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THE INTERMONTANE - OMINECA BELT BOUNDARY
IN THE QUESNEL LAKE AREA, EAST-CENTRAL BRITISH COLUMBIA:
TECTONIC IMPLICATIONS BASED ON
GEOLOGY, STRUCTURE AND PALEOMAGNETISM

by

CHRISTOPHER JOHN REES, B.Sc., M.Sc.

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

Carleton University

Ottawa, Ontario

April, 1987

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in partial fulfilment of the requirements.
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ABSTRACT

The thesis presents a tectonic interpretation of a segment of an important boundary in the Canadian Cordillera, that between the Omineca Belt (OB), representing the (paleo-)continental margin of North America, and the Intermontane superterrane (IM), a composite of tectono-stratigraphic assemblages of oceanic affinity. In the thesis area IM consists of back-arc basin oceanic crust (upper Paleozoic Antler Formation) overlain by island arc rocks (Triassic-Jurassic Takla Group); OB comprises Proterozoic-Paleozoic Snowshoe Group metasediments and Devonian-Mississippian Quesnel Lake (ortho-)Gneiss. This work has established that the boundary is tectonic, marked by a mylonitic shear zone (Quesnel Lake Shear Zone) with upper plate to the east or northeast kinematics, and represents the obduction of a marginal basin and arc (components of IM) from the west or southwest onto OB. This deformation (D_1) spanned the Early-Middle Jurassic boundary and was followed by Middle Jurassic southwesterly-verging folding (D_2) of the coupled IM and OB. This deformation dominates the structure and map pattern and induced the main phase of regional metamorphism, although D_2 strain and metamorphic grade decrease markedly southwestwards into IM across the folded boundary. In effecting D_2 , the OB basement is interpreted to have shortened by horizontal telescoping rather than by major underthrusting; its overlying sedimentary cover was simultaneously folded, thickened and uplifted such that the convergence of IM was deflected downwards, accounting for the vergence of the folds. The tectonic regime is qualitatively analysed in terms of a horizontal quasi-constrictional shortening with a strike-parallel extension, marked by the finite stretching lineation, and a simultaneous transverse or

oblique simple shear component due to the 'undertucking' basement.

D₂ metamorphic grade in the thesis area varies between lower amphibolite facies and prehnite-pumpellyite facies, being mainly greenschist facies. Syn- versus post-kinematic textures indicate diachronous metamorphic quenching due to more rapid cooling in the south versus the north, respectively; this is supported by regional isograd patterns. The transition between D₂ and the kinematically similar late metamorphic D₃ folding and crenulation emulated this migration.

A secondary remanent magnetization is identified in Takla Group volcanoclastics, interpreted to be a mid-Cretaceous overprint; its paleopole is discordant with respect to North America. The paleolatitude is $38 \pm 10^\circ\text{N}$ and the paleodeclination is $73 \pm 15^\circ$ clockwise. Combined with other documented paleopoles, the result provisionally means that most of the southern Canadian Cordillera, including part of OB, has been quasi-rigid and has been rotated and translated northwards 2000 to 2500 km with respect to the craton since the mid-Cretaceous. A major problem emerges in that relative movement of this magnitude is refuted by stratigraphic links between OB and the cratonic miogeocline. It is proposed that presently the paleomagnetism argument is untenable, pending better constraints; the interpretation of the paleopoles is subject to error, the most significant being the non-recognition of crustal tilting or clockwise rotation.

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CHAPTER ONE

INTRODUCTION

[1] 1. GENERAL GEOLOGY OF THE CANADIAN CORDILLERA

INTRODUCTION

The Canadian Cordillera orogenic belt has had a long and complex history, but its present form is due mainly to the emplacement and re-arrangement of allochthonous exotic or suspect (possibly exotic) geological terranes against the evolving North American continental margin in the Jurassic, Cretaceous and early Tertiary. This hypothesis developed in the 1970's and early 1980's out of preliminary plate tectonic models (e.g. Monger et al. 1972, Dercourt 1972) when it was realized that the western half of the orogen in Canada and the United States is a collage of disparate rock assemblages which represent vestiges of ancient Pacific oceanic crust, the rest of which has been consumed by subduction (Hamilton 1969). Detailed stratigraphic analysis of these constituent areas revealed lateral discontinuities which could not be explained simply by unconformities or facies variations. At the same time, paleomagnetic results were accumulating which indicated that

some rocks formed at a latitude much farther south than their present position relative to the North American craton, and at a latitude different from neighbouring rocks. Paleontological work on the climatic affinities of faunas (e.g. Tozer 1970, 1982) supported the notion that some originated in lower latitudes than contemporaneous faunas presently at the same latitude in rocks of the North American miogeocline.

This multidisciplinary evidence was used to delineate "tectonostratigraphic terranes" (see Berg et al. 1972, Davis et al. 1978, Coney et al. 1980, Monger et al. 1982). Their diverse properties indicated that they were not always in their present relative positions, but rather that they were formed apart from one another and commonly apart from the continental margin. Some terranes may be merely telescoped or re-arranged parts of the paleo-continental margin, but they are still considered suspect until their link is 'proved'. It follows that allochthonous or suspect terranes are fundamentally fault-bounded, and that tectonic processes have brought them into their present configuration - primarily the oblique subduction of thousands of square kilometres of proto-Pacific oceanic crust and the accretion of its superstructure to the paleo-continental margin.

Recent geological research in the Cordillera has been aimed at elucidating in detail these tectonic processes by studying the composition of terranes and, in particular, rocks and structures at terrane boundaries. For example, some terranes or parts thereof are often ascribed a plate tectonic affinity, such as island arc, subduction complex, microcontinent, oceanic plateau, or ocean floor (ophiolite). A terrane may form part of two different tectonic settings at two different times; it is important to differentiate the hallmarks of each.

A terrane boundary may be interpreted as a relict subduction zone (suture), obduction zone, a thrust fault, or a strike-slip fault, or a combination of these. The present boundary of a terrane may or may not reflect its original mode of emplacement. Paleomagnetic research is still in the process of defining the itineraries of exotic terranes and that of the craton. It is fundamentally important to find out about the source of terranes, their formation, how, when and where they amalgamated to each other and to the North American continent, and what happened to them afterwards.

This thesis is concerned with the geological composition, structural history, and the paleomagnetic signature of a segment of one such terrane boundary.

GEOLOGICAL BELTS, TERRANES AND SUPERTERRANES

The Canadian Cordillera has conventionally been divided into five longitudinal belts based qualitatively on dominant rock types, structures, metamorphic grade and physiography (Fig. 1.1). More recently, the tectonic subdivision of the Cordillera has incorporated the terrane hypothesis, which is more interpretive (Fig. 1.1). Some terranes transect the boundaries between the five major belts, and their names and limits are frequently revised, reflecting ongoing discoveries and re-interpretations.

In the Jurassic and Cretaceous, individual terranes became attached to each other and (or) to the continental margin (Davis et al. 1978, Monger & Price 1979, Coney et al. 1980, Monger et al. 1982, Monger 1984, Price et al. 1985). The timing of amalgamation is based on the ages of

Fig. 1.1

Tectonostratigraphic terrane map of the Canadian Cordillera, simplified from Price et al. (1985). Legend for terranes in text. Heavy lines outline superterrane: IM in east, IS in west. Uncoloured area in east occupied by Omineca and Foreland Belts (undivided); uncoloured area in west occupied by intrusions of Coast Plutonic Belt. Key to abbreviations: AX = Alexander terrane, BR = Bridge River terrane, CC = Cache Creek terrane, CD = Cadwallader terrane, CN = Chilliwack-Nooksack terrane, MT = Methow terrane, QN = Quesnel terrane, ST = Stikine terrane, WR = Wrangell terrane, YT = Yukon-Tanana terrane. SA = Sylvester allochthon of Slide Mountain terrane, TT = Tintina Trench, NRMT = Northern Rocky Mountain Trench.

Inset shows geologic-physiographic subdivisions of the Cordillera (from Price et al. 1985, Wheeler 1985). V = Vancouver, C = Calgary.

overlapping deposits, reworking and mixing of diagnostic rocks between terranes, and common intrusive suites, especially those that transect terrane boundaries without offset. Similar paleopoles between terranes may be a sign of their continuity. Composites of amalgamated terranes are called superterranes.

The terrane map in Fig. 1.1 is based on the above mentioned work, mainly Price et al. (1985). The bases for the definition of the belts and terranes are described in these works (see also Wheeler 1985) and will not be repeated here, except for the main points relevant to this thesis, which follow in the form of a legend for Fig. 1.1.

INTERIOR PLATFORM (east of the five major belts)

Late Proterozoic to Cretaceous marine sedimentary rocks deposited on North American (cratonic) shelf, by definition unaffected by Cordilleran deformation.

1. FORELAND BELT

Mid-Proterozoic to Upper Jurassic miogeoclinal clastic wedge, composed of craton-derived sedimentary rocks deposited on attenuated continental or transitional crust on the west-facing (paleo-)margin of North American continent. Also known as Rocky Mountain Belt or Rocky Mountain Fold and Thrust Belt.

Includes no allochthonous or suspect terranes.

2. OMINECA BELT

Composed mainly of metasedimentary and igneous rocks derived from mid-Proterozoic to mid-Paleozoic miogeoclinal strata deposited on North

American attenuated continental or transitional crust. Metasediments partly correlatable with Foreland Belt strata. Also includes terranes of Proterozoic to lower Mesozoic metasediments, metavolcanics and intrusions of suspect affinity to North America (Kootenay terrane, Monashee complex, Barkerville terrane, Yukon-Tanana terrane).

Locally overlaps the ill-defined eastern limit of allochthonous terranes (which mainly constitute the INTERMONTANE BELT).

3. INTERMONTANE BELT

Composed of a variety of upper Paleozoic to mid-Mesozoic volcanic, plutonic and sedimentary rocks, which have no affinity to North American (paleo-)continental margin rocks. Subdivided into several terranes by rock-types, stratigraphy, fossils, and internal structure. Diverse paleomagnetic signatures among terranes.

Constituent terranes listed below in general east-west order of occurrence.

Slide Mountain terrane (SM). (Mid-)upper Paleozoic (meta-)basalt, gabbro, ultramafic rocks, chert, volcaniclastics and argillite.

Quesnel terrane (QN). Upper Paleozoic (meta-)volcanic, sedimentary rocks, and orthogneisses, overlain via unconformity by Upper Triassic to Lower (Middle) Jurassic calc-alkaline and alkaline silicic to basic volcanic and comagmatic plutonic rocks, and sedimentary rocks. Stratigraphically linked to SM and CC by the end of the Triassic.

Stikine terrane (ST). Possible upper Precambrian basement. Upper Paleozoic volcanic and sedimentary rocks. Overlain via unconformity by Upper Triassic to Middle Jurassic volcanic and plutonic rocks. Probably

7
linked stratigraphically to CC by the end of the Triassic.

Cache Creek terrane (CC). Tectonic melange of Mississippian to Upper Triassic radiolarian chert, argillite, carbonate, basalt, serpentinite, and rare blueschist. Includes exotic Permian Tethyan faunas.

INTERMONTANE SUPERTERRANE (IM)

Composite of SM, QN, ST and CC, which were amalgamated by the end of the Triassic; signified by stratigraphic linkages or common intrusive suite.

IM analogous to Intermontane Belt.

4. COAST PLUTONIC BELT

Mainly Cretaceous and Tertiary granitic batholiths, with septa of metamorphosed ?Precambrian to Tertiary sediments and volcanics. Igneous rocks intrude ST (of IM) in the north(east) and intrude WR (of IS) in south(west).

5. INSULAR BELT

Composed of two main terranes:

Wrangell terrane (WR). Upper Paleozoic, Triassic, and Triassic-Jurassic sequences of volcanic and sedimentary rocks. Includes low (equatorial) latitude Triassic coralline faunas.

Alexander terrane (AX). Composite terrane of ?Precambrian, Paleozoic and lower Mesozoic volcanic and sedimentary rocks.

INSULAR SUPERTERRANE (IS)

Composite of WR and AX, amalgamated by Upper Jurassic (overlap of Gravina-Nutzotin Belt).

Other terranes in Fig. 1.1

Methow terrane (MT). Triassic basalt(?), Lower Jurassic to Cretaceous sedimentary and volcanic rocks.

Bridge River terrane (BR). Permian to Middle Jurassic melange of chert, argillite, basalt, ultramafics, and carbonate.

Cadwallader terrane (CD). Triassic basalt, carbonate and clastics, overlain by Upper Triassic to mid-Cretaceous clastics.

Chilliwack-Nooksack terrane (CN). Upper Paleozoic sedimentary rocks and dacitic to basaltic volcanics, overlain by Upper Triassic and Lower Jurassic sedimentary and volcanic rocks, and by Jurassic volcanics and Jurassic to mid-Cretaceous clastics.

MT, BR, CD and CN occur roughly between IM and IS, and are affiliated with IM.

[1] 2. GENERAL GEOLOGY, AND PREVIOUS WORK IN THE QUESNEL LAKE AREA

GENERAL GEOLOGY

Quesnel Lake is a large lake in east-central British Columbia (121°W, 52.5°N; Fig. 1.2). The Columbia Highlands, on the edge of the Cariboo Mountains, lie to the northeast of the lake, and the Interior

Fig. 1.2

General geology of the Quesnel Lake region showing location of the thesis area, and of some relevant previous work. Base map from Tipper et al. (1979), and incorporates only minor revisions from more recent work (excluding this thesis), such as major thrusts.

Plateau of British Columbia is to the southwest. The boundary between the Intermontane Belt (or superterrane) and the Omineca Belt passes through the Quesnel Lake area from northwest to southeast.

On the northeast side of the boundary, the Omineca Belt consists of Upper Proterozoic and Paleozoic psammitic and pelitic metasediments of the Snowshoe Group, and Paleozoic granitic intrusive rocks (Quesnel Lake Gneiss; Fig. 1.2). In contrast, the Intermontane Belt consists of upper Paleozoic meta-basaltic and meta-ultramafic rocks (Antler Formation), overlain by lower Mesozoic phyllitic pelites (Black Phyllite) followed by basaltic volcanic, volcanoclastic and sedimentary rocks and cogenetic intrusive rocks collectively called the Takla Group (Fig. 1.2).

There is a marked strain and metamorphic gradient across the boundary. Omineca Belt rocks are highly strained by several phases of deformation, under low to high grade metamorphic conditions. These effects diminish southwestwards towards the Intermontane Belt but transgress the boundary, across which metamorphic grade is low or very low, and deformation is characterized by mainly open to moderate folds and numerous steep faults.

PREVIOUS WORK

The Quesnel Lake area was first mapped on a reconnaissance basis by A. Bowman in the late 1800's, and in the 1960's by R.B. Campbell, both of the Geological Survey of Canada (G.S.C.). Campbell published maps with marginal notes (Campbell 1961, 1963; also Campbell & Campbell 1970) and a revised map (Campbell 1978) covering the Quesnel Lake sheet (NTS 93A).

Notable geological mapping on the stratigraphy in the Cariboo

Mountains has been done by Johnston & Uglow (1926), Holland (1954), and Sutherland Brown (1957, 1963). The most recent contributions are by L.C. Struik of the G.S.C. who has been tracing the Upper Proterozoic through Paleozoic stratigraphy along the western margin of the Qmineca Belt from latitude 54°N southeastwards through the Quesnel Lake area and beyond (Fig. 1.2; Struik 1980 to 1986). Struik has revised the stratigraphy considerably, and has recognized several thrusts and strike-slip faults in the region. He proved the allochthoneity of the Antler Formation in the Black Stuart synclinorium (Fig. 1.2). His resulting tectonic synthesis involved the emplacement of sub-terrane and thrust sheets from the east and west on low-angle faults (Struik 1986b).

Several unpublished theses have been done in adjacent parts of the thesis area (Fig. 1.2). Morton (1976) mapped the Horsefly area south of Quesnel Lake, concentrating on the cyclical stratigraphy and petrography of Mesozoic volcanic rocks, which he designated "Horsefly Group" (= Takla Group in this thesis). He concluded that the volcanics were generated by crustal rifting, because of their alkalic compositions. Bailey (1978) studied the Morehead Lake area, immediately to the west of Morton, with similar emphasis on the volcanic and plutonic rocks and mineral deposits, and their origin. His interpretation differed from that of Morton, proposing that these rocks were formed in an island arc above a convergent plate margin.

The Snowshoe Group immediately east and northeast of the thesis area was studied by Getsinger (1985), who interpreted these rocks as continental shelf to slope deposits. She identified four phases of folding, notably northeasterly-verging first phase folds, and southwesterly-verging second phase folds which were accompanied and

outlasted by lower and middle amphibolite-facies metamorphism. A southeasterly-directed, low-angle, post-metamorphic normal fault in the north of her area, the Little River fault, emplaced lower grade phyllites on top of Snowshoe Group schists and gneisses.

The Intermontane-Omineca Belt boundary continues southeast of Quesnel Lake into the Crooked Lake area (Fig. 1.2) and beyond, where it has been examined by K.V. Campbell (1971) and Montgomery (1978). In the Crooked Lake area, K.V. Campbell (1971) studied the same units as this thesis, namely Omineca Belt Upper Proterozoic metasediments (his "Kaza Group"), and in the Intermontane Belt, upper Paleozoic meta-basaltic rocks (his "Slide Mountain Group") and Upper Triassic and Lower Jurassic metasediments and volcanics. Campbell found that the contact between the "Slide Mountain Group" and the "Kaza Group" (the Intermontane-Omineca boundary) was tectonized during eastward thrusting of unknown magnitude of the former over the latter, with the production of "mylonites" and "cataclastic rocks", prior to regional folding of the contact and metamorphism. However, he argued, partly for reasons now believed to be spurious, that the "Slide Mountain Group" was originally in stratigraphic juxtaposition with the "Kaza Group" and that the contact was later modified by the thrusting, prior to Mesozoic deformation. He also maintained that the several phases of major and minor folding and metamorphic recrystallization were the result of a single, continuous or progressive folding event, rather than of superimposed deformations. Folding or fabric formation related to the earlier thrusting was not mentioned.

Montgomery (1978) worked just south of K.V. Campbell, concentrating on an isolated, 3 km thick composite mafic-ultramafic assemblage which

he called the "Black Riders Complex": This consists of ultramafic rocks at the base, including serpentized and steatized dunite-harzburgite with cumulate layering, followed by meta-gabbroic and amphibolite gneiss, and this overlain via a thin zone of dykes by amphibolitic schists. This complex, which Montgomery correlated with the "Slide Mountain Group", is separated from the underlying Omineca Belt metasediments by a "west-dipping thrust of unknown displacement". Montgomery concluded that the Black Riders Complex is a klippe composed of an ophiolite complex, obducted eastwards from the floor of a marginal basin that separated the Omineca Belt continental margin from the "Nicola-Takla volcanic arc" to the west, after the Upper Triassic and before Middle Jurassic deformation.

Much of K.V. Campbell's (1971) thesis area has been re-studied in recent years by four Master's theses from the University of British Columbia (Fig. 1.2). Fillipone (1985), Montgomery (1985), and Elsbey (1985) all recognized a phase of deformation in the Snowshoe Group which pre-dated the emplacement of the Intermontane Belt, which they interpreted as an allochthonous terrane. Carye (1986) differed somewhat in proposing an extra phase of deformation in the Antler Formation as well as in the Snowshoe Group, but he could not correlate them conclusively.

There are some differences in interpretation between these four workers, but in general (see also Ross et al. 1985 for a synthesis) they recognized F_1 easterly-verging isoclinal folds (confined to the Snowshoe Group), and F_2 easterly-verging folds related to the emplacement of the Intermontane Belt, with mylonitization at the terrane contact. This F_2 event was synchronous with the main metamorphism, which peaked slightly afterwards. Montgomery (1985) dated this metamorphism as Middle

Jurassic from titanite U-Pb geochronometry. F_3 folding was upright to southwesterly-verging and responsible for the regional map pattern.

F. Ferri was the author's assistant in 1981 and completed a B.Sc. Honours thesis (Ferri 1982) on the structure of the Western Quesnel Lake Gneiss within the thesis area.

[1] 3. STATEMENT OF THE PROBLEM

As outlined in Section 1, tectonic reconstructions in the Cordillera depend on the knowledge of the composition of terranes and of the nature of terrane boundaries. This thesis focuses on the Intermontane-Omineca Belt boundary in the Quesnel Lake area, concentrating on the following issues.

- * Is the boundary tectonic, or merely a tectonized unconformity, as proposed by K.V. Campbell (1971)?

- * The nature of the Intermontane Belt rocks. Is Montgomery's (1978) interpretation that the lowermost rocks represent an ophiolite complex tenable also in the thesis area? Are the overlying Mesozoic volcanic rocks rift- or subduction-related? (Morton 1976 versus Bailey 1978). What was the environment of their formation (which would aid in the reconstruction of the pre-deformation tectonic setting)?

- * What was the environment of formation of the Omineca Belt rocks? Are they consistent with being part of the (paleo-)continental margin sedimentary prism?

* If the boundary is tectonic, how and when did emplacement occur? What was the sense of movement and the displacement vector? What metamorphic conditions prevailed? What was the magnitude and rate of displacement?

* Is there any phase (or phases) of deformation in the Omineca Belt that is absent in the Intermontane Belt or vice versa, i.e. one that pre-dates their juxtaposition?

* What deformation(s), if any, followed emplacement of the Intermontane Belt? Can they be correlated with deformations in adjacent areas (see Section 2) and in areas farther into the deformed continental margin? When in the Mesozoic or Cenozoic did they occur? What regional tectonic processes do they reflect?

* What are the timing and spatial relationships between the metamorphism and deformation(s)? What were the prevailing metamorphic pressure and temperature conditions? What is the significance of the metamorphic and strain gradient across the boundary?

* How does the tectonic history compare with other parts of the Intermontane-Omineca boundary in the Canadian Cordillera? Is there an analogous tectonic record in the conterminous United States?

* What is the paleomagnetic signature of the region? What, depending on age of magnetization(s), does it imply about the large scale tectonic history of the region and of the Cordillera?

[1] 4. LOCATION OF THE THESIS AREA, AND FIELD
AND LABORATORY METHODS

LOCATION

The Quesnel Lake area is suitable for a study of the Intermontane-Omineca boundary, as major structures are well-displayed in the map pattern (Fig. 1.2) and because it is not inundated with intrusions as it is in the Kootenay Arc in southern British Columbia (Fig. 2.3). Exposure is quite good southeast of the lake in the Crooked Lake area which has been studied intensively in recent years (see Section 2), but diminishes northwest of the lake which has never before been studied in detail. Thus, the area straddling the boundary between the village of Likely and the North Arm of Quesnel Lake was chosen for the present study (Fig. 1.2) on the advice of R.B. Campbell. The eastern limit is roughly Grain Creek, which is roughly the southwestern limit of Getsinger's (1985) Ph.D. area.

Likely is reached via a partly paved road from Williams Lake (Fig. 1.2)

FIELD WORK

The thesis area is mostly below tree-line except for a series of alpine or subalpine ridges about 2000 m (ca. 6500') high in the north and east of the area. Relief in the area is moderate (Quesnel Lake has an elevation of 725 m or 2400'), and the terrain is easily traversed. Snow covers the best exposures above about 1700 m (ca. 5500') until June or July.

The area is quite heavily logged so access by vehicle on the

numerous major and minor logging roads is very good, although many old roads are in disrepair and are impassable other than on foot or perhaps by 4-wheel drive. Old clearcuts are dense with second growth. Unlogged bush, also, is locally thick with undergrowth and littered with deadfall, which is greater than normal because the trees are diseased.

Field work was conducted between June and August inclusive in 1980, 1981 and 1983. Mapping was done on 1:50 000 topographic maps except in the area around Browntop Mountain, Goose Peak and Badger Peak which was mapped at 1:25 000 scale. The thesis area is covered by NTS sheets 93A-11 and 12. Provincial air photos cover the area below tree-line at 1:15 840 scale, and the whole area at roughly 1:66 000 scale; federal air photos cover the area at 1:63 360 scale. The provincial small-scale photos are particularly useful for navigating in the mazes of clearcuts. Forestry maps, kindly provided by the Likely Forestry service, were also very useful.

The cleanest and highest incidence of exposures, especially of soft lithologies, are on logging roads and in clearcuts; exposure in the bush below 1700 m (ca. 5500') is generally not very good. Traverses were often undertaken along streams or spurs where exposure tends to be better. Exposure is limited by the cover of Quaternary deposits, especially in the Cariboo River valley, and in the southeast and northwest of the thesis area.

Fly camps were undertaken in alpine areas around Badger Peak and Mt. Brew for 2-3 or 5-6 weeks at a time, in July or August. Access was provided by Okanagan Helicopters and Highland Helicopters, both based in Williams Lake. [Air time from Williams Lake to Likely is ca. 30 minutes, unloaded.] Some 4-5 day trips into the high ground were done

by back-packing.

LABORATORY WORK

Geochemical analysis of Antler Formation rocks was done partly by the Analytical Chemistry Laboratory at the Geological Survey of Canada (Ottawa) by the XRF method, and partly by Bondar-Clegg and Co. Ltd. (Ottawa) by XRF and DC plasma emission.

Rubidium-Strontium geochronological work on whole-rock and mineral samples from the Quesnel Lake Gneiss was done by R.L. Armstrong at the University of British Columbia (Vancouver).

Analyses of garnets were done by Peter Jones on a Cambridge, Mark V electron microprobe at Carleton University.

Paleomagnetic analysis was done by the author at the Pacific Geoscience Centre, Sidney, British Columbia (see Chapter 5 for details).

Rock mineralogies, textures and fabrics were studied with a standard thin-section petrographic microscope, and universal stage.

CHAPTER TWO

STRATIGRAPHY

There are five major and two minor rock units in the thesis area (Figs. 2.1, 2.2, Table 2-1). These are described in this chapter in stratigraphic order.

The Antler Formation, the Black Phyllite and the Takla Group are described in some detail because this is pertinent to the tectonic reconstructions and modelling in Chapters 6.1 and 6.2. The Snowshoe Group is described briefly here, and in more detail in Appendix A.

[2] 1. SNOWSHOE GROUP (UNIT HPS)

The metasedimentary rocks in the Omineca Belt in the thesis area are designated Snowshoe Group (Fig. 1.2), after Struik (1983a).

AGE AND CORRELATION OF THE SNOWSHOE GROUP

Struik (1983a) defined the Snowshoe Group as all those rocks of the Omineca Belt which occur in the lower plate of a major thrust fault, the

LEGEND FOR FIG. 2.1

QUATERNARY VOLCANIC ROCKS (QV)

olivine basalt, debris flows

LOWER-MIDDLE JURASSIC ROCKS (LMJ)

volcaniclastic wackes & breccia, chert-pebble conglomerate, limestone, shale, volcanics

UPPER TRIASSIC-LOWER JURASSIC TAKLA GROUP (TrJT)

b = augite porphyry basalt flow & flow breccia

c = volcanic conglomerate (debris flow)

g = volcanic wacke; s = siltstone

f = volcanic wacke

a = tuffaceous phyllite, argillite & siltstone;
v = pillowed mafic volcanics

UPPER TRIASSIC BLACK PHYLLITE (uTrBP)

p = graphitic pelite

c = volcanic (pyroclastic?) conglomerate,
f = volcaniclastic schist

PENNSYLVANIAN-PERMIAN ANTLER FORMATION (PPA)

greenstone, metagabbro, serpentinite, steatite

DEVONIAN-MISSISSIPPIAN QUESNEL LAKE GNEISS (DMWQ & DMEQ)

Western Quesnel Lake Gneiss: granitic orthogneiss

Eastern Quesnel Lake Gneiss: granodioritic orthogneiss

PROTEROZOIC-PALEOZOIC SNOWSHOE GROUP (HPS)

Pelitic & psammitic metasediments, minor pebbly conglomerate, carbonate; dense dots = subunit HPSH

Fig. 2.1 General geology of the thesis area. Axial traces of major F2 folds also shown.

Fig. 2.2

Composite, generalized cross-section through the thesis area. See Fig. 2.1 for legend. Main F_2 axial surfaces shown. Contact between PPA and HPS or DMWQ is Quesnel Lake Shear Zone.

TABLE 2-1: TABLE OF FORMATIONS

Era	Period or epoch	Lithologic or stratigraphic unit	Lithology
Cenozoic	Quaternary	Quaternary Volcanic Rocks	olivine basalt & volcanoclastics
	Unconformity		
Mesozoic	Lower and Middle Jurassic	Unnamed Jurassic Rocks	volcanic wacke, breccia, conglomerate, mudstone, limestone
	Unconformity?		
	Upper Triassic (Karnian to Norian)	Takla Group	tuffaceous argillites & siltstone, volcanoclastic wacke, volcanic conglomerate, basalt flow
		Black Phyllite	graphitic phyllite & siltstone; minor volcanic conglomerate
Paleozoic	Disconformity?		
	lower Pennsylvanian to Lower Permian	Antler Formation	greenstone, greenschist, metagabbro, serpentinite & -breccia, steatite
	Tectonic contact		
	Late Devonian to early Mississippian	Quesnel Lake Gneiss	granitoid orthogneiss (strongly sheared at structural top)
Proterozoic and (or) Paleozoic	Intrusive contact		
	Devono-Mississippian? Cambro-Ordovician? Hadrynian?	Snowshoe Group (base not observed)	gritty siliciclastic & pelitic metasediments, rare carbonate & misc. (strongly sheared at structural top)

Pleasant Valley thrust (Fig. 1.2; Struik 1982a), to as far west as the boundary with the Intermontane Belt or superterrane. Struik (1983a) informally divided the Snowshoe Group into several successions which may be time-correlative with presumed craton-derived Hadrynian through Paleozoic strata in the upper plate of the Pleasant Valley thrust, namely the Hadrynian Kaza Group, the Hadrynian-Cambrian Cariboo Group, and the Ordovician to Mississippian Black Stuart Group, collectively called Cariboo terrane by Struik (1985a). The Snowshoe Group successions that Struik (1983a) recognized in the thesis area are the Hadrynian Ramos Creek and Keithley successions, and the ?Paleozoic Harvey Creek and Goose Peak successions.

The "Snowshoe Group" is a complete revision of the original "Snowshoe Formation" defined by Holland (1954) as the highest unit of the Cariboo Group. This new designation for the Snowshoe was proposed by Campbell (1978) and Struik (1983a) because it has a different stratigraphy to the groups in the upper plate of the Pleasant Valley thrust and cannot be affiliated conclusively with any of them, although Struik (1986a) did suggest a specific correlation. The Snowshoe Group lacks reliable fossil control, preventing confident correlation across the Pleasant Valley thrust with the Cariboo terrane, although it contains a fossiliferous limestone which demonstrates that at least part of it is Paleozoic (Struik 1983a, 1986a).

The Snowshoe Group may be in part correlative with the Paleozoic Eagle Bay Formation of the area northeast of Kamloops, southern British Columbia (Fig. 2.3; Struik 1982b). Parts of the Snowshoe Group may also be correlative with the lower Paleozoic Lardeau Group and the Mississippian-Pennsylvanian Milford Group in southern British Columbia

Fig. 2.3

General geology of the southeastern Canadian Cordillera, based on Tipper et al. (1981), Price et al. (1985), Okulitch (1985), Brown et al. (1986), Brown & Journeay (1987). Pre-Tertiary geology emphasized: Tertiary intrusions and extensive volcanic cover are omitted for clarity.

(Fig. 2.3; Table 2-2).

A note on terrane terminology

The aforementioned Proterozoic and Paleozoic rocks of the western Omineca Belt (Snowshoe, Eagle Bay, Lardeau, Milford) comprise a variety of lithologies, some of which resemble presumed craton-derived clastic metasediments, such as blue quartz grits, psammites and pelites, typical of the Hadrynian Windermere Supergroup (including the Kaza Group). Other lithologies, however, are quite different including mafic and felsic volcanics, chert-carbonate sequences, chert/volcanic/granite/diorite-pebble conglomerates, and 'black shale facies' sediments (Struik 1981b, Klepacki 1983, Klepacki & Wheeler 1985, Preto 1981). They are also distinct from most of the Omineca Belt to the east in containing Devonian-Mississippian orthogneisses and granitoid intrusions (e.g. Quesnel Lake Gneiss, this thesis), and locally a record of lower to mid-Paleozoic metamorphism and deformation, and uplift and erosion not apparent elsewhere (Klepacki 1984, Price et al. 1985).

These western Omineca Belt assemblages are generally considered to be a distal western facies of the (paleo-)continental margin miogeocline, although a substantially different, western source terrane has also been postulated (e.g. Campbell et al. 1973, Struik 1982a,b, Price et al. 1985). The recognition of their unusual characteristics has rendered these rocks "suspect" in terms of their primary relationship to North America; Monger (1984) and Price et al. (1985) distinguished them as the Kootenay terrane (Table 2-2). Struik (1985a, 1986b) further separated the Snowshoe Group as the Barkerville terrane.

A very similar assemblage of rocks is present in southeastern Yukon,

TABLE 2-2: TERRANE CONSTITUENTS, TERMINOLOGY AND PROPOSED CORRELATIONS
RELEVANT TO THIS THESIS

QUESNEL TERRANE	SLIDE MOUNTAIN TERRANE ²	SOUTHERN OMINECA BELT
(Cretaceous & Cenozoic rocks)	Antler Fm. (in Black Stuart synclinalorium)	Snowshoe Gp. ^{3 or 4}
Nicola Gp. ca. = Takla Gp. } Quesnel Belt (geographic division)	Fennell Fm.	Eagle Bay Fm. ⁴
- see text.	Kaslo Gp.	Milford Gp. ⁴
Black Phyllite = Slocan Gp.	(Sylvester Allochthon, Anvil Gp., in northern B.C. & Yukon)	Lardeau Gp. ⁴
Antler Fm. (in Quesnel Lake area) ¹ (= Fennell Fm., Kaslo Gp.)		
Harper Ranch Gp. Mt. Lytton Complex		
Possible Precambrian basement		
¹ Antler Fm. in Quesnel-Lake area is included as part of Quesnel terrane. because they are in probable stratigraphic contact (see text)	² Formerly called Eastern Assemblage and Eastern terrane (see text)	³ Included in Barkerville suspect terrane ⁴ Included in Kootenay suspect terrane

the Yukon-Tanana terrane (Fig. 1.1), characterized by Paleozoic siliciclastic metasediments, metavolcanics and Devonian intrusions.

Mortensen & Jilson (1985) have noted its similarity with the Barkerville and Kootenay terranes, and have even suggested that they are displaced equivalents.

Throughout this thesis, the terms Snowshoe Group or Omineca Belt will be used to describe these rocks in the Quesnel Lake region, rather than "Barkerville terrane".

LITHOLOGY OF THE SNOWSHOE GROUP

It was not possible to subdivide the Snowshoe Group (unit HPS) rocks in the thesis area into stratigraphic units because of the structural complexity and the paucity of marker units. Struik's (1983a) map units could not be followed or supported by the author, with the exception of part of the Harvey Creek succession, which is here called subunit HPS_h. Instead, the rocks have been classified according to lithology; lithological subunits form the basis of the map, and the following brief descriptions. More detailed descriptions are in Appendix A.

Most of the Snowshoe Group in the thesis area is metamorphosed to the greenschist facies. (The prefix meta- is usually omitted in the rock names.)

Subunit HPS_g. This subunit is characterized by a predominance of competent 'gritty' siliciclastic lithologies over pelites, and includes quartz psammites, (feldspathic-) quartzites, and quartzofeldspathic pebbly conglomerates (Plate 1B). They are usually fine to medium grained, well bedded, and well foliated. Blue quartz is fairly common. Thin

interbeds of grey, fine grained, phyllitic pelite (locally calcareous) are subordinate.

Rocks of this subunit at or near the contact with the Antler Formation are very fine grained, and finely foliated due to strong deformation and dynamic recrystallization.

Subunit HPSq. This is a separate subunit consisting of significant beds of quartzite. Quartzites are pale coloured, fine to medium grained, and some are well foliated and lineated.

Subunit HPSp. This subunit is the most widespread and typical in the Snowshoe Group. It comprises well- and thinly bedded, fine grained, phyllitic pelite, and psammite (Plate 1A,C). In the highest grade area east of Mt. Brew (see below), these rocks are coarse grained (garnet-) mica schists.

The interbedded psammities are more thinly bedded and finer grained than those that characterize subunit HPSg.

An uncommon variety of this subunit is calcareous pelitic phyllite (HPSc), consisting mainly of carbonate, with foliae of mica. It may have been a marl originally.

Subunit HPSk. This subunit is compositionally similar to HPSp and HPSg, with which it is usually associated. It is distinguished only by its distinctive olive to khaki colour, and by the common presence of a white or pink-white quartzite. (This might be useful to future workers in the area.)

Subunit HPSm. This subunit comprises rare, thin bands (>3 m thick) of thinly bedded marble within subunits HPSg and HPSp. The marbles are creamy-white, and consist mainly of medium- to coarse grained calcite.

Subunit HPSh. This subunit consists of black psammites and pelites. It was mapped by Campbell (1978) as the ?Midas Formation (but is no relation to the type-Midas Formation of the Cariboo Group), and as the Harvey Creek succession by Struik (1983a).

Subunit HPSh is dark blue-grey to black, fine to medium grained graphitic psammite and phyllitic pelite. Its bedding and foliation are particularly irregular, disrupted by the deformation. Locally the subunit contains fine grained, dark charcoal grey impure limestone or calcareous siltstone, which is also graphitic.

Subunit HPSl. This pale to dark charcoal grey, fine grained impure limestone was found only in Locality 829. It is slightly argillaceous and graphitic, and is interbedded with dark pelites. It is virtually identical to the limestone within subunit HPSh, but lacks the associated black psammites.

Subunit HPSd. This is a very sporadic rock type consisting usually of pale, creamy-grey, fine to medium grained 'silicate-carbonate'. The predominant carbonate has replaced ca. 60% of the original rock. It is usually non-foliated, and is characterized by much limonite, pyrite, and locally fuchsitic mica. This rock is most probably altered and carbonatized diabase (or basalt), based on the author's previous experience of this lithology.

Subunit HPSi. This subunit includes a variety of apparently late intrusive rocks (excluding HPSd), such as massive, dark green-grey diabase-gabbro, ?lamprophyre, and meta- ?diabase. There is one occurrence (locality 866) of white pegmatite (labelled HPSip on Map 1).

Subunit HPSa. This is a dark green, medium grained garnet amphibolite, which was found in situ only at locality 151. The amphibole has the habit of hornblende, but the extinction angle of actinolite. This rock is interpreted to be a pre-metamorphic mafic intrusion. It is probably not equivalent to the para-amphibolites mapped by Getsinger (1985) in the area to the north.

The Snowshoe Group south and east of Mt. Brew.

The Snowshoe Group in this part of the thesis area is distinctive in being metamorphosed to the lower amphibolite facies grade, the highest in the thesis area. Consequently, some aspects of the preceding descriptions such as grain size, do not apply to this area, although these rocks are the same as elsewhere in terms of bulk composition and the character of the bedding. They are probably not stratigraphically distinct, therefore.

Foliations are medium- to coarse-grained, especially in the pelites where garnet porphyroblasts are up to 2 cm across, much larger than elsewhere. Texturally, these rocks could be described as paragneisses (Plate 1E), whereas the rest of the Snowshoe consists of metasedimentary schists and phyllites. Locally pelitic rocks are streaked with trondhjemitic segregations and veins which may be concordant or discordant with the compositional layering (Plate 1D). The veins are generally up to 10 to 20 cm thick and locally constitute up to 20% of the volume the rock.

INTERPRETATION OF DEPOSITIONAL ENVIRONMENT OF SNOWSHOE GROUP

The sedimentological character of the Snowshoe Group has been modified by ductile deformation and metamorphic recrystallization to an extent that makes a detailed analysis of the depositional environment impossible. The only primary structure that is preserved is graded bedding. Original bed shape and thickness have been altered by tectonic transposition, but a general appreciation of their original form is still possible.

The main and typical properties of unit HPS that are relevant to facies analysis are as follows.

(a) It is a several kilometre-thick pile of meta-sandstone and meta-mudstone.

(b) These lithologies are moderately to thinly interbedded. Beds are of uniform thickness, parallel-sided, laterally persistent, and not incised by channels.

(c) 'Gritty' and pebbly lithologies, representing medium to coarse grained sandstones, are moderately well sorted and graded. Maximum grain size is 1 cm in rare pebbly conglomerates. In good exposures, the tops of grits grade into pelites. No cross-bedding, cross-lamination or ripple marks were found.

(d) Pelites, representing mudstone units, are predominant. They are finely laminated in low grade rocks. Thickness is variable.

(e) There is no discernible fining- or coarsening-upward trend in the stratigraphy.

(f) Carbonate is uncommon, very thin, and laterally discontinuous.

Marine sand-mud lithofacies are associated mainly with shallow seas, or with turbidites forming submarine fan complexes on continental slopes. The former, shallow marine environments, are characterized by current-induced sedimentary structures, good sorting, and indigenous fossils or bioturbation. These features are absent in the Snowshoe Group; this is not likely to be due entirely to non-preservation, and is interpreted to preclude the possibility of a shallow marine environment.

Submarine fan complexes are composed of turbidites forming a variety of sand-mud lithofacies, which have been summarized by Howell & Normark (1982). Considering earlier models, these authors recognized several facies distinguished by sand-mud ratio, bed thickness, channelling, grain size, and sedimentary structures. The Snowshoe Group conforms with their turbiditic lithofacies C and D, which are characteristic of the outer fan apron of a submarine fan complex. This corresponds to the traditional 'distal turbidites'.

It is concluded that this is the most reasonable depositional model for the Snowshoe Group. This implies a deep, clastic marine environment, probably below the carbonate compensation depth (CCD). The minor occurrence of carbonate as thin marbles (subunit HPSm), impure limestone (subunits HPSH and HPSl), or as calcareous cement in some lithologies can be explained by infrequent influxes of CaCO_3 , mixed in with the turbidity current, probably derived from the erosion of limestone from the shelf at the head of the fan. Also, it could merely represent transient and local biogenic activity or chemical precipitation in situ.

Struik (1985b) proposed that the Barkerville terrane strata, of which the Snowshoe Group forms the main part, were deposited on "a

continental slope subject to periodic volcanism in the Paleozoic, possibly related to rifting". This is probably compatible with the interpretation made here, although in another article Struik (1986b) described the Barkerville terrane as "continental shelf clastics, carbonate and volcanics".

[2] 2. QUESNEL LAKE GNEISS (UNIT DMWQ & DMEQ)

This name has been given to two varieties of granitoid orthogneiss in the thesis area (Campbell 1978). One variety outcrops east and west of Cariboo River, and in a strip adjacent to the Quesnel Lake Shear Zone north of Seller Creek (Fig. 2.1). This orthogneiss is herein called the Western Quesnel Lake Gneiss (Fig. 2.4). The other orthogneiss occurs as the most western part of an elongate body, 50 km by 5 km, which extends from just east of Mt. Brew in the thesis area to south of the east arm of Quesnel Lake (Fig. 2.4). Although it is discontinuous and somewhat variable in composition, it is probably a single magmatic body. This orthogneiss, where it occurs in the thesis area, is called herein the Eastern Quesnel Lake Gneiss.

The Western- and Eastern Quesnel Lake Gneiss are quite distinct isotopically and geochemically, with somewhat different ages, and with very different $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratios (Armstrong, pers. comm. 1982). They also differ in modal composition and texture, described below. However, the geochemical and geochronological evidence is not conclusive enough to confidently differentiate their origins, so it is inferred that they represent different members of a single intrusive suite.

Fig. 2.4

Quesnel Lake Gneiss and probably related gneisses in the Quesnel Lake region: terminology used in this thesis. See Table 2-3.

AGE AND CORRELATION OF THE QUESNEL LAKE GNEISS

Several attempts have been made to date the various bodies of the Quesnel Lake Gneiss, mainly by R.L. Armstrong and associates at the University of British Columbia. The results range from Proterozoic to mid- and upper Paleozoic (Table 2-3). Obviously, the ages are not very consistent, and only general comments and provisional interpretations are given here.

The Rb-Sr whole-rock method has yielded a Late Ordovician age (447 ± 32 Ma) for the Western Quesnel Lake Gneiss, and a Mississippian age (351 ± 70 Ma) for the Eastern Quesnel Lake Gneiss (Fig. 2.5, Table 2-3).

Zircons from the East Arm body of the Quesnel Lake Gneiss and from a separate but probably related granitic orthogneiss (Boss Mountain Gneiss, Fig. 2.4) have recently been analysed by Mortensen et al. (in press; Table 2-3). The data for the Quesnel Lake Gneiss is variably discordant and yields early- to mid-Paleozoic U-Pb and Pb-Pb ages. Zircons from the Boss Mountain Gneiss are also discordant, and indicate a minimum emplacement age of 339 ± 10 Ma (mid-Mississippian). Taken together, the data sets suggest that both bodies were emplaced in the range of 375 to 350 Ma (Mortensen et al., in press). These results are compatible with the Early Mississippian Rb-Sr date for the Eastern Quesnel Lake Gneiss, and cast doubt on the early Paleozoic Rb-Sr date for the Western Quesnel Lake Gneiss. It is tentatively concluded that the Quesnel Lake Gneiss suite is Late Devonian to Early Mississippian.

Porphyritic granitoid orthogneisses of Devonian to Mississippian age are indeed widespread, though not very common, in the Omineca Belt from Yukon to southern British Columbia (Mortensen 1983, 1985 and in press,

TABLE 2-3: GEOCHRONOLOGICAL DATA FROM THE QUESNEL LAKE GNEISS

Locality	Method	Age (Ma)	Initial $^{87}\text{Sr}/^{86}\text{Sr}$	Notes
Western QLG	Rb-Sr, whole-rock	447 ± 32	0.7199 ± 0.0018	Collected by author, analysed by Armstrong
Eastern QLG	Rb-Sr, whole-rock	351 ± 70	0.7073 ± 0.0005	Collected by author, analysed by Armstrong
East Arm QLG	U-Pb, zircon	317 - 340	-	Dave Parkinson (UBC) pefs. comm.
Boss Mt. Gn.	U-Pb, zircon	340 - 350	-	Dave Parkinson (UBC) pers. comm.
East Arm QLG	U-Pb, zircon	350 - 375	-	Estimated age of emplacement (Mortensen et al., in press)
Central segment of QLG	Rb-Sr, whole-rock	752 ± 76	0.7035 ± 0.001	Ref: Blenkinsop 1972
Western QLG	Rb-Sr, whole-rock	746 ± 56	0.7040 ± 0.0008	Combined data sets of author and of C.J.N. Fletcher (UBC)
East Arm QLG	U-Pb, titanite	174 ± 4	-	Age of metamorphism (Mortensen et al., in press)
Western QLG	Rb-Sr, plagioclase	161 ± 5	0.7478	From matrix plagioclase + qz

TABLE 2-3 (concluded):

Locality	Method	Age (Ma)	Initial $^{87}\text{Sr}/^{86}\text{Sr}$	Notes
Eastern QLG	Rb-Sr, biotite	106 ± 3	0.7088	Collected by author, analysed by Armstrong. Mesozoic resetting (Armstrong, pers. comm.)
East Arm QLG	K-Ar, biotite	108 ± 4	-	Mid-Cretaceous resetting, interpreted by Okulitch (see text)
Boss Mt. Gn.	Pb-U, monazite	116.7 ± 1.4 113.8 ± 2.0	- -	Possibly reset during metamorphism (Mortensen et al., in press)
East Arm QLG	K-Ar, hornblende	153 ± 8	-	Ref: Stevens et al. 1982
Western QLG	Rb-Sr, K-feldspar	276 ± 11	0.7372	K-feldspar phenocrysts Collected by author, analysed by Armstrong

Fig. 2.5

Rb - Sr isochrons for the Western (WQLG) and Eastern (EQLG) Quesnel Lake Gneiss, determined by R.L. Armstrong. Note that WQLG is Rb-enriched, while EQLG is Rb-poor. Decay constant = $1.42 \cdot 10^{-11} \text{ a}^{-1}$.

Okulitch 1985). Inherited Proterozoic components present in zircons from many of the these bodies, and Sr isotopic data both suggest a component of assimilated early Paleozoic or Precambrian crustal material (Mortensen et al., in press, Okulitch 1985). A Proterozoic Rb-Sr whole-rock date of 752 ± 76 Ma from the central segment of the Quesnel Lake Gneiss (Blenkinsop 1972), and a similar date obtained by combining Blenkinsop's data with the author's data (746 ± 56 Ma; Table 2-3) may reflect this postulated crustal assimilation. Similarly, Okulitch (1985) interpreted that while the Quesnel Lake Gneiss has a maximum intrusive age of 400 to 350 Ma, some discordant zircons were derived from Proterozoic sediments with a cratonic source, or from Precambrian basement partially melted at depth.

The other important information provided by the radiometric data pertains to the timing of metamorphism and other isotopic disturbances. A titanite U-Pb date of 174 ± 4 Ma from the East Arm Quesnel Lake Gneiss was interpreted by Mortensen et al. (in press) as a metamorphic age. A Rb-Sr date of 161 ± 15 Ma from matrix plagioclase from the author's Western Quesnel Lake Gneiss is similar, and also indicates mid-Mesozoic resetting (Armstrong, pers. comm. 1984). These two Middle Jurassic dates span the range in which the regional metamorphism in the southern Omineca Belt is generally considered to have occurred (see Pigage 1977, Archibald et al. 1983; see also Chapter 6.2, sub-section 19).

A biotite Rb-Sr date of 106 ± 3 Ma from the author's Eastern Quesnel Lake Gneiss agrees with a biotite K-Ar date of 108 ± 4 Ma from the East Arm of the Quesnel Lake Gneiss. The latter was interpreted by Okulitch (in Stevens et al. 1982, p.13) to reflect isotopic resetting due to Early to mid-Cretaceous plutonic activity in the region. Similar dates,

116.7 $\pm$ 1.4 Ma and 113.8 $\pm$ 2 Ma, were obtained from monazite from the Boss Mt. Gneiss but their significance is problematical (Mortensen et al., in press). Partial isotopic resetting of K-Ar in hornblende from the East Arm gneiss may account for an otherwise anomalous date of 153 $\pm$ 8 Ma (Stevens et al. 1982).

In summary, the Western- and Eastern Quesnel Lake Gneiss in the thesis area and in the Quesnel Lake region is provisionally interpreted to be mid-Paleozoic, probably Late Devonian or early Mississippian. The various gneiss bodies record evidence of Middle Jurassic metamorphism and later Mesozoic, possibly mid-Cretaceous, isotopic resetting.

WESTERN QUESNEL LAKE GNEISS

The main characteristic of this gneiss (unit DMWQg) is the presence of potassium feldspar megacrysts, in a generally mafic-poor matrix of quartz, plagioclase and potassium feldspar, and white mica mainly derived from the metamorphic alteration of feldspars. The protolith is presumed to be porphyritic granite, but it is not preserved. As a result of deformation and recrystallization, strongest at the contact with the Antler Formation, the rock is everywhere a granitic augen orthogneiss (Plates 3F, 4F). In this section, the composition of the gneiss is described; tectonic features and microstructures are described further in Chapter 3.2.

The Western Quesnel Lake Gneiss is essentially a tabular body which has been sheared, foliated and folded by deformations described in Chapter 3. The present thickness of the gneiss east of Cariboo River is about 400 to 700 m; the body west of the river is thicker due to folding

(see cross-sections).

Lithology

The Western Quesnel Lake Gneiss is moderately-well exposed, forming conspicuous white exposures in clearcuts, although they are commonly not exactly in place. The gneiss is generally pale- to medium grey, inequigranular, and is well- to extremely well foliated and lineated. White- to pale pink potassium feldspar megacrysts (porphyroclasts, originally phenocrysts) range from up to 18 cm long (typically 6 cm) in less deformed gneiss to rounded augen less than 1 cm long (Plates 1F,G; 3E; 4F). Some crystals are dark bluish-grey, probably because of internal strain. Some have quartz-filled extension fractures, possibly parallel to 001 cleavage planes. The potassium feldspar is mainly perthitic microcline. No plagioclase megacrysts were found, but there is plagioclase in the matrix. The potassium feldspar is generally altered to white mica to some degree, and it is difficult to assess its original percentage in the rock; staining suggests that it formed at least 30%. Locally, the retrogression to white mica is so advanced that few feldspars remain and the rock resembles a pale grey, fine grained micaceous schist.

The matrix consists of plagioclase, quartz, white mica, minor chlorite and clinozoisite, and potassium feldspar, some of which was disintegrated from megacrysts. The gneiss is generally very mafic-poor, but north of Sellar Creek several per cent of biotite are present. Black tourmaline occurs rarely. Porphyroclasts of blue to grey to milky white quartz are either primary crystals or remnants of transposed veins. Most quartz in the matrix, however, is in the form of fine- to very fine grained polygonized crystals in lenses, layers and ribbon

grains (Plates 3H; 4D-G).

Twinned and untwinned feldspar which does not stain by the cobaltinitrite method is presumed to be plagioclase. It is partly altered to white mica, but less so than the potassium feldspar. The plagioclase is albite, approximately An_{5-10} , from the Michel-Levy method, and probably formed about 10 to 20% of the original rock.

Sericite and muscovite, with or without minor biotite and chlorite, constitutes most of the remainder of the matrix. The proportion of white mica depends on the degree of retrogression of feldspar. It is generally in the form of dense, fine grained lenticular mats which define the foliation, within which relict potassium feldspar may be present. The micaceous foliae locally have a dull dark green cherty texture which could be mistaken for 'ultramylonite' (it could be called phyllonite, following Sibson 1977, and see Chapter 3.2); this material can be several centimetres thick, and very little of the potassium feldspar remains.

Contacts and inclusions

The contact between the Quesnel Lake Gneiss and the Snowshoe Group is sheared but was originally intrusive, from the evidence of tabular or lenticular pelitic or psammitic inclusions with wispy margins in the gneiss near their contact (corroborated by Struik 1983a). These inclusions, up to a few metres thick, are xenoliths or remnants of country rock in the marginal fringe of the intrusion. No inclusions of the Antler Formation occur in the Quesnel Lake Gneiss, however.

Dykes and veins

Small dykes, usually about 10 to 30 cm thick but rarely up to 1.0 m

thick, are sporadic in the gneiss (denoted on Map 1 as DMWQg1). Most were probably basaltic originally, but are now green to grey, fine- to medium grained, homogeneous amphibolite or chloritic greenstone. They vary from massive to well foliated, and generally have sharp and discordant contacts with the gneiss. Some dykes have been silicified and carbonatized, appearing as pale buff or creamy grey with a thick, rusty brown, limonitic weathering rind.

Locally there are 10 to 20 cm thick veins or anastomosing sheets in the gneiss which may have been primary aplitic or microgranite fractions. They are probably primary because they are oriented parallel to the foliation, are themselves foliated, and they have fairly diffuse contacts with the enclosing coarser grained gneiss.

EASTERN QUESNEL LAKE GNEISS

This gneiss (unit DMEQg) is restricted to the extreme southeast of the thesis area, and is rather poorly exposed (Figs. 2.1, 2.4). Its homogeneity and the presence of flattened schlieren (5 to 20 cm long, 1 to 3 cm thick) of fine grained schist, which appear to be xenoliths, indicate that this gneiss is meta-igneous. Based on the map pattern, the Eastern Quesnel Lake Gneiss is interpreted to be a 'solid' intrusive body, about 4000 m wide, with an unknown (vertical) thickness. However, the central segment of the Quesnel Lake Gneiss (Fig. 2.4) has been mapped by Fletcher (1972) who interpreted the body as an isoclinally folded granitic sill, with a screen of Kaza Group (= Snowshoe Group) metasediments in its core (Fig. 6.17); each 'limb' is about 1400 m thick. No such screen of Snowshoe Group was detected in the Eastern Quesnel Lake Gneiss to corroborate this.

Lithology

The Eastern Quesnel Lake Gneiss is homogeneous, medium grained, and streaked black and grey-white. It is well lineated and well foliated, defined by thin foliae about 1 mm thick containing biotite, muscovite, epidote and clinozoisite, which comprise about 25 to 30% of the rock. The remainder is composed of layers and lenses of quartz and feldspar, separated by the foliae. No (meta-)phenocrysts were seen.

The epidote and clinozoisite are metamorphic, not primary; they may have been derived from a primary amphibole or from plagioclase or both, but probably mainly the latter. The composition of the protolith is uncertain, therefore, but it is considered to be approximately granodiorite.

Quartz grains are inequigranular, ranging from less than 0.1 to 0.5 mm. Undulose extinction and deformation bands are moderately developed, but generally features of major strain and dynamic recrystallization are absent. However, the quartz is not strongly annealed: most grain boundaries are irregular or serrated from grain boundary migration, and few triple junctions are present.

Feldspar is recognized in thin section by mild alteration to sericite and by mechanical twinning. Like the quartz it is inequigranular, being finer grained where associated with micaceous and epidote-clinozoisite foliae.

Biotite is greenish-brown and mildly altered to chlorite. It is associated with muscovite, some of which forms larger porphyroblasts, and with granular aggregates or trains of epidote and clinozoisite. One epidote grain is cored by allanite.

Structure

Two steeply-dipping foliations of similar development are present in the gneiss; they probably represent the first and second phases of deformation, described in Chapter 3. One is usually predominant, and the other is separated by about 20 to 30°. Commonly their interference tends to diminish the foliated effect and the rock appears more lineated than foliated. There may be a stretching lineation present as well, roughly coincident with the intersection of the foliations, which plunges gently to moderately to the northwest. • No folds were seen.

[2] 3. ANTLER FORMATION (UNIT PPA)

This unit is the lowest in the Intermontane Belt in the thesis area and in the Quesnel Lake region (Figs. 1.2, 2.1, 2.6). The main rock types are greenstone, greenschist, amphibolite, serpentinite and steatite, and are usually easily distinguished from the metasediments and orthogneisses of the Omineca Belt. In the thesis area the unit is generally thin, ranging between about 800 m in the extreme west to only a few metres in other places; it is typically about 250 m thick. Because of its general thinness and readily weathered lithologies, the Antler Formation is rather poorly exposed, though its position can usually be located quite precisely from the occurrence of bounding rock units.

AGE AND CORRELATION OF THE ANTLER FORMATION

The Antler Formation in the thesis area is herein correlated with

Fig. 2.6

Schematic stratigraphic column showing distribution of facies in the three main units of the Quesnel terrane, including the Antler Formation. Thicknesses are approximate. The main points to be noted are the general coarsening upward composition of the Black Phyllite and Takla Group, and the gradual replacement of pelites by volcanoclastics and finally by volcanics, as the dominant deposit. Fig. 2.10 is intended to show that this facies change takes the form of a progradation of the Takla Group into the 'Black Phyllite basin' (see also Chapter 6.1).

the type-Antler Formation in the Cariboo Mountains to the north, and with the Fennell Formation, 150 km southeast of Quesnel Lake (Fig. 2.3; Table 2-2; Campbell & Campbell 1970, Campbell 1978, Campbell & Tipper 1971, Campbell et al. 1973, Struik 1983a). The author has examined the Antler Formation in the Crooked Lake area (see Campbell 1971) and it is identical to that in the thesis area. Similarly, the Antler and Fennell Formations throughout the region have been compared by R.B. Campbell (see K.V. Campbell 1971), Campbell (1971), Hall-Beyer (1976) and Montgomery (1978), and all these workers attest to the overall similarity of these occurrences, on the bases of age, lithology, and chemical composition.

The Antler Formation in the Cariboo Mountains was formerly mapped as the upper part of the Slide Mountain Group (Sutherland Brown 1957, 1963), and consists of an 1100 m thick succession of pillow basalt, ribbon bedded radiolarian cherts, argillite, and subordinate diorite, gabbro, ultramafics, volcanic breccia and greywacke (Campbell et al. 1973, Struik 1980). The lower part of the Slide Mountain Group as originally defined consists of polymict conglomerate, argillite, basic volcanics and limestone of the Mississippian to Permian Guyet Formation (Fig. 1.2), which is autochthonous and unconformable on rocks of the Cariboo Group and the Black Stuart Group. Monger (1977) incorporated the whole of the Slide Mountain Group as part of his "Eastern Assemblage" (see next sub-section). Campbell et al. (1973), however, were the first to speculate that the Antler Formation was separated by a thrust from the Guyet Formation, mainly because no feeders to the mass of pillow lavas are present in the underlying rocks, and this was corroborated by Struik (1981b) who demonstrated a stratigraphic inversion by finding lower Pennsylvanian and Lower Permian conodonts in

the Antler Formation cherts. As a result, the significance of the Antler Formation changed, and the terms "Slide Mountain Group" and "Eastern Assemblage" as originally used have been revised (see next sub-section).

The Fennell Formation (Fig. 2.3) is very similar to the type-Antler Formation, comprising low metamorphic grade pillowed metabasaltic greenstone, amphibolite, and interlayered bedded chert, argillite and phyllite (Campbell & Tipper 1971, Aggarwal et al. 1984). The age of the Fennell Formation is Upper Devonian to Upper Permian, based on the age of zircons in quartz porphyry intrusions, and on the age of conodonts in the sedimentary units (Preto 1981).

The Antler Formation in the thesis area (and in the Crooked Lake area: Campbell 1971) is apparently much more deformed than the type-Antler and Fennell Formations, and lacks many of their diagnostic features like pillow lavas and bedded cherts. Consequently, no independent age determination is possible, but it is likely that their correlation is valid (Campbell 1978, Struik 1983a). Thus, the age of the formation is assumed herein to be lower Pennsylvanian to Lower Permian, though the range may extend to the upper Mississippian and the Upper Permian.

The Antler Formation is also correlated with the Kaslo Group of southern British Columbia (Fig. 2.3, Table 2-2), another member of Monger's (1977) "Eastern Assemblage" or of his more recent (Monger 1984) "Slide Mountain terrane" (see next sub-section). Like the Antler and Fennell Formations, the Kaslo Group consists of tholeiitic pillow lavas and greenstone, serpentinite breccia and talc schist (Klepacki & Wheeler 1985). It is dated as Lower Permian from its conodont fauna (Orchard

1985) and from the U-Pb age of zircons (Klepacki & Wheeler 1985). The major contribution of Klepacki (1983) and Klepacki & Wheeler (1985) was that the Kaslo Group is in depositional contact with the underlying McHardy assemblage of the Milford Group (Fig. 2.3, Table 2-6), part of the continental margin-type stratigraphy of the southern Omineca Belt (see Chapter 6.3). In contrast, the Antler Formation in the Quesnel Lake region is everywhere structurally separated from the Omineca Belt (see Chapter 3).

A note on terminology

In an important paper, Monger (1977) emphasized the 'oceanic and volcanic' character of certain upper Paleozoic assemblages in the Canadian Cordillera. He placed the Slide Mountain Group in the Black Stuart synclinorium, which included the type-Antler Formation, into the "Eastern Assemblage" (Table 2-2).

After it was realized that the Antler Formation and other members of the Eastern Assemblage were actually separated from underlying rocks by major faults (Montgomery 1978, Struik 1980, 1981b, Rees 1981), these allochthonous rocks were distinguished as the Eastern terrane (Price et al. 1981, Monger et al. 1982). Monger (1984) renamed the Eastern terrane "Slide Mountain-terrane".

If the Antler Formation in the thesis area is correlated with the Antler Formation in the Black Stuart synclinorium, as is done here, then technically it is part of the Slide Mountain terrane. However, the Antler Formation in the thesis area is overlain (probably a deformed depositional contact) by rocks of the Quesnel terrane, thus linking them into a single composite terrane by the Upper Triassic at the latest (cf.

Monger 1984, Price et al. 1985; see Chapter 6.1). Terranes are supposed to be fundamentally in tectonic juxtaposition, so to suppress the implication that the Antler Formation in the thesis area is tectonically separated from the Quesnel terrane, it will be called simply that, rather than Slide Mountain terrane.

LITHOLOGY OF THE ANTLER FORMATION

The Antler Formation in the thesis area consists of three general rock types: greenstone, metagabbro, and meta-ultramafic rocks. Greenstone is usually present in exposures, commonly accompanied by variable amounts of meta-ultramafic rocks; metagabbro is less common.

Greenstone

The greenstones (subunit PPAb) vary from pale to dark bluish-green or greyish-green, depending on the amount of plagioclase and (or) carbonate present. They are fine- to medium grained and have a pervasive foliation (or cleavage) defined mainly by chlorite, giving the rock a fine fissility, or a flaggy parting on the scale of a few centimetres (Plate 11). The main constituents are chlorite and albite; epidote is usually present and locally it is a major constituent. Minor phases are clinozoisite, actinolite, biotite, magnetite, quartz, carbonate and pyrite. No hornblende or garnet was found. The plagioclase is commonly sericitized, and locally the biotite appears to be breaking down to chlorite, muscovite and magnetite. Various-sized veins of quartz ± epidote, and of carbonate occur in the greenstones.

A fine- (millimetre-) scale compositional banding is variably developed, defined by variations in the proportions of chlorite, albite

and epidote. Such banding is well displayed in the most deformed greenstones which are found at the contact with the Omineca Belt. It is the result of highly flattened, and transposed primary layering or perhaps of an early development of metamorphic segregation (Plate 3C). Otherwise the greenstones can be quite uniform and homogeneous. The rocks are completely metamorphosed: no relict olivine or pyroxene was found.

A coarser, centimetre-scale layering is also characteristic of the greenstones (Plate 1H). This also probably represents transposed primary layering or structures like pillows, veins or dykes, but this could not be proved. However, pillows were tentatively identified in a few localities. It is considered quite feasible that the colour and compositional banding in some greenstone exposures represents transposed metamorphosed pillows, pillow rims and inter-pillow selvages: the same feature has been documented in the Western Alps where the strain decreases into the pillowed protolith, proving the origin of the banding (Jean Bertrand [Universite de Geneve] pers. comm. 1981; see also Holland & Norris 1979). The author has seen this kind of transition in deformed pillow lavas in the Central Metasedimentary Belt of the Grenville Province in the Canadian Shield.

Lensoidal bodies, up to about 30 cm across, of bright green epidosite occur in the greenstone at a few localities (e.g. 750). Their origin is not known to the author; although their form resembles pillows, this possibility is not favoured.

The greenstones are locally altered by carbonatization to some degree, varying from small spots of rusty orange-brown carbonate porphyroblasts to the replacement of thin layers, podiform lenses, or the entire rock by carbonate. The carbonate may be ankeritic or ferro-

dolomite. Pyrite, later hydrated to limonite, is usually associated and fuchsite is developed locally. The alteration has proceeded to completion in areas of some exposures, producing a rusty carbonate-limonite schist with thin foliae of chlorite. This kind of alteration is usually found at the contact between the Antler Formation and the Black Phyllite (unit **uTrBR**) and locally with the Snowshoe Group (unit **HPS**).

Metagabbro

Metagabbro, and amphibole schist which may be metagabbro (subunit **PPAg**), are sporadic in the Antler Formation. Metagabbro is best exposed at locality 204 where it is layered: each layer, about 5 to 10 cm thick, consists of an amphibole-rich part grading into an altered plagioclase-rich part (Plate 2A); however, the direction of the top is not known. This structure closely resembles cumulate layered gabbros in ophiolite complexes and layered intrusions. The fairly-well preserved relict primary texture comprises (pleochroic) pale green, coarse (1 to 2 cm long) tremolite-actinolite (or Fe-poor actinolite) + chlorite laths after original pyroxene oikocrysts; there was probably an intermediate stage of hornblende, of which traces are visible. These crystals are intergrown with a white saussuritic 'matrix' consisting of a very fine grained matte of albite, epidote, clinozoisite and chlorite, after the original ?intermediate plagioclase. The rock is foliated but the texture is definitely a relict igneous one.

Speckled greenstones containing coarser amphibole (2 or 3 mm across) and altered plagioclase are thought to be also metagabbro because the texture is closer to porphyroclastic than porphyroblastic.

Meta-ultramafic rocks

The meta-ultramafic constituents of the Antler Formation include serpentinite, steatite (talc + carbonate rock), breccia comprising a mixture of these, and ultramafic schists.

Serpentinite (subunit PPAs) is dark green or indigo and weathers greenish-white or dun brown. It is very fine grained, hard and has a flinty texture. It has a patchy appearance due to the irregular replacement of primary minerals by serpentine. Locally a subtle compositional layering (≤ 5 cm) is present in exposures, suggesting a relict cumulate structure.

The serpentine is believed to consist mainly of lizardite, in mainly non-pseudomorphic form. Some mesh-textured pseudomorphs indicate derivation from olivine, but most appears to be lizardite bastite after olivine and pyroxene. No relict olivine or pyroxene was identified so it is not possible to determine the protolith (whether dunite, harzburgite or lherzolite). Locally, the lizardite is overgrown by randomly-oriented blades of what is interpreted to be antigorite, although this was not verified by X-ray analysis. Minor amounts of talc and carbonate are usually associated with serpentine, replacing it. Minor tremolite is probably derived from the CaO from clinopyroxene. Skeletal or anhedral chromite is present locally.

The serpentinite may be transected by veins of asbestiform chrysotile and asbestiform tremolite, between 1 mm and 20 cm thick. These minerals are also present as very coarse white or green fibrous sheaves in minor anastomosing shear zones in the serpentinite (e.g. locality 843).

Steatite (subunit PPAt) is the most common form of meta-ultramafic

rock in the thesis area. It nearly always accompanies the serpentinite, while usually occurring by itself. It is consistently composed of pale blue-grey very fine grained talc, with abundant orange-brown subequant porphyroblasts (up to 1 cm across) of magnesite (determined by X-ray diffraction in samples from several localities). Iron-bearing carbonate also occurs, judging by the limonite pits on weathered surfaces. The only other constituents are magnetite and chlorite. A moderate to strong undulose foliation is usually present. Large bodies of vein quartz are typically associated with steatite, and contain large crystals of talc and tremolite-actinolite; they are probably a by-product of metamorphic reactions. Similarly, Montgomery (1978) noted "quartz boulders" in the Kaza Group (cf. Snowshoe Group) immediately below ultramafic rocks of the Black Riders Complex; he related them to metasomatic interactions between the metasediments and the ultramafic rocks.

Steatite identical to that described above (but without other Antler Formation lithologies) occurs in scattered isolated lenses, a few metres to a few tens of metres long, within the Snowshoe Group within about 200 m of the contact with the Antler Formation. Exactly the same thing was found by Campbell (1971) in the Crooked Lake area, and by Montgomery (1978) in his area. Nowhere deeper ~~in the Snowshoe Group~~ mapped by the author was steatite found. It is concluded that ~~this~~ steatite is Antler Formation. Its incorporation into this structural position is discussed in Chapter 6.2 (Fig. 6.8).

Ultramafic breccia is associated with serpentinite and steatite in a few localities. The best examples are at locality 28 (UTM 0597058411) where clasts and blocks from less than 1 mm to 50 cm or more across of dark green serpentinite are supported in a matrix of very fine grained

carbonate (Plate 2B). The serpentinite clasts are subangular to rounded, and are composed mainly of lizardite, partly altered to tremolite. While the matrix is primary, probably authigenic or biogenic carbonate, some carbonatization of the serpentine is secondary (i.e. metamorphic; see Chapter 4.2).

An important feature is that part of this breccia almost completely lacks foliation or flattening and preferred orientation of clasts. For this reason the origin of at least some of the breccias (possibly all of them) is interpreted to be sedimentary rather than tectonic (as in melange or fault breccia). In support of this, this breccia is apparently virtually identical to those found in ophiolites in the Mt. Geneve Massif in the Western Alps which were interpreted to have formed in a submarine talus slope composed of serpentinitized upper mantle detritus shed from an active fault or fracture zone on the flank of an oceanic spreading centre (from discussion with Jean Bertrand, Université de Geneve, 1981). Furthermore, the breccia in Plate 2B is practically identical in appearance and composition to that described in Bonatti et al. (1974, their Fig. 2) which was dredged from a fracture zone offset in the Mid-Atlantic Ridge. They interpreted the rock as either fault breccia formed by shear fracturing, or fault zone talus.

The matrix of the breccia at locality 372 is, however, well- but irregularly foliated, and the creamy-grey carbonate, talc and tremolite matrix is scored locally by slickensides or chrysotile fibre growth. These latter features are not pervasive nor consistently oriented; they probably formed during Jurassic deformation as the matrix flowed heterogeneously around the serpentinite clasts.

Ultramafic schists are restricted varieties of the Antler Formation.

They include foliaceous talc-chlorite schists (with minor carbonate) and tremolite-talc schists. Greenish tremolite or Fe-poor actinolite occurs in coarse grained lenses or in radiating clusters in a matrix of talc. These rocks are probably meta-serpentinites, although little or no serpentine is preserved.

CHEMICAL ANALYSIS AND TECTONIC SETTING OF THE ANTLER FORMATION

Chemical analyses were obtained from representative rocks from fifteen Antler Formation localities. These comprise major and trace element analyses of six greenstones, two metagabbros, one ultramafic rock (steatite) and one mafic metasediment, and selected trace element analyses of five other greenstones (Tables 2-4 and 2-5). This is obviously not an exhaustive geochemical study, but serves to compare the chemical composition of the Antler Formation in the thesis area with analyses from other occurrences of the unit which have been documented more fully by other workers (see below).

All the sampled rocks are metamorphosed, and minor weathering alteration on some samples could not be avoided. Consequently, the major element analyses should be regarded with caution as some elements are mobile during metamorphism. Bearing this in mind, all the greenstones are metabasalts, and they and the metagabbros are subalkaline and tholeiitic according to the alkali-silica and AFM variation diagrams (Figs. 2.7 and 2.8).

Certain minor and trace elements such as Ti, Zr, Y, Cr and Nb are considered more reliable indicators of original igneous composition because they are much less mobile during metamorphism than other elements. Empirical discrimination diagrams using these elements have

TABLE 2-4: MAJOR ELEMENT CHEMICAL ANALYSES OF THE ANTLER FORMATION IN THE THESIS AREA

	1	2	3	4	5	6	7	8	9	10	11
SiO ₂	41.2	50.2	49.0	50.1	49.6	49.2	48.6	46.9	41.0	40.2	48.2
TiO ₂	1.57	1.53	1.45	1.82	2.03	1.81	0.47	0.22	0.03	1.33	1.7
Al ₂ O ₃	15.1	15.7	16.0	17.1	14.7	15.1	15.1	12.8	1.2	29.7	15.6
Cr ₂ O ₃	0.02	0.02	0.02	0.01	0.02	0.02	0.03	0.12	0.26	0.03	0.02
Fe ₂ O ₃	10.8	8.3	9.4	5.0	5.8	5.3	2.6	3.4	8.0	2.5	7.4
FeO	2.0	4.2	5.8	7.6	6.8	6.6	6.1	5.7	-	5.1	5.5
MnO	0.2	0.21	0.18	0.13	0.12	0.23	0.16	0.18	0.08	0.09	0.18
MgO	9.57	5.05	5.09	4.08	3.97	6.76	9.29	14.2	33.9	6.2	5.75
CaO	4.95	5.37	3.49	3.79	6.45	6.97	10.1	9.03	0.32	0.41	5.17
Na ₂ O	2.7	3.3	3.7	4.4	3.4	3.5	0.9	0.9	0.0	0.1	3.5
K ₂ O	0.21	0.36	0.09	0.07	0.87	0.11	0.04	0.06	0.05	6.92	0.28
H ₂ O	4.9	3.9	3.5	3.5	3.0	3.5	4.8	5.4	7.2	7.0	3.7
CO ₂	6.2	0.1	0.8	0.8	2.1	0.2	0.4	0.1	7.7	0.1	1.7
P ₂ O ₅	0.14	0.17	0.13	0.26	0.24	0.18	0.03	0.01	0.0	0.23	0.19
S	0.09	0.05	0.06	0.06	0.06	0.08	0.06	0.08	0.05	0.08	0.07
Total	99.7	98.5	98.7	98.7	99.2	99.5	98.7	99.1	99.8	100.0	99.0

Notes: Values in weight per cent. Columns 1-6 = greenstones (metabasalts), localities 28, 100, 124, 128, 144, 454, respectively; columns 7-8 = metagabbros, localities 187 & 204, respectively; column 9 = meta-ultramafic (steatite), locality 28; column 10 = mafic metasediment, locality 256; column 11 = mean for six greenstones. All analyses by X-ray diffraction fused, except FeO, H₂O, CO₂ and S by rapid chemical.

TABLE 2-5: TRACE ELEMENT ANALYSES OF THE ANTLER FORMATION IN THE THESIS AREA

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
Rb	9	14	3	3	15	1	2	n.d.	1	255	-	-	-	-	-
Sr	59	107	61	140	207	100	108	165	2	73	-	-	-	-	-
Ba	70	130	70	60	570	80	50	40	30	2410	-	-	-	-	-
Y	40	42	36	64	46	44	14	6	1	42	10	32	22	29	21
Zr	107	117	89	223	145	131	27	9	1	417	31	76	179	76	72
Nb	3	3	2	5	3	3	2	n.d.	2	29	-	-	-	-	-
Cr	tr	tr	tr	tr	tr	tr	300	1200	2600	300	115	144	226	184	70
NI	170	120	100	110	120	120	180	420	2520	170	-	-	-	-	-
Zn	170	190	180	130	180	180	120	130	130	150	-	-	-	-	-
Mo	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	1	1	1	n.d.	-	-	-	-	-
As	16	22	10	9	13	4	24	11	20	8	-	-	-	-	-
Br	1	2	n.d.	1	1	1	1	2	1	n.d.	-	-	-	-	-
Pb	12	12	11	13	14	11	12	16	8	9	-	-	-	-	-
Th	6	4	3	4	4	5	8	5	2	23	-	-	-	-	-
U	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	3	3	2	8	-	-	-	-	-
(TiO ₂)	1.57	1.53	1.45	1.82	2.03	1.81	0.47	0.22	0.03	1.33	0.58	1.36	2.19	1.2	1.0)

Notes: All values in parts per million, except TiO₂ in weight per cent (values in columns 1-10 repeated from Table 2-4). - = not analysed, n.d. = not detected, tr = trace amount. Columns 1-6 & 11-15 = greenstones (metabasalts), localities 28, 100, 124, 128, 144, 454, 372, 457, 696, 956, 1107, respectively; columns 7-8 = metagabbros, localities 187 & 204, respectively; column 9 = meta-ultramafic (steatite), locality 28; column 10 = mafic metasediment, locality 256.

Fig. 2.7

Alkali - silica diagram for Antler Formation greenstones (solid circles) and metagabbros (open circles). Dividing line from Irvine & Baragar (1971).

Fig. 2.8

(a) AFM diagram for Antler Formation rocks. A = $\text{Na}_2\text{O} + \text{K}_2\text{O}$, F = $\text{FeO} + (0.8998)\text{Fe}_2\text{O}_3$, M = MgO . Dividing line from Irvine & Baragar (1971). Solid circles = greenstones, open circles = metagabbros, triangle = ultramafic (steatite). Note probable tholeiitic differentiation trend. (b) TiO_2 - P_2O_5 - K_2O diagram for Antler Formation greenstones and metagabbros; fields from Pearce et al. (1975). Symbols as for (a). Helps to distinguish oceanic from continental basalts.

been devised for basaltic rocks from known tectonic environments (e.g. Pearce & Cann 1973, Floyd & Winchester 1975, Perfit et al. 1980), and the eleven greenstones have been plotted in several of these diagrams (Fig. 2.9). The plots confirm that the greenstones are oceanic tholeiitic metabasalts (especially Fig. 2.9a). Two of the samples (localities 128 and 696) have a tendency to plot in the 'within-plate' field, which is the typical domain of alkali basalts, apparently by virtue of their unusually high Zr contents; however, all other indications are that the metabasalts are not alkalic.

Subalkaline oceanic tholeiites may belong to the ocean floor, abyssal or mid-ocean ridge type, or they may belong to the low-potassium type found in island arcs. Still other oceanic tholeiites, such as back-arc basin basalts and fracture zone basalts, tend to have ambiguous characteristics. Relevant to this study is whether the Antler Formation metabasalts represent ocean floor or island arc volcanism. The trace element data (Fig. 2.9) is biased towards an ocean floor origin, although locality 372 usually plots in the island arc field due to unusually low Ti, Zr, Y and Cr. [This sample is lithologically different from most of the others, and it may not be a volcanic but possibly a plutonic rock; it does tend towards where the metagabbros plot.] The otherwise consistent ocean floor trend of the data is largely due to the TiO_2 values which fall between 1.0 and 2.2%, a range characteristic of ocean floor basalts, while the vast majority of arc tholeiites (and calc-alkaline basalts) contain less than 1.1% TiO_2 (Pearce & Cann 1973, Perfit et al. 1980), and oceanic island (alkalic) basalts contain at least 2% TiO_2 . TiO_2 is a good discriminant. Similarly, the recorded Y values at between 21 and 64 ppm are higher

Fig. 2.9

Trace element variation in Antler Formation greenstones: empirical discrimination diagrams indicating tectonic setting.

(a) Ti-Y-Zr diagram from Pearce & Cann (1973). (b) Ti-Sr-Zr diagram from Pearce & Cann (1973). Note that this diagram is suitable only for unaltered basalts, as Sr tends to be mobile. Some Sr depletion is suggested here.

Fig. 2.9 (continued)

(c) TiO_2 -Zr diagram from Pearce (1980). (d) Cr-Y diagram from Pearce (1980). Helps to distinguish ocean floor from island arc basalts by their higher Y values.

Fig. 2.9 (continued)

(e) Ti-Cr diagram from Pearce (1975). (f) Zr/Y-Zr diagram from Pearce (1980). Zr/Y > 4 separates within-plate basalts from ocean floor basalts, while lower Zr values distinguish island arc tholeiites. (g) TiO₂-Zr/P₂O₅ diagram from Winchester & Floyd (1976).

Fig. 2.9 (concluded)

(h) TiO_2 -Y/Nb diagram from Floyd & Winchester (1975). Y/Nb > 1 separates tholeiitic from alkali basalts. (i) Nb/Y-Zr/ P_2O_5 diagram from Floyd & Winchester (1975). Vertical trend for alkali basalts, horizontal trend for tholeiitic basalts. (j) P_2O_5 -Zr diagram from Winchester & Floyd (1976). Supposed to be a good indicator of magma type.

than for arc tholeiites.

Other geochemical studies of the Antler and Fennell Formations have been carried out by Campbell (1971), Hall-Beyer (1976) and Aggarwal et al. (1984), and their results are worth summarizing here. Campbell (1971) included samples from the type-Antler Formation and the Fennell Formation as well as from his own Crooked Lake area. His results consistently indicated oversaturated to undersaturated tholeiitic metabasalt, with a few mildly alkaline samples. Again, with few exceptions, TiO_2 values fall between about 1.4 and 2.0%. He did not obtain trace element data.

Hall-Beyer (1976) made a rigorous study of metavolcanics from the Antler-Fennell belt, concluding that most are ocean floor tholeiitic basalts with some komatiitic varieties. He found virtually no tendency towards island arc or alkalic basalts. The trace element values of his tholeiitic suite plot entirely within the field of ocean floor basalts in the Ti-Zr-Y diagram of Pearce & Cann (1973), practically duplicating the results of this thesis. The TiO_2 range is the same as for Campbell (1971): 1.4 to 2.0%.

In a review, Souther (1977) noted that analyses from volcanic rocks of the "Slide Mountain Group" fall into the "tholeiitic field typical of oceanic ridge basalt".

The only study to differ from the above conclusions is Aggarwal et al. (1984) who interpreted the metabasalts as originally alkalic or transitional basalt, possibly formed in an oceanic island setting. This was based on the chemical composition of primary pyroxenes and amphiboles (kaersutites) from a single suite of samples from the eastern margin of the Fennell Formation. These authors conceded that whole-rock

major element and most trace element data, including that for the kaersutite-bearing rocks, plot according to tholeiitic basalt, but argued that this is an effect of chemical mobility. They concluded that at least this part of the Fennell Formation does not represent ocean floor formed at a mid-ocean ridge (despite the recorded range of TiO_2 , i.e. 1.6 to 2.2%, which is lower than typical oceanic island basalts), but represents basalt produced at an off-axis or mid-plate oceanic island or fracture zone, or possibly a back-arc basin.

[2] 4. BLACK PHYLLITE (UNIT uTrBP)

This unit is here given the informal name Black Phyllite after the dominant lithology. Although these rocks have been mapped by several previous workers in the region, no formal name has been applied to them.

The Black Phyllite occurs in a northwest-southeast trending belt from Prince George to the Raft batholith (Fig. 2.3). The unit lies between the underlying Antler Formation and equivalents to the northeast and the overlying Takla Group to the southwest (Figs. 2.1 and 2.2). Campbell & Campbell (1970), Campbell & Tipper (1971), Campbell (1971), and Campbell et al. (1973) believed that the Black Phyllite was originally unconformable or disconformable on the Antler and Fennell Formations, although contacts are generally masked by deformation. Most of these workers agreed that the top of the Black Phyllite is conformably overlain by sedimentary and volcanoclastic facies of the Upper Triassic to Lower Jurassic Takla Group (Section 5); the author also agrees. On these bases, the present thickness of the Black Phyllite in the thesis area is approximately 2500 m. Its original

thickness, before tectonic thinning, is roughly assessed to have been 3000 m.

The Black Phyllite stratigraphy cannot be defined very well because of the poor exposure and strong deformation of these incompetent rocks. Nevertheless, according to the structure in cross-section K-K'-K'', this section through the Black Phyllite is basically a stratigraphic one, apparently lacking major fold repetition, although the existence of thrusts cannot be ruled out. Accepting this, the subordinate, volcanoclastic rocks north of Spanish Lake (subunits uTrBPf and uTrBPc, Fig. 2.6; Map 1) are part of the Black Phyllite despite their resemblance to the Takla Group. The change that occurs in the Spanish Lake valley from dark argillaceous rocks southwestwards to volcanoclastic and tuffaceous rock types is taken as the stratigraphic transition between the top of the Black Phyllite proper and the base of the Takla Group, respectively (Fig. 2.6).

AGE AND CORRELATION OF THE BLACK PHYLLITE

The age of the Black Phyllite is mainly Upper Triassic (Karnian to early Norian, Campbell & Tipper 1971, Campbell 1978); Struik (1986b) noted that Middle Triassic conodonts have been recorded. The fauna has Tethyan affinities (Campbell & Tipper 1971).

The Black Phyllite is probably correlative with the Slocan Group in the southern part of the Quesnel terrane in British Columbia (Fig. 2.3; Table 2-2), on the basis of age, lithological similarity and stratigraphic relationships. The Slocan Group is a thick unit of mainly grey phyllite or slate, with local interbedded siltstone and limestone,

and lies unconformably on the Kaslo Group (Read & Wheeler 1976, Klepacki 1983). Conodonts extracted from the limestone are Karnian and Norian (Orchard 1985).

LITHOLOGY OF THE BLACK PHYLLITE

The following descriptions are based on rock types within the outcrop of the Black Phyllite throughout the study area, not only on the sequence in cross-section K-K'-K" which is not necessarily typical.

Subunit uTrBPp. The Black Phyllite is predominantly composed of charcoal black, dark grey or pale bluish-grey, very fine grained, well-cleaved carbonaceous argillite and siltstone (Plate 2C). The metamorphic grade is low (middle greenschist and lower). The term 'phyllite' is often used and is retained to describe these rocks because of the sheen on some surfaces (Plate 2C), even though it is due to the very strong cleavage and disseminated graphitic material rather than to a phyllite grade micaceous fabric, which is restricted to a small higher grade area (garnet zone) in the extreme northeast of the outcrop (Fig. 4.1).

Bedding is defined by the millimetre- to centimetre-scale alternation (not rhythmic) of the black argillite and fine grained, usually paler grey and brown-weathering, well sorted siltstone or fine grained sandstone. Beds of siltstone range in thickness from less than 1 mm to about 10 cm, and are usually discontinuous or lenticular, mainly due to tectonic transposition. Thicker beds (20 cm to 50 cm) with flaggy micaceous partings are not common. No persistent marker horizons were found. Some of the siltstone beds, especially the thicker ones, have a calcareous cement; a stronger reaction to dilute HCl indicates

impure limestone, but this is rare. No rhythmic bedding sequences were noted.

The fine grain size and scale of bedding suggests a low energy environment of sedimentation. Current-produced structures were observed in only one locality (46): a thin (3 cm) truncated cross-bedded sandstone, indicating an at least transient higher energy regime. The same locality also reveals possible soft-sediment deformation in the form of ripped up siltstone beds with internal laminations oblique to the general bedding.

Characteristics of the Black Phyllite are its strong cleavage-bedding fissility, and orange-brown limonite in spots, pits, fine fractures, and as a general weathering stain; both render the rocks brittle and friable where the rocks are not recrystallized. The limonite was produced by the oxidation of pyrite which was widely distributed, in varying amounts; pristine pyrite is rare. Limonitic cubic moulds range from 1 mm to 10 mm. Locally the rusty alteration is so pervasive that no structures in the rocks are discernible. The preserved cubic form of the pyrite and limonite suggests that they postdate most ductile deformation. Syn-metamorphic quartz veins are also characteristic of the Black Phyllite: they occur on all scales as transverse fracture fillings in siltstone beds or as highly transposed, folded and boudinaged layers in argillaceous beds. Visible gold was found in one large quartz vein at the west end of Spanish Lake: dendritic filaments of gold line the limonitic cavity of a former pyrite crystal. The vein is known to prospectors.

Subunits uTrBPc and uTrBPf. The only other notable rock types within the Black Phyllite are the volcanic conglomerates and the

fragmental schists of subunits uTrBPc and uTrBPf, respectively. The volcanic conglomerate (subunit uTrBPc) occurs in a few lenticular outcrops up to about 4 km long and 1 km wide, northeast of Spanish Lake. The clasts in this mainly matrix-supported conglomerate are exclusively volcanic and range from about 20 cm across to sand-size (Plate 2D). They consist of porphyritic (pyroxene), and feldspar-phyrlic vesicular and amygdaloidal (meta-)basalt or andesite. Rarely they are zoned (Plate 2D), and these may be pyroclastic bombs. Other clasts may be flow breccia. Some vesicles in the clasts are contorted, suggesting that those clasts were still viscous when deposited and transported. Most clasts are rounded, but the preservation of vesicular rims suggests that abrasion, and perhaps transportation, were limited. The clasts are usually strongly flattened parallel to the dominant foliation, which may contribute to a locally framework-supported appearance.

The matrix of the conglomerate is pale green to creamy grey, and consists exclusively of pyroclastic and hyaloclastic tuffaceous material including poorly sorted lapilli-sized fragments and crystals. The matrix has essentially the same composition as the clasts.

Subunit uTrBPf is a pale green to pale grey or buff, fine- to medium grained chloritic schist containing highly flattened incompetent fragments up to 2-3 cm, probably of volcanic origin. It is usually altered somewhat, with calcite porphyroblasts and limonitic alteration. This rock is clearly a matrix-rich equivalent of the volcanic conglomerate, subunit uTrBPc, where the two are associated; isolated occurrences of similar rocks are on structural trend with uTrBPc and are mapped as uTrBPf.

Other subunits. Other lithologies in the Black Phyllite are minor.

White, quartz-feldspar rhyolitic porphyry (subunit uTrBP1) was found in a few localities. It is inequigranular, massive or foliated, with quartz (up to 10 mm) and white or pink feldspar phenocrysts (up to 5 mm). These intrusions are usually a few metres thick, and are structurally concordant with the Black Phyllite. No significant contact metamorphic effects or chilled margins were seen, though the contacts are mildly sheared and altered.

Altered and carbonatized rocks (subunit uTrBPd) occur within possible Black Phyllite on the west shore of south Cariboo Lake (see Chapter 3.6). These are very similar to subunit HPSd, and so are inferred to be altered basic dykes or sills.

Rare, green, medium grained chloritic greenstone in thin lenses within the Black Phyllite are probably small, transposed mafic apophyses. Bright green, fuchsite micaceous foliae associated with quartz veins are present locally, especially near the contact with the Antler Formation.

INTERPRETATION OF DEPOSITIONAL ENVIRONMENT OF BLACK PHYLLITE

The Black Phyllite was derived from carbonaceous mud and silt and minor fine sand, which was slightly calcareous locally. The flow regime was very low energy, indicated by the extreme rarity of current-induced structures and bedforms, and the absence of (upper flow regime) parallel laminations in siltstones. It appears that sedimentation, as far as can be deduced through the effects of deformation, took the form of a rain of very fine clastic material and planktonic algal remains through the water column. The organisms exhausted the oxygen in the water; the oxygen was not replenished efficiently by oceanic circulation (see

Tourtelot 1979). These features attest to a generally stagnant, euxinic basin. The paucity of carbonate suggests that the Black Phyllite accumulated below the CCD; fairly deep water is implied, although the CCD is shallower in small versus large ocean basins (Jenkyns 1978).

The original silt of the Black Phyllite could have come from a distant mature erosion surface, such as the speculative metamorphic basement of the Quesnel terrane inferred to lie to the west, or from the Omineca Belt continental margin to the east (see Chapter 6.1). The silt represents detritus carried by surface currents from a distant source, and which settled through the water column in the 'Black Phyllite' basin along with planktonic remains (see Fig. 2.10 in Section 5). This may apply also to the thicker siltstone beds, or alternatively these are the D components of infrequent turbidity currents, the 'background' silt and mud representing the E component. In general, however, the Black Phyllite does not particularly resemble turbidites. Influxes of ash from volcanoes in the incipient Quesnel Belt (Takla Group; see Section 5) are another possible source for the silt, especially the larger silt beds.

The volcanic conglomerate of subunit uTrBPc is interpreted as a pyroclastic flow (not an ash flow tuff) or agglomerate. (This is similar to a debris flow or lahar, but differs in that the clasts are not so heterolithic, that there are pyroclastic constituents, and that there was probably limited transportation during deposition.) The picture is incomplete because no lava, pillowed or otherwise, was found in situ to indicate the source of the volcanic clasts. It is not known whether the deposition was entirely subaqueous, or a subaerial eruption which flowed into water. Final deposition was probably submarine

because of the association with subunit uTrBPp, but it is emphasized that the nature of the contact between these units is not definitely depositional, although there is a transitional lithology locally to suggest that it was. A possible scenario is that a volcano composed of basaltic to ?andesitic vesicular and amygdaloidal lava flows, flow breccia, pyroclastic bombs and lapilli tuff built up, possibly but not necessarily to above sea level. The edifice of volcanics and tephra collapsed periodically due to instability, and settled in the water after only limited transportation and sorting. Finer material flowed farther away in a dense suspension to form subunit uTrBPf.

The ?subaerially formed subunit uTrBPc is at variance with the postulated deep water environment of the ambient uTrBPp phyllite. Either the phyllite is not so 'deep water' or, more likely, the sea floor was locally and (or) temporarily elevated by volcanism.

In conclusion, the depositional environment of unit uTrBP is considered to be a barred oceanic basin with restricted circulation and limited current activity, in which was deposited carbonaceous, hemipelagic mud and silt derived from a distant ?terrigenous source. The floor of the basin was the upper Paleozoic Antler Formation. The basin was barred by the incipient Quesnel Belt to the west. Most sedimentation occurred below the CCD, which was not particularly deep (perhaps less than 2000 m). It is inferred that the sediment was derived from the Quesnel terrane (its basement or overlying volcanics) or possibly in part from the continental margin. The volcanic subunits may have been 'satellites' of the Quesnel Belt.

[2] 5. TAKLA GROUP (UNIT TRJT)

This is the youngest major unit of the Intermontane Belt in the thesis area. It occurs in the southwest of the thesis area, part of a belt of Upper Triassic to Lower Jurassic volcanic, sedimentary and plutonic rocks that extends from north-central British Columbia to the U.S. border (Fig. 2.3). This belt forms the superstructure of the Quesnel terrane and will be called the Quesnel Belt herein (Table 2-2).

Accepting the definition of the base of the Takla Group given in Section 4, about 9000 m of the unit is present in the study area as a result of the steep dip of the strata (see cross-section K-K'-K'', Fig. 2.2). This figure ignores the possibility of tectonic repetition, but there is no particular evidence for that (see Chapter 3). The exposure over most of the outcrop area is poor, but is good on logging roads and clearcuts from Spanish Lake to Mt. Warren, affording a fairly complete stratigraphic section.

AGE AND CORRELATION OF THE TAKLA GROUP

The Takla Group as used in this thesis is correlated with the mainly Upper Triassic Nicola Group of southern British Columbia (Campbell & Tipper 1971), and with the Takla Group in the Fort St. James area, northwest of Prince George (Fig. 2.3), which ranges from Upper Triassic to Lower Jurassic. Campbell (1978) and Tipper et al. (1981) made the same correlations, and designated the rocks in the thesis area Quesnel River Group. However, the term Takla Group is preferred herein because of its wider usage. Takla Group is preferred over Nicola Group because the latter is generally applied to the Quesnel Belt in the Okanagan area

2

MicroD

of southern British Columbia, while Takla Group is applied to their presumed equivalents farther north in the belt (Souther 1977, Davis et al. 1978). (The rocks have also been called the Quesnel Trough.) The Takla Group may also be equivalent to the Lower Jurassic Rossland Group in the extreme southeast of the Quesnel terrane (Fig. 2.3; Beddoe-Stephens & Lambert 1981, Tipper 1984).

No datable fossils were found by the writer. The age control is provided by Bailey (1978) from fossils in the Morehead Lake area: his "unit 2" is Norian (uppermost Triassic) up to the base of the Jurassic, and is probably equivalent to most of unit TrJT based on lithological comparisons and structural extrapolation.

LITHOLOGY OF THE TAKLA GROUP

As explained in Section 4, the map pattern of the Takla Group in the thesis area is effectively a steeply dipping stratigraphic section beginning at the top of the Black Phyllite in the northeast around Spanish Lake, and younging to the southwest. This section reveals a gradual change from the predominance of Black Phyllite-type phyllitic argillites and siltstones in the northeast to rocks containing less pelitic material and more, coarser grained, more thickly bedded siltstones, sandstones and fragmental rocks composed of volcanogenic detritus. The stratigraphically highest rocks are augite porphyry lava flows on Mt. Warren, which are typical of the volcanics in the main part of the Quesnel Belt. This sequence is typical of the region: a very similar transition is described by Campbell (1971) in the Crooked Lake area, and the rocks also closely resemble Campbell & Tipper's (1971)

description of the Nicola Group in the Bonaparte Lake map area. A previous interpretation by the author (Rees 1981) that the Black Phyllite and Takla Group were in fault contact, is incorrect; this was inferred from the observed strain contrast between them, but this contrast is not as abrupt as first thought, and is merely due to heterogeneous deformation in the Black Phyllite and basal Takla Group lithologies.

The rock types are described in general order of their occurrence from their base in the northeast, up stratigraphic section to the southwest (Fig. 2.6).

The lower subunits (below subunit TrJTg, Fig. 2.6) are characterized by a variety of grey to black (meta-) argillites and siltstones (subunit TrJT_a), associated with fragmental rocks (subunit TrJT_f), and minor mafic volcanics (subunit TrJT_v). (Recall that these rocks differ from the Black Phyllite by being more heterogeneous and generally less pelitic.)

Subunit TrJT_a. The rocks in this subunit are generally grey-black to very pale grey, well sorted, very fine grained (phyllitic) argillites and siltstones (Plate 2E). The rocks weather grey to creamy white, with a general rusty stain from widespread limonite in small pits and fractures. The argillites are usually fairly graphitic and well cleaved, and resemble the Black Phyllite although they are more friable and the fissility is irregular. The siltstones tend to be massive or may be internally disintegrated to the point of brecciation. A cherty texture is quite common in some argillites, although such rocks are actually composed of very fine micaceous material. The paler grey siltstones are composed of fine quartz and feldspar that may be

tuffaceous. Some lithologies are rich in limonite and carbonate, unfoliated, and are probably altered basaltic dykes or lava flows.

Bedding is defined by fine compositional laminations, and by flaggy, uneven beds of siltstone up to 10 to 20 cm thick. Beds containing granules, graded bedding and channels are rare, and generally the primary structures indicate low energy deposition of distal clastic material, partly from volcanic sources, judging from the tuffaceous rocks. Extremely heterogeneous deformation of these incompetent rocks has locally disrupted bedding, producing a phacoidal or flaser structure resembling tectonic melange.

Subunit TrJTf. This subunit is similar to subunit TrJTg (see below) but since it is associated with and occurs within the outcrop of subunit TrJTg (Fig. 2.6), it is distinguished from TrJTg.

The rocks are mainly coarse grained fragmental wackes. They are usually pale to mid-grey, and weather creamy-buff to grey-white. They are moderately well sorted, with the grain size of the matrix ranging from very fine grained silt to sand, enclosing rounded to angular fragments up to several centimetres in size. The matrix forms at least 80% of the rocks. The clasts consist mainly of lithic fragments, with some mineral grains; many are altered. Some are very fine grained and highly flattened, and these might have been lapilli or coarse volcanic ash. Some exposures are homogeneous and unstratified, but usually some bedding is present (similar to subunit TrJTg, see below), ranging from fine lamination to beds a metre thick. Locally a weak to moderate cleavage is developed, but generally this subunit is not foliated, unlike the cleaved argillites of subunit TrJTg which may form part of the same exposure.

The rocks in subunit TrJTf probably formed underwater under moderate to high energy conditions, from the local evidence of scouring and truncations (from which younging was determined). The mineral clasts are finer and less mafic than subunit TrJTg, and may be largely volcanic ash or reworked pyroclastic deposits.

Subunit TrJTv. This subunit consists of mafic pillowed volcanics, and is restricted to a few isolated occurrences within subunit TrJTg. The presumed conformable contact between them was not observed but there is evidence of mixing of the deposits suggesting that it is a primary, if deformed association.

The pillows and massive volcanics are very pale green, very fine grained, and strongly fractured. Part may be flow breccia. The pillows are 20 to 30 cm across, usually rounded and lenticular, and flattened in the cleavage. The pillow selvages and matrix are pale to dark grey-green, fine grained to fragmental, with a cleavage or flaser fabric. The fragments are composed of flattened hyaloclastite and pillow breccia. Numerous creamy white weathering, coalescing variolites up to about 2 cm across indicate the former presence of volcanic glass. The volcanics are altered to a dark matrix of fine micas and iron oxide, with epidote, clinozoisite, calcite, and ?tremolite-actinolite.

Subunit TrJTg. This subunit is composed of pale to mid-greenish grey, moderately sorted volcanic wackes. Some isolated lithic clasts are 1 or 2 cm long but generally the coarsest fraction is about 5 mm and less, and comprises subangular to rounded, green pyroxene and creamy white sericitized plagioclase crystals and fragments of volcanic rock including variolitic glass. Quartz grains are conspicuously absent.

The matrix is fine grained, pale greenish grey, and includes altered brownish basaltic glass (palagonite), hyaloclastic debris, and very fine grained carbonate cement. Some delicate clast shapes imply limited transportation and attrition. The alteration permeates across clast-matrix boundaries which might indicate that this mixture was hot when deposited.

Also included in subunit TrJTg are black and rusty siltstones which resemble parts of the Black Phyllite, but differ in containing fragmental, volcanic detritus.

Rocks of subunit TrJTg are usually homogeneous, non-foliated and unstratified or display crude bedding locally, but they grade into medium- and fine grained sandstone and siltstone of the same composition which are better sorted and bedded (subunit TrJTb, see below).

Subunit TrJTg is associated with and is probably derived from basaltic to andesitic flows (represented by subunits TrJTv and TrJTb, see below) and volcanic conglomerates (subunit TrJTc, see below). The large volume of rock and crystal detritus may have been produced by the syn-eruptive disintegration of the volcanics as a result of the explosive interaction between subaqueous lava and water (cf. Campbell & Tipper 1971). The weak stratification, suspended clasts and poor sorting and grading suggest that these rocks were deposited en masse (see later section on depositional environment).

Subunit TrJTb. The bedded sandstones and siltstones present within subunit TrJTg can locally be mapped out as a separate if discontinuous subunit, TrJTb. In more detail, these rocks weather buff and pale greenish grey, giving exposures a distinctive colour banding. They are thinly bedded, with beds between 1 mm and 3 or 4 cm (Plate 2F). Grain

size varies from coarse sand to cherty-textured silt. Sandstone beds commonly are graded, and composed of lithic and crystal fragments. Siltstone layers are darker green-grey, very fine grained and finely laminated. Rounded, suspended cherty lenses and blobs in the siltstones could be silicified mud balls (Plate 2F). Rather high energy conditions are suggested by local scouring, truncated current bedding, and soft sediment deformation such as folded laminae and convolutions. Way-up indicators show that the sequence is inverted.

Current bedding and distinct channels are rare or absent in subunit TrJT_s, so these rocks are not considered to be alluvial. They resemble rhythmic layering (turbiditic?), and their association with subunit TrJT_g most likely means that they formed as finer suspended sediment flows, deposited on or between massive mud flows.

Subunit TrJT_c. This subunit is unsorted volcanic conglomerate, usually matrix supported, containing a wide variety of angular to subrounded volcanic and plutonic clasts, some more than 20 cm across. The volcanic fragments are typically plagiophyric, and may have a pilotaxitic groundmass and calcite amygdules. They may include andesitic as well as basaltic rocks. Most of the primary minerals are at least partly chloritized, sericitized or serpentinized. Other clasts include black pyroxenite nodules, and gabbro-diorite.

The matrix of the conglomerate is mid- to dark green and poorly sorted, consisting of sand- or granule-sized altered lithic fragments and abundant plagioclase and ferromagnesian crystal debris, and palagonite and variolitic glass. Impure carbonate is common in the matrix, but it is not clear whether it represents interstitial lime mud or authigenic carbonatization. Some clasts are reddened on the rims due

to oxidation, while other clasts and the matrix generally are not.

Subunit TrJTb. This subunit is green, augite porphyry basalt flow and flow breccia (Plate 2G,H), the latter predominating. It is probably equivalent to "unit 2c" of Bailey (1978). Pillows and flow breccia are well exposed at the summit of Mt. Warren, indicating that at least some of the volcanism was subaqueous. The pillows are not zoned, and interpillow material comprises basalt debris and hyaloclastite. The clasts of the autobrecciated flows are subangular to rounded, from <1 cm to 20 cm across, unzoned, and are usually monomict, though some variation is present locally such as plagiophyric basalt. These features suggest that limited transportation and mixing followed the fragmentation of the lava. The matrix of the breccias consists of fine grained, green basalt or basaltic debris, and locally limestone. The rocks in this subunit are non-foliated and undeformed apart from some flattening of the basalt clasts. The thickness and shape of the depositional units are not known.

The diopsidic augite (Bailey 1978) phenocrysts are euhedral, optically zoned and partly altered; some are 1.5 cm long. Plagioclase phenocrysts are also zoned and partly altered, and also occur as glomerocrysts. Olivine was not found. White to pink amygdules are very common in the volcanics; some are albite, but those in the extreme south are mostly prehnite. The groundmass is very fine grained and contains microlites of plagioclase.

Intrusive rocks. Rocks which are clearly intrusive (subunit TrJT1) are not common in unit TrJT. Among them are a medium- to coarse grained, speckled green-grey, massive diabase-diorite that occurs in a dyke on Spanish Mountain, and a possible ultramafic rock at locality

351, containing green ?pyroxene phenocrysts in a fine grained, pale green foliaceous chloritic groundmass.

Just northeast of Likely is a speckled grey, medium to coarse-grained, massive monzodiorite or monzonite. It comprises mainly plagioclase and amphibole, minor mica and pyrite; no quartz or nepheline was detected by X-ray diffraction of a few mineral separates.

INTERPRETATION OF DEPOSITIONAL ENVIRONMENT OF TAKLA GROUP

The grey thinly bedded rocks (subunit TrJT_a) interbedded with Black Phyllite-type argillites at the base of the Takla Group are interpreted as volcanic tuffs (*sensu lato*). They were probably water-lain pyroclastic ash which was introduced into the ambient hemipelagic carbonaceous mud with increasing frequency up-stratigraphic section as the volcanic sources became more active with time, or became closer, or both. Some were probably transported in submarine channels emanating from ?siliceous or intermediate volcanoes in the Quesnel Belt (Fig. 2.10).

Other lithologies in subunit TrJT_a, higher up in the section, are richer in tuffaceous material in thinly bedded, graded units. These could be moderate- to distal turbidites, or submarine fan deposits. One exposure contains a fragmental rock composed of small (1-2 mm) flattened particles which could be coarse ash or formerly-glass shards; this particular rock could possibly have been a warm submarine pyroclastic ash flow.

There is evidence for only one volcanic centre (subunit TrJT_v) in the lower part of the Takla Group. These pillow lavas and hyaloclastic tuffs represent a small, probably juvenile submarine volcano - there is no extensive debris flow fan in its vicinity to suggest that it was any bigger.

Fig. 2.10

Schematic block diagram showing depositional environments of the various facies of part of the Black Phyllite, and of the Takla Group in the thesis area. Compare Fig. 2.6.

The volcanic wackes of subunit TrJTg are not easy to interpret: their moderately-poor sorting (although they lack clasts larger than a few centimetres), their poor stratification, and the limited abrasion of clasts all have to be reconciled. The depositional environment must be similar to that of the conglomerates because they are associated.

A reasonable model for subunit TrJTg is that these rocks represent submarine sediment gravity flows (cf. Middleton & Hampton 1976). This type of deposit is basically a cohesionless mass with a high sediment concentration in which grains are supported by "dispersive pressure" and moved by grain to grain contact, probably in laminar flow. This mechanism explains the observed general lack of sorting, grading and bedding (the sediment is not 'fluidized' enough to allow density-stratification or settling). Presumably the flow is also slow and gentle enough to preserve glassy fragments and coarse crystals. The main problem is how to explain the absence of large clasts, which in effect distinguishes these rocks from the conglomerates of subunit TrJTc. It is speculated that these sediments were originally sorted in a more turbulent (fluidized) sediment flow, winnowing out the larger clasts; similarly they could have formed the suspension cloud associated with a submarine debris flow. Thus, it was probably formerly bedded and graded, and subsequently remobilized as a grain flow, described above, obliterating the original primary structures.

The graded bedded volcanic sandstones and siltstones of subunit TrJTb probably represent turbidite flows of tuffaceous sediment in suspension at the peripheries of debris- and grain flows.

The matrix-rich conglomerates of subunit TrJTc are interpreted as deposits of submarine debris flows. The conglomerate contains a wide

variety of clast composition, size and shape, implying derivation from numerous volcanic and rare (exhumed) plutonic sources. Limestone in the matrix, and post-depositional oxidation of some clasts indicate that the water depth was not great, certainly above the CCD. The debris flows may have been channelled in submarine canyons, incised in a broad shelf or apron of volcanic and volcanoclastic rocks (subunit TrJTg, stratigraphically below the conglomerate) surrounding subaqueous and probably some subaerial volcanic centres. Lavas, flow breccia, and pyroclastic and hyaloclastic conglomerates and wackes (crystals, glass and rock fragments) may have accumulated by rapid erosion on volcanic flanks at the heads of these channels or smaller tributaries (on land or under water) before being re-mobilized in slumps or liquefied mass-flows, triggered by gravitational instability, earthquakes, faulting or violent eruptions. No vesicular, pumiceous or 'bomb-like' clasts were found, so the volcanism is thought to have been not very explosive, nor silicic. The crystal- and lithic-rich matrix mainly represents pyroclastic and hyaloclastic debris and products of attrition. There was probably much erosion, mixing and re-sedimentation.

Some conglomerate is clast-supported and could represent a submarine talus fan. One specimen contains little-travelled vitrophyric rock fragments and palagonite matrix, possibly derived from hyaloclastic debris.

A schematic sketch diagram (Fig. 2.10) attempts to integrate the above depositional models of the various volcanoclastic lithologies of unit TrJT. This kind of depositional model is very similar to those described by Fisher (1982), Grippi & Burke (1980) and Cas et al. (1981). The overall environment is interpreted to be a volcanic island arc made up of numerous coalesced volcanic centres, linked by extensive subaerial

but mainly submarine volcanoclastic aprons, incised by channels. These deposits conformably overlapped the hemipelagic mud and silt of the Black Phyllite.

CHEMICAL COMPOSITION OF THE TAKLA GROUP

No chemical analysis of Takla Group rocks was done by the author but a considerable amount has been done by Morton (1976) and Bailey (1978) in areas immediately adjacent to this thesis (Chapter 1.2), and it is pertinent to summarize their results and conclusions.

Takla Group volcanic rocks in both the Horsefly (Morton 1976) and Morehead Lake (Bailey 1978) areas are characterized by an alkalic suite of basalts to trachytes. The basalts are mostly clinopyroxene-phyric porphyries, and most are nepheline normative.

Average whole-rock chemical analyses of basalts (excluding trachytes, trachybasalts and analcime-bearing basalts) were calculated by the author from the data of Morton and Bailey and are shown in Table 2-6. One of Bailey's samples is actually from Mt. Warren (Map 1) in the author's area (subunit TrJTb; Table 2-6, column 3); like most of the rest it is alkali olivine basalt.

The interpretation of the chemical composition of unit TrJT involves regional considerations and comparisons, so a full discussion of this matter is left until Chapter 6.1.

[2] 6. UNNAMED JURASSIC ROCKS (UNIT 1-J)

Some distinctive sedimentary rocks occur sporadically around Likely.

TABLE 2-6: . CHEMICAL ANALYSES OF BASALTS FROM THE QUESNEL LAKE AREA

	1	2	3
SiO ₂	48.06	49.40	48.7
TiO ₂	0.98	0.76	0.88
Al ₂ O ₃	13.47	14.30	13.6
Fe ₂ O ₃	2.74	2.66	2.68
FeO	8.06	7.54	7.67
MgO	8.41	6.49	6.35
CaO	10.18	9.20	11.5
Na ₂ O	3.01	3.46	2.85
K ₂ O	2.04	2.48	1.93
P ₂ O ₅	0.34	0.44	0.37
H ₂ O	2.44	loss 2.94	loss 3.78
Total	99.73	99.67	100.55
K ₂ O/Na ₂ O	0.68	0.72	0.68
Sr	995	-	-
Zr	59	-	-
Y	21	-	-
Cr	181	-	-

Key to chemical analyses:

1. Average of 11 alkali olivine basalts, alkali basalts, analcime basalt, nepheline basalt, Horsefly Group (XRF method; Morton 1976).
2. Average of 35 alkali olivine basalts, alkali basalts, Morehead Lake (XRF method; Bailey 1978).
3. Alkali olivine basalt, Mt. Warren, in this area (from Bailey 1978).

They are tentatively distinguished from the Takla Group (unit TrJT) for several reasons. First, they are not metamorphosed, and apart from fracturing and local, steep tilting, they are not significantly deformed. Second, they are apparently younger than the Takla Group. Third, the lithologies imply a different environment of deposition from the Takla Group.

AGE AND CORRELATION OF UNIT 1mJ

The rocks in this unit are the same as those mapped by Bailey (1978), who designated them "unit 6". The sedimentary rocks are Pliensbachian to early Bajocian (late Early Jurassic to early Middle Jurassic), based on ammonites collected by R.B. Campbell (see Bailey 1978, p. 64), and so are younger than at least the main part of the Takla Group. The plant remains in subunit 1mJg (see below and Bailey 1978) are apparently not diagnostic of stratigraphic age.

It is suggested that this unit could be correlated with the Sinemurian to Callovian (upper Lower to upper Middle Jurassic) Ashcroft Formation (Fig. 2.3; Price et al. 1985), if its distinction from the Takla Group is indeed warranted. If it is not warranted, in spite of the arguments offered below, then the Takla Group range would have to be extended to the Middle Jurassic.

LITHOLOGY OF UNIT 1mJ

Subunit 1mJg. This subunit (equivalent to units 6C & D of Bailey 1978) includes volcanoclastic wacke, breccia and conglomerate, and

carbonaceous mudstone. The wackes comprise volcanic rock fragments and are rusty-weathering, green to brown to maroon, and fine to medium grained. They are fairly well sorted and locally bedded or laminated. The rocks are fractured and partly altered, but show no trace of metamorphic recrystallization.

The lithic wackes are closely associated with matrix supported conglomerate and breccia, comprising rounded to mainly angular, pebble-sized clasts of black, very fine grained volcanic rock, green porphyritic lava, limestone and siltstone. The matrix is unsorted volcanic detritus.

Dark blue-grey to black, very fine grained carbonaceous mudstone is interbedded with the lithic wackes and conglomerates. It is thinly bedded with a fissility parallel to bedding producing a friable, flaggy, parting. It may be slightly calcareous. Locally there is abundant plant stem and leaf debris. According to Bailey (1978) they are monocotyledons; they suggest a non-marine environment.

Subunit 1mJl. This subunit (equivalent to unit 6A of Bailey 1978) is impure, possibly bituminous limestone, and calcareous and carbonaceous mudstone and sandstone. The limestone is pale to dark grey, fine grained to crystalline, and partly clastic. It is generally uniform and poorly bedded, although there are interbeds of sandstone and black shale. No macrofossils were found.

The calcareous mudstone is dark grey to black, with thin flasered bedding. The dark grey sandstone contains volcanic grains and is graded locally. A rhythmic sequence was noted in one exposure, comprising sandstone, limestone and shale, probably younging in that order.

Subunit 1mJc. This distinctive subunit (equivalent to unit 6B of

Bailey 1978) comprises multicoloured, moderately sorted chert- and volcanic-pebble polymictic conglomerate (Plate 2I). It is coeval with subunit 1mJg. The well-rounded clasts are pebbles and cobbles up to 20 cm across, and include blue-black, grey or green chert, limestone, siliceous volcanic rock, and porphyritic and amygdaloidal-vesicular volcanic rocks. Possible plutonic clasts, and white and grey quartz clasts were noted rarely. The conglomerate is mainly clast-supported and there is some weak primary clast imbrication. The matrix is composed of moderately sorted, coarse-sand sized multicoloured lithic fragments, which are angular to rounded. The clast-matrix ratio varies. Overall, bedding is obscure. There is minor fracturing, but no fabric or significant flattening.

Subunit 1mJv. This is a poorly exposed, pale to dark green, fine to medium grained, equigranular volcanic rock, probably basaltic in composition.

INTERPRETATION OF DEPOSITIONAL ENVIRONMENT OF UNIT 1mJ

Non-marine conditions are suggested by the fossil flora in subunit 1mJg. Furthermore, the well-rounded pebbles and variety of clast types in subunit 1mJc imply much fluvial transportation and mixing. The close association of these two subunits suggests, perhaps, an alluvial fan draining into a lake or peri-marine lagoon, bordered by vegetation (cf. Bailey 1978), or possibly a fairly high energy river system with interfluvial vegetated mudbanks, marshes or lagoons. (The possibly correlative Ashcroft Formation is also partly non-marine: Price et al. 1985.)

The limestone of subunit 1mJ1 is not conclusively coeval with these other subunits. If it is, it could be a freshwater limestone, or brackish, shallow water limestone in a backshore environment.

Ammonites were exclusively marine. If the reported ammonites (Erycites sp., Tmetoceras sp. and Utonia sp.) are representative of the environment described above, proximity to at least a shallow sea is implied. Marine conditions are not conclusive, however, because dead ammonites could be washed into tidal or backshore (terrestrial) areas by storm waves and currents. So even with the apparent presence of ammonites, overall a terrestrial or marginal marine (lagoonal?) environment is suggested by the lithologies of unit 1mJ.

Porphyritic basalt of the Takla Group (unit TrJTb, probably Norian) occurs sporadically as inliers among rocks of unit 1mJ. The contact between them was not observed, but is presumed to be an unconformity because of the age difference and the lithological contrast. The amount of erosion, if any, represented by the hiatus is not known, but some erosion of the Takla Group might account for the apparent difference in metamorphic grade between the unmetamorphosed unit 1mJ and the weakly metamorphosed (zeolite or prehnite-pumpellyite facies?) Takla Group. This is very speculative, as it is difficult to quantify.

The origin of the abundant chert in the conglomerate (subunit 1mJc) is interesting. There are two main possible sources. An eastern provenance, from the cherts of the Antler Formation or the Black Stuart Group, is unlikely because the conglomerate lacks other rock types which would be expected from the geological terrane to the east, such as sandstone, grit or gneiss clasts. A far more likely provenance is the

cherts and volcanics of the Cache Creek Group of the Cache Creek terrane to the west (Fig. 2.3), as suggested by Bailey (1978). This implies an east-facing paleoslope, and would confirm that the Cache Creek and Quesnel terranes were contiguous at this time, if they weren't always.

[2] 7. QUATERNARY VOLCANIC ROCKS (UNIT QV)

Basaltic volcanic and volcanoclastic rocks of unit QV occur in the southeast of the thesis area. They are part of a suite of late Tertiary and Quaternary alkali olivine basalts in the Quesnel Lake region (Fiesinger & Nicholls 1977) known as the Anahim Volcanic Belt (Souther 1977).

Fiesinger (1985) concluded that the lava flows immediately west of the North Arm of Quesnel Lake were contemporaneous with Recent alpine glaciation. She suggested that the conical landform with a central depression in the east of the thesis area, at the northern end of the QV outcrop, is an extinct volcano (or cinder cone: Campbell 1978).

LITHOLOGY OF UNIT QV

Subunit QVb. This is the most common lithology, a dark gray to black basalt with olivine phenocrysts (1-2 mm) in a very fine grained groundmass, and locally nodules of dunite or peridotite. The basalt is slightly vesicular to highly vesicular, scoriaceous and cindery. A subhorizontal primary layering is discernible.

Subunit QVc. This is an unsorted, matrix supported polymictic conglomerate, probably a debris flow derived from basaltic tephra,

volcanics, and pre-Quaternary bedrock. The clasts include vesicular basalt, and angular pebbles of orthogneiss and paragneiss. The matrix is pale to mid-grey, completely unsorted, and composed of rock and mineral detritus.

[2] 8. SUMMARY OF STRATIGRAPHY

SNOWSHOE GROUP (UNIT HPS)

Occurring west of the Pleasant Valley thrust, the Snowshoe Group consists of Upper Proterozoic and Paleozoic metasedimentary rocks possibly equivalent to the Kaza Group, Black Stuart Group and Cariboo Group, which occur east of the Pleasant Valley thrust (Fig. 1.2; Struik 1983a). They may be partly correlative with the Paleozoic Eagle Bay Formation and Milford Group, which occur farther south in the Omineca Belt (Fig. 2.3, Table 2-2).

The relationship of the Snowshoe Group to the paleo-continental margin is not clear, and these rocks have been assigned to the 'suspect' Barkerville terrane by Struik (1985a,b). The present author tentatively interprets the mainly moderately- to thinly interbedded pelitic and 'gritty' psammitic metasediments as the outer fan apron of a submarine fan complex, which corresponds to the traditional 'distal turbidites'.

QUESNEL LAKE GNEISS (UNIT DMWQ and DMEQ)

Two varieties of gneiss, both intrusive into the Snowshoe Group, occur in the thesis area, herein named the Western- and Eastern Quesnel

Lake Gneiss (Figs. 2.1, 2.4). The Western Quesnel Lake Gneiss (DMWQg) forms a sheet-like body about 400 m thick, in the north and northwest of the thesis area. It is a well foliated and lineated to mylonitic, megacrystic (mainly perthitic microcline) granite orthogneiss.

The Eastern Quesnel Lake Gneiss (DMEQg) in the southeast of the thesis area, is a non-porphyritic, homogeneous ?granodioritic orthogneiss which is well foliated and lineated.

Although not well defined, the emplacement age of the gneiss is interpreted to be between 375 and 350 Ma (Mortensen et al., in press), or Late Devonian to early Mississippian (Table 2-3, Fig. 2.5). The radiometric data also records isotopic disturbances: Middle Jurassic Rb-Sr and U-Pb dates have been interpreted as due to resetting during regional metamorphism (Mortensen et al., in press), and Rb-Sr and K-Ar mineral ages around 100 to 110 Ma probably represent resetting due to mid-Cretaceous plutonic activity in the region (see Okulitch in Stevens et al. 1982, p. 13).

ANTLER FORMATION (UNIT PPA)

This unit (typically about 250 m thick) forms the base of the Intermontane Belt (or superterrane) in the thesis area and in the Quesnel Lake region (Figs. 1.2, 2.1, 2.2). The main rock types are greenstone (or greenschist), metagabbro, serpentinite and steatite. Chemical analysis indicates that the greenstones represent subalkalic tholeiitic metabasalt formed in an ocean floor setting.

Unit PPA is herein correlated with the type-Antler Formation in the Cariboo Mountains, with the Fennell Formation to the southeast, and with the Kaslo Group of the Kootenay Arc (Fig. 2.3, Table 2-2). Together

they constitute part of the Slide Mountain terrane (formerly the Eastern Assemblage and the Eastern terrane).

BLACK PHYLLITE (UNIT uTrBP)

The Upper Triassic (Karnian to early Norian) Black Phyllite overlies the Antler Formation (Figs. 1.2, 2.1, 2.2, 2.6); it is probably a sheared depositional contact. It is predominantly composed of black, well cleaved carbonaceous argillite and siltstone, with minor volcanic conglomerate and mafic volcaniclastic schist. Its present thickness is approximately 2500 m. It is probably correlative with the Slocan Group of southern British Columbia (Table 2-2, Fig. 2.3).

The Black Phyllite was laid down in a very low energy environment, probably a stagnant euxinic basin with localized volcanic activity. It is inferred that the basin was barred to the west by the incipient (volcanic) Quesnel Belt (see below), from which at least some of the sediment was derived.

TAKLA GROUP (UNIT TrJT)

This unit forms part of a belt of Upper Triassic to Lower Jurassic volcanic, sedimentary and plutonic rocks, herein called the Quesnel Belt, that forms the superstructure of the Quesnel terrane. About 9000 m of the unit is present in the thesis area.

Unit TrJT gradationally overlies the Black Phyllite and comprises tuffaceous argillites and siltstones, followed by volcaniclastic wackes, volcanic conglomerates (debris flows) and volcanic lavas (alkalic

olivine basalt: Table 2-6) and flow breccias (Fig. 2.6).

The inferred environment of deposition is summarized schematically in Fig. 2.10, which shows a volcanic island arc composed of coalescing volcanic centres, linked by extensive subaerial but mainly submarine volcanoclastic aprons. These deposits conformably overlapped the hemipelagic mud and silt of the Black Phyllite.

UNNAMED JURASSIC ROCKS (UNIT 1mJ)

These Pliensbachian to early Bajocian rocks are considered distinct from the Takla Group by virtue of their age, fossils and lithological characteristics. There are no obvious correlatives of this unit in the Quesnel terrane, except possibly the Ashcroft Formation (Fig. 2.3).

The unit consists of volcanic wacke, breccia, conglomerate and mudstone (containing plant stem and leaf debris), limestone, and chert-and volcanic pebble polymictic conglomerate, the latter probably derived from the Cache Creek Group. Non-marine environmental conditions are suggested by the fossil flora, or possibly a peri-marine lagoon.

QUATERNARY VOLCANIC ROCKS (UNIT QV)

Quaternary basaltic and volcanoclastic rocks in the southeast of the thesis area are part of a suite of late Tertiary and Quaternary alkali olivine basalts in the Quesnel Lake region known as the Anahim Volcanic Belt. They include black, olivine-phyric, vesicular basalt, and polymictic conglomerate, interpreted as a debris flow deposit.

CHAPTER THREE

DEFORMATION AND METAMORPHIC RECRYSTALLIZATION

[3] 1. INTRODUCTION

The deformation in the thesis area is dominated by two major, consecutive phases. The first phase (D_1) was the tectonic emplacement of the Quesnel terrane allochthon (including the Antler Formation at its base) onto the Snowshoe Group. This produced a foliation (S_1) and locally a stretching lineation (L_1) in both plates; the fabrics are mylonitic at the terrane contact, which is named the Quesnel Lake Shear Zone.

The second major phase (D_2) deformed the now coupled Quesnel terrane and Snowshoe Group into large-scale, syn-(peak) metamorphic, southwesterly-verging sub-recumbent folds, which plunge to the northwest.

This chapter concentrates on these two phases. There are few mesoscopic and map-scale structural features of D_1 because it is largely overprinted by D_2 , so the section that deals with D_1 (Section 2) concentrates on the foliation and microstructure of S_1 in the Quesnel

Lake Shear Zone, where it is best preserved, and on the available kinematic indicators that demonstrate the sense of shear. These are described in detail, as it is important for the tectonic reconstructions to document as many of the parameters of D_1 as possible, such as displacement paths and metamorphic conditions.

The second phase of deformation is important because it is responsible for the main mesoscopic structures (Section 3) and the map pattern (Section 6), and because its development and demise is intimately related to the metamorphic history in the area. For this latter reason, a separate section (Section 4) focuses on the metamorphic textures and their relationship to the D_2 foliation (S_2) and lineation (L_2), a relationship that varies across the thesis area and which has interesting implications. These matters are raised again in the chapter on metamorphic conditions (Chapter 4) and are discussed fully in Chapter 6.

A third phase of deformation (D_3) comprises post-(peak) metamorphic folding and crenulations, and is described in Section 5.

Note on timing of deformations

This is discussed in Chapter 6.2, but for convenience the inferred timing of deformations is introduced here. The maximum age of first phase deformation is probably late Early Jurassic, from the age of the youngest marine sediments in the upper plate of the Quesnel Lake Shear Zone. Second phase deformation followed soon after in the Middle Jurassic, on the bases of isotopic resetting due to the accompanying metamorphism, and dating of syn-kinematic intrusions. D_2 probably continued into D_3 without significant hiatus, in the Late Jurassic.

[3] 2. FIRST PHASE DEFORMATION: D₁

QUESNEL LAKE SHEAR ZONE

Recognition and definition of the Quesnel Lake Shear Zone

The strain in metasediments of the Snowshoe Group and in the Western Quesnel Lake Gneiss is very strong at and towards their respective contacts with the Antler Formation at the base of the Quesnel terrane. In the Western Quesnel Lake Gneiss, grain size decreases and a finely-spaced compositional banding, foliation, and stretching lineation are increasingly developed (Plates 3A, 4A). In the Snowshoe Group, quartzites and psammites also show grain size reduction, strong flattening, preferred crystallographic orientations in quartz microfabrics, and highly transposed compositional layering (Plate 3B). Strong flattening and transposition of compositional layering was also recorded locally in Antler Formation greenstones (Plate 3C,D).

These features of high strain are interpreted to have been produced by dynamic (syn-kinematic) recrystallization in a ductile shear zone generated by relative movement between the Quesnel terrane and the Omineca Belt. That this is a major tectonic contact (Rees 1981), herein called the Quesnel Lake Shear Zone (Brown et al. 1986), is one of the principal findings of this thesis. In this it agrees with Montgomery's (1978) conclusion farther south, and resolves the ambivalence of Campbell's (1971) interpretations of this contact in the Crooked Lake area (see Chapter 1.2).

The shear zone was called the Quesnel Lake Fault Zone by Brown & Rees (1981). Its northwestern extension, between the thesis area and the Naver pluton (Fig. 2.3) has been called the Eureka thrust by Struik (1985a, 1986b; Fig. 1.2)

The Quesnel Lake Shear Zone is shown on Map 1 as the terrane boundary, i.e. the base of the Antler Formation, although actually it is a zone of high strain of variable thickness in the order of tens of metres extending above and below the terrane contact. This high strain is most evident in the fabrics and mylonitic microstructures in quartzofeldspathic lithologies at the top of the lower plate, as described above and below. The high strain is not obvious in pelitic and greenschist lithologies in the shear zone because these did not deform dominantly by mylonitization, although they are well foliated and sheared.

High strain may not be confined to the terrane contact: such zones may occur many metres or kilometres away from the contact, as reported by Phillipone (1985), although the present author would urge caution against confusion with locally strong D_2 grain size reduction and stretching lineations (see Section 3) - a misidentification he himself made originally.

The reason that the Quesnel Lake Shear Zone is D_1 .

The Quesnel Lake Shear Zone is folded by large-scale, southwesterly-verging folds of a later generation (D_2 : described in Section 3); minor folds related to these structures, in both the upper and lower plates of the shear zone, fold only one tectonic fabric. Therefore, the Quesnel Lake Shear Zone, and related pre- D_2 fabrics (S_1 and L_1), are interpreted to be the principal expressions of the oldest deformation (D_1) in the thesis area. To emphasize the point, no evidence was found for a deformation in either plate that pre-dated the Quesnel Lake Shear Zone.

FIRST PHASE STRUCTURES AND FABRICS

General points.

Very little evidence for F_1 folding was found in the thesis area. Two F_1 structures are suspected, but they are relatively small and their geometry and facing could not be worked out because of lack of exposure and D_2 and D_3 overprinting; they will not be elaborated on. Otherwise, the preserved exponents of first phase deformation are limited to very rare intrafolial folds, a foliation (S_1), a stretching lineation (L_1), and microfabrics interpreted to have formed during D_1 .

First phase deformation can be inferred throughout the lower plate of the shear zone from the presence of S_1 , and is present in the lower parts of the upper plate, diminishing in the higher, cooler structural levels in competent volcanic and volcanoclastic lithologies of the Takla Group. Bedding (S_0) in metasedimentary rocks in both plates was transposed into S_1 to become S_{0-1} , except in rare cases where an angle between S_1 and S_0 is preserved (see below).

The first phase stretching lineation (L_1) is rarely preserved outside quartzofeldspathic lithologies in the Quesnel Lake Shear Zone (except in the Western Quesnel Lake Gneiss, where it is virtually always preserved: Fig. 3.1), being overprinted by later deformations and recrystallization. Similarly, S_{0-1} has been transposed into the foliation (S_2) of second phase folds (F_2), and has been recrystallized by the M_2 metamorphism. This overprinting means that in general D_1 fabrics can be distinguished only:

* In the hinge zones of F_2 minor folds, where S_2 is spaced and crenulates S_{0-1} , and transposition is less well developed.

Fig. 3.1

Equal-area stereogram of L_1 stretching lineations, measured from megacryst alignment and ribbon grains in Western Quesnel Lake Gneiss. All structural domains.

All those lineations trending between 008° and 052° in the northeast come from the gneiss in the Badger Peak area, on the overturned limb of the Seller Creek syncline; furthermore, no L_1 lineations in this sub-domain plot anywhere else. Thus it is a distinct domain in terms of L_1 orientation.

In contrast, L_1 lineations from the outcrops of gneiss east and west of Cariboo River generally trend west-northwesterly and constitute the rest of the data in the stereogram.

* In competent lithologies in the Quesnel Lake Shear Zone because there the conspicuously strong D_1 fabric and microfabric are discernible through the comparatively 'mild' effects of D_2 flattening and recrystallization.

The occurrence of non-parallel S_0 and S_1 :

At about 10 localities, S_1 lies at an angle to S_0 . In all cases, S_2 is also distinguishable, and crenulates S_1 ; S_3 is usually also distinguishable.

Usually the asymmetry of S_1 with respect to S_0 is vague or ambiguous but in a few localities (e.g. 23, 238, 429, 492) it is relatively clear: after the effects of D_2 folding are removed, the apparent dip of S_1 always dips southwestwards more steeply than S_0 . If the flattening and sense of overturning indicated here represents upward-facing D_1 kinematics, then northeastward vergence is implied. (In these cases, S_2 has the opposite vergence - otherwise S_1 would not be distinguishable.) Northeastward vergence for S_1 is consistent with the sense of shear determined from D_1 strain in the Western Quesnel Lake Gneiss (see below).

* * *

First phase deformation is described in more detail in the next two sections by reference to mesoscopic structures and microstructures in the Western Quesnel Lake Gneiss and in quartzofeldspathic metasediments of the Snowshoe Group in the Quesnel Lake Shear Zone, where they are comparatively well preserved, so that D_1 kinematic and metamorphic conditions can be ascertained.

FIRST PHASE DEFORMATION IN THE WESTERN QUESNEL LAKE GNEISS

Introduction

The Western Quesnel Lake Gneiss is moderately to highly deformed, manifested by much grain size refinement and the development of a very strong foliation and lineation (Plate 4A). The strain is described in a general way as mylonitic because the grain size has been reduced by dynamic recrystallization, described below. This strain is inferred to have occurred during D_1 for the following reasons. First, rare folds and crenulations in the Western Quesnel Lake Gneiss, which are most likely F_2 , fold an already highly strained and recrystallized fabric, including polygonized quartz ribbons and those micaceous layers derived from the mechanical and metamorphic breakdown of feldspars (see Chapter 2.2). Second, no comparable microstructures have been recognized conclusively to have formed due to second or third phase deformation. Certainly there may have been some shearing on the limbs of second phase folds, but evidently strain rates were not high enough to produce mylonitic microstructures, and there are no post- D_1 'shear zones'.

It is emphasized that even in the Western Quesnel Lake Gneiss, the 'mylonitic' S_1 is overprinted to some extent by D_2 flattening (cf. cross-sections A to F), forming a crenulation and crenulation foliation (S_2) visible very locally. In general, therefore, the fabric of the Western Quesnel Lake Gneiss is strictly S_{1-2} .

Kinematic study

At locality 16 (UTM 0606158415) there is a large, clean exposure of Western Quesnel Lake Gneiss in which the potassium feldspar megacrysts are systematically but not symmetrically oriented with respect to the

mylonitic foliation. Because the protolith was a porphyritic granite (Chapter 2.2), it is assumed that the megacrysts were originally randomly oriented, and have been re-oriented by D_1 . A study was made here to ascertain the kinematic history of the deformation, the sense of shear, and the deformation mechanisms. A summary of this study was given by Rees & Ferri (1983). It is realized that some D_2 flattening has contributed to the megacryst orientation and finite strain, but the rarity of D_2 minor structures implies that internally D_2 strain in the gneiss was minor and not great enough to invalidate the main interpretations that follow.

The mylonitic foliation in the gneiss, assumed to be parallel to the plane of shear, is subhorizontal (061/10 NW). It contains a strong stretching lineation (10/277), marked by megacrysts and quartz ribbon grains, assumed to approximate the X axis of the finite strain ellipsoid (Fig. 3.2a; Plate 1G). The rock is cut by two roughly orthogonal, vertical joint sets which are respectively parallel and perpendicular to the stretching lineation. Thus, the set of joint faces parallel to the lineation displays the strain corresponding to the XZ plane of the strain ellipsoid; the other set of joint faces display approximately the strain in the YZ plane (Fig. 3.2a). Seven separate XZ surfaces, each of about $(50 \text{ cm})^2$, were examined (e.g. Plate 3E). In each, measurements were made of the long axis (a) and short axis (b) of intact megacrysts considered to have behaved as rigid inclusions, and also the angle (θ) between the long axis and the upward normal to the mylonitic foliation. Smaller or strongly comminuted feldspars were not measured as in effect they behaved as part of the 'matrix'. A total of 504 megacrysts were measured from the seven XZ surfaces; there was no significant variation

Fig. 3.2

(a) Block diagram showing exposed surfaces in the Western Quesnel Lake Gneiss at locality 16 in terms of strain geometry. The top surface is the plane of shear in the gneiss and is close to the XY finite strain plane; the plane perpendicular to that and to the stretching lineation is approximately the YZ plane.

(b) Rose diagram illustrating orientation distribution of megacryst long axes in XZ plane, looking north. Degree notation refers to θ , as in Fig. 3.3.

in the data between the surfaces. The axial ratio R , where $R = a/b$, is plotted against ϕ in Fig. 3.3. In this diagram, $\phi = 0^\circ$ corresponds approximately to a vertical line or the finite Z axis, and $\phi = 90^\circ$ corresponds to a subhorizontal line trending approximately east, parallel to the stretching lineation, approximately in the X direction.

Two important observations are made. First, the angle ϕ generally increases as R increases. The significance of this systematic relationship is discussed in Appendix B. Second, the majority of the megacrysts have their long axes lying approximately parallel to the stretching direction in the shear plane, or dipping to the west, in the XZ plane (Fig. 3.2b; Plate 3F). The remainder have their long axes dipping to the east. The west/east ratio is about 3 to 1. If the long axes are thought of as representing small rigid material marker lines which have suffered rotation in progressive deformation, the orientation bias indicates a non-coaxial strain path, with a clockwise (looking north) rotation. This implies a component of upper plate to the east simple shear, with the mylonitic foliation being parallel to the plane of shear (as has been assumed) and the stretching lineation in this foliation being parallel to the displacement vector.

Alternatively, if the strain had been produced by pure shear flattening alone, the megacryst orientations would have been symmetrical about the mylonitic foliation, assuming that they were originally randomly oriented. The observed asymmetry indicates that pure shear was not solely operative; it does not mean that simple shear and pure shear are mutually exclusive models for the deformation, as is discussed in Appendix B.

Other kinematic indicators in Locality 16. In at least two separate

Fig. 3.3

R versus ϕ diagram (axial ratio versus finite angle of rotation) for megacryst distribution in Western Quesnel Lake Gneiss at locality 16. $\phi = 0^\circ$ is approximately vertical, up, and $\phi = 90^\circ$ is subhorizontal and trending east (looking north). $\phi = 90^\circ$ is parallel to the plane of shear in the gneiss. Note the higher its axial ratio, the closer is the megacryst to the plane of shear. Conversely, the lower the axial ratio, the more random the orientation. For explanation of parabolic curves, see Appendix B.

places in the same exposure, the mylonitic foliation is not subhorizontal, but dips about 20° to the west. These zones, about 50 to 100 cm thick, are bounded by gneiss with the typical subhorizontal mylonitic foliation, and are separated from it by discrete zones of shear (Plate 3G). They represent a discontinuity or heterogeneity in the bulk shear strain in which the rock has suffered less flattening (represented by the foliation and the long axes of the megacrysts), and less rotation of the flattening foliation towards the shear plane. That the foliation dips west implies rotation to the east, compatible with an upper plate to the east sense of shear. This is exactly analogous to the method of Berthe et al. (1979) who used the asymmetry of S surfaces (flattening foliation) and C surfaces (planes of shear) in orthogneiss to infer the sense of shear in the South Armorican Shear Zone in Brittany, France.

Locally, shear band foliation (White et al. 1980, Platt 1984, Harris & Cobbold 1985) is present in the gneiss. It dips eastwards at about 20° to the mylonitic foliation (Plate 3F,H). The shear bands generally propagate from the corners of feldspar porphyroclasts and extend for a few centimetres before attenuating into the shear plane. These foliations definitely do not crenulate the mylonitic foliation in a compressional way, indicating that they are syn-kinematic with D_1 and do not represent S_2 .

The sense of asymmetry of the shear bands implies an upper plate to the east sense of shear, consistent with the other kinematic indicators. However, some authorities (e.g. Platt 1984) have noted that most shear bands, or "extensional crenulation cleavage", are likely due to a partitioning of the bulk strain into domains of coaxial flow which can produce conjugate sets of 'normal-sense' shears whose geometry has no simple relation to the bulk simple shear flow. The point is taken;

however, only one set of shear bands is present in the gneiss, and their displacements are most compatible with the inferred sense of shear.

Kinematic indicators from elsewhere. Kinematic indicators were also observed in the Western Quesnel Lake Gneiss around locality 715, close to the terrane contact on the overturned limb of a major D_2 syncline (Seller Creek syncline). Again the mylonitic foliation, defined by micas, quartz ribbons and attenuated feldspars, is gently dipping to the northeast and there is an excellent stretching lineation defined by feldspar megacrysts and quartz ribbons which trends consistently about 025° (Fig. 3.1). In XZ sections, feldspar megacrysts and micas in the flattening plane dip southwestwards with respect to the very strong mylonitic foliation, which is assumed to mark the plane of shear (Plate 3I); this C and S fabric asymmetry indicates an upper plate to the northeast sense of shear. (It is not confused with S_{1-2} relationships because the latter would exhibit crenulation, which is absent.)

Ferri (1982) studied the microfabric in quartz ribbons in the XZ kinematic plane in the Western Quesnel Lake Gneiss at locality 700, with a universal stage. The resulting fabric (Fig. 3.4) shows pole free areas, and girdles at a high angle to the mylonitic foliation. Ferri recognized a slight asymmetry of the maxima, from which he inferred an upper plate to the northeast sense of shear. Although this locality (and 715) is on the overturned limb of a major F_2 fold, unfolding the structure should conserve the displacement vector as well as the shear sense because the stretching lineation is practically perpendicular to the fold axis (compare Fig. 3.1 with F_2 fold axes in Fig. 3.6). There are ~~no~~ serious restrictions of the petrofabric analysis, however, such as (see Ferri 1982): the origin of asymmetric c-axis fabrics, which is

Fig. 3.4

C-axis fabric stereogram (equal area) of quartz ribbons in the Western Quesnel Lake Gneiss at locality 700 (from Ferri 1982). Contours are 0.5, 1, 2, 3 and 4% per 1% area. The mylonitic foliation, parallel to the shear plane, is represented by the XY plane (horizontal line) in the stereogram. The girdle is asymmetrically centred about the normal to the shear plane (subvertical), from which Ferri (1982) inferred an upper plate to the northeast sense of shear (see text). However, this interpretation is suspect because of the flattening imposed by D₂ folding, such that the petrofabric is really a composite S₁₋₂ fabric.

controversial; the reliability that can be placed in a single result; and the effects of D_2 and D_3 overprinting on the quartz fabric.

Microstructure

The D_1 metamorphic grade (M_1) may be ascertained from the characteristics of the quartz and feldspar microstructure in the Western Quesnel Lake Gneiss. Furthermore, it will be shown later that the D_2 metamorphic grade increases from northwest to southeast in the thesis area, so another question arises: was there a similar or different metamorphic gradient during D_1 ? To help answer these questions, the microstructure in the gneiss is described with reference to two localities: one (16) is in the northwest and about 3000 m structurally beneath the Quesnel Lake Shear Zone (i.e. removing D_2 folding), and the other (700) is farther east and right in the shear zone, very close to the terrane contact.

Locality 16. Potassium- and plagioclase feldspar deformed mainly by brittle fracturing and disintegration, mechanical twinning, and kinking, but with some development of subgrains and new grains by dynamic recovery and recrystallization. Associated with this strain was the retrograde alteration of the feldspar to muscovite and sericite (Plate 4B), which acted as a strain-softening mechanism for the rock as a whole (cf. 'reaction-softening' of White et al. 1980).

Feldspar megacrysts (perthitic microcline) are larger and less disintegrated in this gneiss compared to that at locality 700. Many feldspars are intact; others have extension fractures ('pull-aparts') filled with polycrystalline quartz (Plate 4B). Undulose extinction is generally present, as is some degree of mechanical twinning, especially

at stress points at the margins of megacrysts or where they impinge on other porphyroclasts. Chessboard albite is a common feature of this. The twin planes themselves may be wavy due to a plastic bending, although conjugate kink bands in some megacrysts attest to their general rigidity. Pressure shadows around the megacrysts contain quartz and albitic plagioclase (Plate 4C); some of the latter was possibly derived from the adjacent perthitic microcline during the alteration which produced the white mica, or during strain-enhanced diffusion. Some megacrysts contain large, irregular subgrains, but generally features of dynamic recrystallization and non-brittle grain size reduction are not well developed. The lattice structure of feldspar impedes the motion of dislocations (White et al. 1980), and the crystals strain-harden rapidly at low temperatures, leading to brittle fracturing (Plate 4B). The most disintegrated and strung-out feldspars were probably the first to fracture, and are the most altered to white mica.

Although variably recrystallized and re-deformed during D_2 , quartz clearly sustained much dynamic recrystallization during D_1 , forming irregular, polycrystalline lenses and attenuated layers with high aspect ratios, bounded by foliae of white mica or feldspar porphyroclasts. There are larger porphyroclasts of quartz, about 5 mm across, which are probably relict primary crystals. These are invariably strained to some degree, the microstructures in a single porphyroclast ranging from undulose extinction to deformation bands, subgrains and new grains (Plate 4D,E). The shape fabric of the quartz lenses is interpreted to be D_1 , but some of the undulose extinction and recovery is probably younger (D_2 or even D_3). The polycrystalline aggregates of new grains formed mainly by subgrain rotation rather than by grain boundary migration, and there was apparently not much grain growth. This is

inferred because there is generally a subtle transition from a subgrain-filled porphyroclast to an aggregate of new grains, and the new grains are inequigranular, irregularly-shaped with few triple-junctions, and no serrated grain boundaries. All this is evidence that these are 'low-temperature mylonites' (sensu lato), and that grain boundaries were not particularly mobile; also there was limited annealing at this locality. 'Mylonitic' is a more appropriate, general qualifier for these rocks than 'mylonites', because non-brittle grain size reduction is shown by only a portion of the original quartz, and relatively little by the feldspar. They are certainly very deformed rocks, but the term 'mylonite' should be reserved for tectonites where little or none of the original grain size escaped significant reduction by dynamic recrystallization, when compared to the protolith. 'Protomylonite', following the terminology of Sibson (1977), is a suitable term for highly strained domains in the gneiss, and 'augen gneiss' for the remainder.

As mentioned in Chapter 2.2, the alteration of potassium feldspar to sericite and muscovite is locally intense, the rock resembling a mica schist, with the micaceous foliae anastomosing around quartz and relict feldspar porphyroclasts (Plate 3H). The extremely-fine grained, cherty-looking micaceous zones may be termed phyllonite (after Sibson 1977); the mica alteration appears to be replacing quartz locally. They were evidently very ductile, and mark zones of discrete shear in outcrop exposures, and are locally scored by slickensides (which are probably D_2).

Locality 700. The Western Quesnel Lake Gneiss at locality 700 (UTM 0617258368) in the north of the thesis area is a few metres or tens of metres away from the contact with the Antler Formation, and is much more strongly foliated and lineated than the gneiss at locality 16 (Plates

3A, 4F). [It has also suffered a stronger annealing, but this is due to D_2 (M_2).] The feldspars are more rounded and flattened than at locality 16, and the quartz ribbons are extremely attenuated (Plate 4F,G). These could be effects of greater D_1 strain, and (or) a higher grade of M_1 here, allowing the feldspars to deform more ductilely; it will be concluded later that the greater strain is due to a strain gradient with respect to the structurally much deeper locality 16, versus higher M_1 temperatures enhancing the ductility or strain rate.

The gneiss in this locality was also affected by D_2 flattening, as at locality 16. This is locally evident from a 'herringbone' type of recrystallized mica crenulation parallel to the S_1 mylonitic foliation, which is therefore really a composite S_{1-2} .

Potassium feldspar megacrysts form rounded augen or porphyroclasts up to a few centimetres across (Plates 3A,I; 4F). In the 'tails' of the augen, where stress was concentrated, are flattened pressure shadows with subgrains and equant recrystallized grains of feldspar, as well as quartz and muscovite (Plate 4G). Mechanical twinning and undulose extinction are generally developed. Some augen are partly replaced by an aggregate of 'new grains'; these could be both the products of dynamic recrystallization, and (or) the annealing of a brittlely-deformed and strung-out porphyroclast.

The D_1 strain and the degree of presumed D_2 (M_2) syn- to post-kinematic recrystallization are both greater than at locality 16. The main effect of the M_2 recrystallization is that the quartz in thin layers (less than 1 or 2 mm thick) and polycrystalline ribbon grains (up to 1 mm by 5 cm in sections parallel to the lineation; Plate 4F,G) derived from D_1 mylonitization is now coarser grained than at locality 16; features of D_1 dynamic recovery and recrystallization are not

obviously preserved. Grain boundaries are straight or somewhat serrated, and triple junctions are common, indicating that significant grain growth and grain boundary migration have operated. Quartz is finer grained in layers rich in mica and in pressure shadows because grain boundaries were pinned by other phases. Undulose extinction and deformation bands are present, but much less so than at locality 16; most are probably post- D_1 anyway.

Muscovite and biotite form well defined foliae, which with the ribbon quartz define the foliation (S_{1-2}). The micas wrap around the feldspar porphyroclasts, but since the feldspars are not large and are flattened, the deflection is weak.

FIRST PHASE DEFORMATION IN SNOWSHOE GROUP METASEDIMENTS IN THE IMMEDIATE LOWER PLATE OF THE QUESNEL LAKE SHEAR ZONE

Some quartzofeldspathic metasediments in the Quesnel Lake Shear Zone in the southeast of the thesis area, are remarkably finely foliated, lineated and finely compositionally banded. Some have been converted to phyllonites. The compositional banding is extremely transposed bedding.

Microstructure

As in the Western Quesnel Lake Gneiss, the microstructure in these metasediments has been modified by recrystallization during D_2 , but the main effects of D_1 strain are discernible. The polygonization of the original clasts in a rodded quartz pebble conglomerate from locality 205 attests to grain size reduction by dynamic recrystallization. The relict boundaries of the original pebbles are defined by enveloping foliae of fine grained muscovite, biotite, chlorite and clinozoisite.

The recrystallized aggregates contain inequigranular, finer grained quartz which is typically 0.1 or 0.2 mm or less across, but ranges up to 0.8 mm. Undulose extinction, deformation bands and subgrains are present, but some of this strain is probably post-D₁. Judging by the fanning array of extinction angles, the high angle boundaries in these aggregates were probably produced by subgrain rotation during D₁. Grain boundaries in the present rock, however, are extremely irregular and serrated, indicating grain boundary migration, probably during D₂. The quartz in this rock is not particularly fine grained but it is important because it can be compared with the much coarser relict texture of the protolith and the rock can accurately be called a protomylonite.

Other quartzofeldspathic metasediments in the shear zone in the Mt. Brew area are very fine grained and well foliated, but they cannot formally be designated 'mylonites' because their protoliths are not preserved for comparison. However, in the author's opinion they are definitely recrystallized mylonites (e.g. Plates 3B, 4H). In these psammitic rocks, the foliation is defined by muscovite, chlorite, biotite (usually altered to chlorite) and clinozoisite. These foliae are typically only about 0.1 or 0.2 mm (or less) apart. In between are ribbon grains of quartz (typically about 6:1 aspect ratio), usually subdivided into subgrains (Plate 4H). The quartz in some layers has high angle boundaries; these are also probably dynamically recrystallized ribbon grains but which have been annealed, resulting in rather straight grain boundaries typically subnormal to the bounding foliae, which prevented further grain growth. The overall effect in these rocks is a microscopic ribbon texture.

Some quartz-psammite mylonites have a well developed preferred crystallographic orientation (including Plate 4H), probably a girdle

fabric, discernible with the gypsum plate. Such good examples were invariably found out of place so no oriented samples were obtained for petrofabric study. In any case, it would be very difficult to defend any interpretation of analysed petrofabrics, given the D_2 (including M_2) and D_3 overprinting that has affected these rocks.

D_1 METAMORPHIC GRADE

Due to the overprinting by D_2 deformation and metamorphic recrystallization, the metamorphic grade during D_1 in the thesis area is not known precisely. Once again, the best rocks to help answer the question are those containing known S_1 fabrics.

D_1 in the Western Quesnel Lake Gneiss appears to have occurred under lower to probably upper greenschist facies conditions, at both localities 16 and 700. This is based on the 'brittle-ductile' nature of the feldspar deformation and on the presence of muscovite, sericite, biotite, clinozoisite and albite in the 'mylonitic' S_1 foliation. Dynamic recrystallization of feldspar is only common or well developed in the amphibolite facies and higher (Tullis 1983), but limited plastic deformation is possible at not excessive strain rates under upper greenschist conditions, especially if water is available to soften the feldspars by hydrolytic weakening (Tullis & Yund 1977) and by promoting replacement by white mica, albite and quartz (reaction-enhanced ductility or reaction-softening: White et al. 1980). True, the feldspar deformation in the gneiss is also compatible with the amphibolite facies, but this grade is inferred to be precluded by the D_1 production of albite (versus more calcic plagioclase) with other typical greenschist minerals. Furthermore, there is no indication in

surrounding metasediments of S_{0-1} mineral assemblages higher than upper greenschist facies.

The difference in the quartz and feldspar microstructure in the gneiss between localities 16 and 700 (and 715) is interpreted to be due to a higher strain or strain rate at locality 700 (and 715), it being closer to the 'centre' of the Quesnel Lake Shear Zone, rather than to there being higher M_1 temperatures at locality 700. Despite the high strain rate at locality 700, feldspars deformed ductilely as much as brittlely due to strain-softening mechanisms, as mentioned above; probably the availability of water in the shear zone was good, and better than at locality 16.

All these interpretations are very tentative, however, because the nature of feldspar deformation is a function of factors other than temperature and strain rate, such as pore fluid pressure, fluid composition, and lithostatic pressure (Tullis 1983).

Mylonitized Snowshoe Group metasediments in the southeast of the thesis area appear to be the most ductile tectonites in the Quesnel Lake Shear Zone. Still, there is nothing conclusive to indicate that metamorphic grade surpassed the greenschist facies during D_1 here either, such as mylonitized garnet or staurolite. (Garnet is present, but it is post- D_1 ; see Sections 3 and 4.)

Even structurally deeper in the lower plate in the thesis area, there is still no clear evidence of amphibolite facies conditions during D_1 . Where S_{0-1} is observable in F_2 fold hinges, it comprises chlorite and muscovite.

Metamorphic grade during D_1 decreased upwards in the upper plate of the Quesnel Lake Shear Zone, indicated by the decreasing development of S_{0-1} (and S_2) recrystallization and ductile deformation.

[3] 3. SECOND PHASE DEFORMATION: D₂

MAJOR STRUCTURAL FEATURES

The structural geometry of the thesis area is dominated by major folds formed during the second phase of deformation. This deformation folded the Quesnel terrane and the Omineca Belt, which were superposed during D₁ along the Quesnel Lake Shear Zone, into a pile of close to very tight, southwesterly-verging sub-recumbent folds. As a result of their moderate to gentle northwestern plunge and of an en echelon arrangement of their axes, these major folds give rise to the prominent Z-shape of the terrane boundary in the regional map pattern (Fig. 1.2).

The two main F₂ folds in the thesis area, responsible for the Z-shape, are the Seller Creek syncline and the Spanish Lake anticline; their axial surface traces are shown in Map 3 and Fig. 2.1. Two F₂ parasitic folds that make important impressions on the map pattern are the Badger Peak syncline and the Pits anticline.

The first major point to be emphasized about these D₂ structures is that they fold the Quesnel Lake Shear Zone (see Plate 4I) and are therefore unequivocally younger than the emplacement of the Quesnel terrane. The alternative, that the contact was originally depositional and that the shearing evident at the terrane contact occurred only during or after the major folding, is not tenable. In addition to the reasons given for this in Section 2 (for example, F₂ minor folds of mylonitic rocks) this alternative is rejected because if flexural shearing, for example, accompanied the D₂ folding its effects should diminish towards the hinge zones of the major folds where flexural shear is, theoretically, insignificant (notwithstanding the possibility of

hinge migration); yet some of the most strongly sheared rocks were observed right in these hinge zones, e.g. localities 52 (Snowshoe psammite) and 1062 (Antler Formation greenstone). Furthermore, if shearing accompanied or post-dated the southwesterly-verging folding it would have truncated the major folds such that the Antler Formation could not be traced around the fold closures as well as it can be.

So, all the evidence indicates that the terrane contact was sheared (D_1) before it was folded (D_2), and probably not at all (of any significance) during D_2 or afterwards. To make the point in another way, the Quesnel Lake Shear Zone cannot be dynamically implicated in the formation of the southwesterly-verging folds.

The second major point to be made about these southwesterly-verging folds is as follows. Some other workers have correlated these major structures with a generation of folds in the region which have fairly steep, northeasterly-dipping axial surfaces and associated crenulation and crenulation cleavage (Campbell 1971, Ross et al. 1985). This generation most probably corresponds to the author's third phase of folding (D_3 or F_3) which is clearly superimposed on the F_2 folds in the author's area. They are approximately coaxial. This later folding has probably tightened up or amplified the major F_2 folds, especially in the Crooked Lake area where F_2 folds are consequently more upright, and because of their near-coaxiality may give the impression that its steep crenulation or crenulation cleavage is axial planar to them. This may in effect be the case, but it is coincidental: from the author's observations the two phases can definitely be distinguished (see below), and it is the earlier of the two, the author's F_2 , that originally formed the major folds and the map pattern.

The third major point to be emphasized is that the metamorphic recrystallization that operated during (and slightly beyond) F_2 folding was the most intensive that the thesis area has suffered. The S_2 foliation is the most prominent one. The S_{0-1} foliation has been transposed by recrystallization into S_2 and is usually only distinguishable in F_2 fold hinges, and as a mylonitic fabric in the Quesnel Lake Shear Zone.

The geometry of D_2 major structures is illustrated in Map 3 (in pocket) and in the cross-sections (in pocket).

D_2 MINOR STRUCTURES

The following description applies to rocks in both plates of the Quesnel Lake Shear Zone.

Minor folds, foliation, vergence, facing, and lineation

Second phase minor folds are usually very tight and have strongly overturned to recumbent axial surfaces from which vergence was determined (Fig. 3.5; Plates 1C; 5A,B,F). They fold a composite of compositional layering and foliation, namely S_{0-1} . These folds (F_2) have an axial planar foliation (S_2 ; Fig. 3.6) which crenulates S_{0-1} in fold hinges (Fig. 3.7) and which crenulates or overgrows S_{0-1} on fold limbs (Fig. 3.8a; Plate 5C), where they have been transposed and recrystallized to form a composite fabric, S_{0-1-2} . At the contact between competent and incompetent metasediments, even on fold limbs, S_2 may form a small angle to S_{0-1} ; vergence was also deduced from this asymmetry.

The vergence indicated by F_2 minor folds and the asymmetry of S_{0-1} and S_2 surfaces is systematically related to the facing of the major F_2

Fig. 3.5

Schematic diagram illustrating the principal mesoscopic differences between D₂ and D₃ minor structures. F₂ minor folds (upper sketch) are tight, strongly overturned and 'syn-metamorphic'. F₃ minor folds (lower sketch) are upright to moderately overturned, and 'post-metamorphic'.

Fig. 3.6

Equal-area stereograms of S_2 foliation, F_2 fold axes and intersection lineations of S_2 and S_{0-1} ($S_{0-1} \times 2$). Domains shown on Map 2 (in pocket). Dashed line represents an average orientation of S_2 in that structural domain, based on the apparent maximum of S_2 poles and on the plunge of F_2 fold axes. Vergence is known for every F_2 and $S_{0-1} \times 2$ point shown, but for clarity is not indicated here: see Map 3.

Fig. 3.6 (continued)

(c) Domain III. (d) Domain IV. For F_2 and S_0-1x_2 , vergence downplunge to the northwest. is mostly Z; to the southeast, mostly S. This domain is mostly in the overturned limb of the Seller Creek syncline.

Fig. 3.6 (continued)

(e) Domain X. (f) Domain XI. Note strong northwest plunge of F_2 and $S_{0-1 \times 2}$ (cf. L_2 in Fig. 3.9b); nearly all have S vergence.

Fig. 3.6 (continued)

(g) Domain VII. (h) Domain VI. (i) Domain IX. Most F_2 and $S_{0-1 \times 2}$ lineations in southeast have Z vergence. This domain is mostly on the common upright limb between the Seller Creek syncline and the Spanish Lake anticline. (j) Domain V.

Fig. 3.6 (concluded)

(k) Domain XII. Spread of data reflects fact that this domain includes parts of both upright and overturned limbs of Spanish Lake anticline. (l) Domain VIII. (m) Domain XIII.

Fig. 3.7

(a) Micrograph of Snowshoe Group pelite from locality 660 in Badger Peak syncline. Section normal to S_2 in hinge of F_2 minor fold. Scale bar = 1 mm. Notes: Bedding defined by quartz-rich and mica-rich layers. S_0-1 preserved in microfolds between S_2 crenulation foliation, parallel to axial surface of syncline.

(b) Micrograph of Snowshoe Group pelite from locality 1054 in Seller Creek syncline (cf. chart (b) in Fig. 3.12). Section normal to S_2 in hinge of F_2 minor fold; scale bar = 0.5 mm. Notes: Bedding in fold hinge defined by quartz-rich layer (upper part) enveloped by mica-rich layer (lower part). Chlorite and muscovite of S_0-1 are preserved between spaced S_2 crenulation foliation, also chlorite and muscovite. Garnet grew in presumably Fe-Al rich layers parallel to S_0-1 ; it overgrows S_2 with no pressure shadow effect.

Interpretation: Garnet and metamorphic peak post-kinematic w.r.t. S_2 .

Fig. 3.8

(a) Micrograph of Snowshoe Group pelite from locality 695 in Seller Creek syncline (cf. chart (b) in Fig. 3.12). Section normal to S_2 and ' S_3 '; scale bar = 1 mm. Notes: garnet subidioblastic, overgrows S_2 without rotation or pressure shadows. S_4 marked by ?ilmenite grains continuous with S_e . S_2 clearly a crenulation, with S_0-1 preserved in hinges. Random, post-kinematic muscovite cuts S_2 . Warping of S_2 by F_3 microfolds (' S_3 '). Garnet not retrograded in this rock.
Interpretation: Garnet and metamorphic peak post-kinematic w.r.t. S_2 .

(b) Micrograph of Black Phyllite pelites from Seller Creek syncline (cf. chart (b) in Fig. 3.12). Upper half from locality 707; section normal to S_0-1-2 and ' S_3 '; scale bar = 1 mm. Notes: Idioblastic garnet overgrows S_0-1-2 with little or no pressure shadow. S_1 (even in retrograded mantle) is continuous with S_e (S_0-1-2). F_3 crenulations post-date garnet, producing minor pressure shadows and chlorite 'beard'. Garnet almost completely retrograded, apart from relict core.
Interpretation: Garnet and metamorphic peak post-kinematic w.r.t. S_2 .

Lower half from locality 708; section normal to S_2 , and parallel to ' S_3 ', scale bar = 1 mm. From right in hinge zone of Seller Creek syncline. Notes: In hands specimen, S_0-1 cuts across S_2 but is not parallel to F_2 axis, reflecting hinge zone folding. This feature is recognizable in micrograph only as an S_1 of black ilmenite inclusions at 90° to S_2 . Note two-stage garnet growth: early stage probably syn-kinematic with F_2 , followed by later stage post-kinematic overgrowth. Latter appears more retrograded than core. Pressure shadow around garnet probably mainly due to D_3 flattening (sketch is parallel to D_3 flattening plane).
Interpretation: Garnet and metamorphic peak post-kinematic w.r.t. S_2 . Followed by retrogression (pre- D_3 ?) and D_3 flattening.

folds that dominate the regional structure. That is, they are southwesterly-verging on-upward-facing limbs, northeasterly-verging on downward-facing limbs, and statistically non-verging in hinge zones. Their fold axes usually plunge northwest, less usually southeast or northeast (Fig. 3.6).

In some metasedimentary rocks of the Snowshoe Group, such as pebbly grits and quartzites, there is a stretching lineation in S_{0-1-2} . The rocks concerned are L-S tectonites. The lineation is not developed equally everywhere, but is very prominent in the Mt. Brew area, where the plunge of F_2 folds is quite high (about 20°). It is marked by the elongation of quartz or feldspar clasts (Plate 1B), by rodding of thin quartzose layers and trondhjemitic veins (Plate 1D), and by elongated pressure shadows (Fig. 6.10a; Plate 5G,H). This stretching lineation is approximately parallel to F_2 and F_3 minor fold axes (and to S_{0-1x2} and $S_{0-1-2x3}$) and so normally trends northwest-southeast (Fig. 3.9). It could be mistaken, therefore, to represent linear microlithons formed by the intersection of planar fabric elements. It might also be taken for a re-oriented L_1 . However, this lineation is interpreted to mark the D_2 finite extension (L_2).

Evidence:

- * It affects M_2 porphyroblasts and matrix minerals, so post-dates D_1 .
- * It is not simply a refolded L_1 because it is unlikely it would be so consistently parallel to the F_2 axis (or to S_{0-1x2}), even if F_2 was parallel to F_1 . Even if some of the finite extension was inherited from D_1 strain (by physical re-orientation) there must have been an additional extension applied in the F_2 axial direction to produce their coincidence. [Some of the finite extension, at least locally, was

Fig. 3.9 Equal-area stereograms of L₂ stretching lineations (L₂ axial extension). (a) L₂ in domains in the west and north of the thesis area; no variation between domains. (b) L₂ in domains in the southeast of the thesis area, mainly in the Mt. Brew area, where L₂ is prominent. Typical plunge here is ca. 18° to the northwest.

inherited from D_1 . The evidence for this is a strong mylonitic lineation in a (recrystallized-) mylonitized quartzite from the Quesnel Lake Shear Zone, which is parallel to L_2 . Because mylonitization was restricted to D_1 , at least some of this stretch is L_1 . L_2 extension re-oriented L_1 and possibly enhanced it.] The apparent magnitude of L_2 may thus be the sum of D_1 strain and superimposed D_2 strain; the orientation of L_2 is a D_2 feature.

* Locally the lineation is twisted around F_3 fold hinges: therefore it pre-dates the third phase of deformation.

The L_2 extension appears to have been at least locally non-coaxial (rotational). This is inferred from microtextures in thin sections cut parallel to L_2 and normal to S_{0-1-2} : here, two syn-kinematic mica fabrics (Plate 5D) and rotated porphyroblasts (Plate 5G) attest to a layer-parallel southeasterly-verging simple shear component of the strain regime (Fig. 3.10). Although this was noted in only a few rocks from the Mt. Brew area, the sense of shear is consistently top to the southeast. Oriented specimens of micaceous psammites were collected from this vicinity, and thin sections parallel to L_2 and normal to S_{0-1-2} were examined with the universal stage by the author and by Ferri (1982); the resulting quartz c-axis diagrams are shown in Fig. 3.11. Some of the specimens possess a very crude type 1 girdle fabric, but none are very convincing. Ferri (1982) managed to recognize an asymmetry in the girdles in his study, from which he inferred a top to the southeast sense of shear, which is the same as that indicated by the mica fabrics referred to above. However, another limitation of these petrofabrics (cf. the problem with D_1 petrofabric analysis) is the D_3 flattening and crenulation which affected these rocks; that is,

(a)

(b)

Fig. 3.10

(a) Micrograph of Snowshoe Group psammite from locality 1010 in Spanish Lake anticline (cf. chart in Fig. 3.12). Section normal to S₀-1-2 and parallel to L₂, looking SW; scale bar = 0.5 mm. Notes: Upper part of sketch shows irregular garnet porphyroblasts aligned in S₀-1-2 foliation with calcite, chlorite and biotite in pull-aparts. Foliation defined by muscovite, chlorite, partly chloritized biotite, and clinozoisite. Quartz shows unannealed strain. Lower part of sketch shows kinematic indicators with top to the southeast (to left) sense of shear. On left, garnet with micaceous pressure shadow clearly aligned oblique to S₀-1-2. On right, large flake of biotite (peak M₂) which was being altered to chlorite as it rotated anticlockwise. Also represented are possible shear bands (dipping to left).

(b) Micrograph of Snowshoe Group pelite from locality 153 in Spanish Lake anticline (cf. chart (a) in Fig. 3.12). Section normal to S₀-1-2 and parallel to L₂; scale bar = 1 mm. Notes: Flattening and rotation of garnets in S₂, and C & S type asymmetric mica fabrics (upper right and lower left) indicate shear parallel to L₂. [Specimen not oriented but elsewhere (e.g. Fig. 3.10a) shear sense consistently upper surface to southeast - see text.] Elongate garnets in upper left and lower right have been 'over-rotated', past plane of shear. Some retrogression of garnet and pulling apart; no post-kinematic overgrowth. Quartz grain boundaries in matrix are extremely serrated: dynamically recovered and recrystallized but not at all annealed.

Interpretation: Peak of metamorphism (garnet) syn-kinematic (r.t. S₂). D₂ flattening and especially L₂ axial extension continued post-metamorphic peak, rotating porphyroblasts and straining and reducing grain size of quartz.

Fig. 3.11

C-axis fabric stereograms (equal-area) of quartz ribbons in Snowshoe Group quartzites and quartz psammities from the Mt. Brew area. Sections oriented normal to sub-horizontal S_{0-1-2} , parallel to L_2 . (a) = locality 205, (c) = locality 213, (d) = locality 204. (b) is from Ferri (1982, from locality 205); an extra contour is included = 0.5%. Ferri inferred a top to the southeast sense of shear for (b), although this is very tenuous.

shortening parallel to S_{0-1-2} and normal to L_2 , shown by a very strong, partially-recovered flattening of quartz grains. Nevertheless, whether these petrofabrics are valid or not, the non-coaxial component related to L_2 is probably real on the basis of the other evidence. Over the whole area, however, the L_2 extension may be statistically co-axial (irrotational). The question should be investigated further.

This D_2 stretching lineation is called the L_2 axial extension, emphasizing its coincidence with the F_2 axis. Its kinematic significance has interesting implications for the D_2 tectonic regime, and this is discussed in Chapter 6.2.

Porphyroblasts

Porphyroblasts such as garnet, staurolite, chloritoid, biotite and muscovite vary spatially in their relationship with S_2 : some are syn-kinematic and some are syn- to post-kinematic (see Section 4 for details).

[3] 4. D_2 METAMORPHIC RECRYSTALLIZATION: MICROTEXTURES

The major folding and prograde regional metamorphism in the thesis area occurred during the second phase of deformation. This was, therefore, the time of the most pronounced metamorphic recrystallization, foliation development and porphyroblast growth. Temperatures reached their peak during or slightly after D_2 , so the pattern of metamorphism is a D_2 feature (see Chapter 4.1).

During F_2 folding, compositional layering and the first phase foliation (S_{0-1}) were transposed into the flattening plane of D_2 strain by a process of physical re-orientation and dynamic recrystallization,

locally followed by annealing (static recrystallization). New minerals recrystallized in the D_2 incremental flattening plane, gradually obliterating the pre-existing fabric by resorbing it, with or without prograde or retrograde mineral reaction. D_2 flattening and transposition was greatest in F_2 fold limbs, so S_2 recrystallization is the most complete there. D_2 flattening and transposition were much lower around F_2 minor fold hinges, and S_{0-1} is commonly preserved there as a crenulation (see Section 3).

S_2 foliation

The S_2 foliation in pelitic and psammitic rocks is marked mainly by muscovite and chlorite over most of the thesis area. Biotite was stable through at least part of D_2 , mainly in the east, but commonly it is at least partly retrogressed to chlorite (see Chapter 4.2).

Porphyroblasts

M_2 minerals that form porphyroblasts in the Snowshoe Group are biotite, muscovite, garnet, staurolite and chloritoid. In the Black Phyllite, garnet is restricted to a small area in the hinge zone of the Seller Creek syncline (Map 4, Fig. 4.1). [Nearby, there is an occurrence of staurolite, but as there is doubt about whether this exposure of Black Phyllite is outcrop or not, it will not be considered further. However, it will be said that this staurolite possesses all the textural characteristics common to this sub-area, as will be detailed below.]

The properties of these porphyroblasts such as their shape, continuity of internal inclusions (S_1) with the external foliation (S_2), pressure shadows versus post-kinematic overgrowth, stability with minerals in the S_2 foliation or retrogression, all provide information

on the relationship between D_2 strain and metamorphic equilibria. (Not necessarily all these criteria are observable in a given thin section; a judgment may be based on the relationships observed in a number of rocks from a given vicinity.) The observations show that this relationship is not straightforward, and that it changes over the thesis area from the north to the southeast. The relationships are illustrated in Fig. 3.12 and are summarized as follows.

(a) In the southeast of the thesis area, in the Spanish Lake anticline around Mt. Brew, there is abundant evidence that the metamorphic peak was syn-kinematic with respect to S_2 recrystallization, F_2 fold formation, and early L_2 axial extension, and that D_2 strain, especially the axial extension, continued after the metamorphic peak when the rocks were retrograding.

Evidence:

- * Syn-kinematic garnet growth, shown by rotated S_1 inclusion trails which are commonly not continuous with S_2 (Fig. 3.10b; Plate 5G).

- * Trondhjemitic veins are folded by F_2 minor folds, and are extremely attenuated and pulled-apart by the L_2 axial extension (Plate 1D).

- * S_2 is usually flattened against garnet porphyroblasts forming pressure shadows (Fig. 3.10b; Plate 5F,G), although locally there is limited idioblastic overgrowth into S_2 .

- * Small skeletal garnet and staurolite in thin mineral lineations parallel to L_2 (Fig. 3.13; Plate 5I). This suggests either that pre-existing garnet and staurolite were pulled apart by the axial extension, or that garnet and staurolite grew preferentially in a 'ferro-aluminous trail' created by L_2 extension, or both.

Fig. 3.12

Metamorphic recrystallization in relation to D₂ and D₃ deformation. Mineral abbreviations given in Table 4-1. Horizontal time axis covers D₂ and D₃ deformation; vertical axis corresponds to metamorphic grade. Neither axes are calibrated nor do the scales on (a) equal (b): they are schematic only, in that they attempt to show the spatial variations in 'prograde deformation' versus 'retrograde deformation'.

(a) Chart for the southeast of the thesis area, around the Spanish Lake anticline: highlights the fact that while deformation starts during prograde metamorphism (left half of chart), much strain and dynamic recrystallization continue during declining temperature (right half), after the metamorphic peak. Note also the implication that the onset of D₃ is mainly via transition rather than via significant hiatus. This is discussed in Chapter 6.2.

(b) In variance to (a) is the chart for the north of the area, in and north of the Seller Creek syncline in which the 'peak' of metamorphism is biased to the right or younger stages of the deformation - in fact, close to the D₂-D₃ transition. This is reflected in the textures which show much more late- or post-kinematic recrystallization than in rocks covered by (a).

It follows from this interpretation that neither the metamorphic peak nor the D₂-D₃ transition represent absolute time datums; they are both diachronous (see Chapter 6.2). For a guide, therefore, an arbitrary time datum is shown in the diagrams (diagonal shading).

Fig. 3.13

Micrograph of Snowshoe Group quartz psammite from locality 153 in Spanish Lake anticline (cf. chart (a) in Fig. 3.12). Section normal to S_0 -1-2 and parallel to L_2 ; scale bar = 0.5 mm. Notes: Staurolite (light stipple) confined to thin layers parallel to L_2 , with muscovite between the fragmental grains. Similarly garnet (heavy stipple) forms a train of grains parallel to L_2 with chlorite between. Quartz texture was formerly coarse but shows much dynamic recovery and recrystallization and serrated boundaries, and no annealing.

Interpretation: Staurolite and garnet probably formed poikiloblasts in S_2 . L_2 axial extension, probably accompanied by retrogression later which weakened the poikiloblasts, stretched and pulled them apart. Quartz strain continued without grain growth. Metamorphic peak synchronous w.r.t. S_2 , followed by retrogression during the remainder of D_2 .

* Locally garnet and staurolite have pull-aparts filled with chlorite and muscovite, parallel to L_2 (Fig. 3.13; Plate 5I), indicating that axial extension outlasted their growth and stability. Similarly, biotite was retrograding to chlorite during L_2 axial extension (Fig. 3.10a).

* Quartz is generally not annealed. In quartz-rich metasediments in sections parallel to the L_2 stretching lineation, quartz is very inequigranular and has undulose extinction, deformation bands and sutured grain boundaries (Figs. 3.10, 3.13; Plate 5I). It is dynamically recovered and partially recrystallized, but it is not statically recrystallized (annealed). This strain was superimposed on a pre-existing coarse grained quartz fabric of S_1 or early S_2 age. The later plastic strain is interpreted to be due to continued deformation, especially the L_2 axial extension, which persisted into retrograde metamorphism.

[In pelitic rocks in the vicinity, quartz is not so strained, probably due to the different deformation mechanisms in these rocks, namely sliding on mica grain boundaries.]

(b) In the north of the thesis area, around the Seller Creek syncline, biotite, muscovite, garnet, and chloritoid porphyroblasts overgrow the S_2 foliation (Figs. 3.7b, 3.8; Plate 5C,E); this feature is common to both plates of the (folded) Quesnel Lake Shear Zone. There appears to be little post-porphyroblast D_2 flattening, nor very much L_2 axial extension. This implies that the peak of M_2 metamorphism was syn- to post-kinematic with respect to D_2 .

Other evidence:

* The garnet first-appearance isograd cuts through the Seller Creek

syncline at a high angle (Map 4, Fig. 4.1); that is, it is not attenuated, truncated, tightly folded or otherwise disrupted by post-metamorphic D_2 strain.

* In garnet and chloritoid porphyroblasts, S_1 is continuous with S_e ($\approx S_2$; Figs. 3.7b, 3.8a; Plate 5E). Locally, S_1 is straight (indicating no rotation of porphyroblast during its growth), but elsewhere may be sigmoidal (indicating some syn-kinematic growth).

* Practically no pressure shadows around porphyroblasts (Plate 5C,E) (except those attributable to D_3 flattening, which is not uncommon; e.g. Fig. 3.8b).

* Idioblastic or subidioblastic porphyroblast shapes (Fig. 3.8b; Plate 5C,E).

* Large, randomly oriented porphyroblastic flakes of muscovite overprinting S_2 (Fig. 3.8a).

This spatial textural variation in the relationship between D_2 strain and metamorphic recrystallization has interesting implications, which are discussed in Chapter 6.2 (see also Chapter 4.1).

Retrogression

Peak D_2 mineral assemblages throughout the thesis area underwent a general retrogression. As described in the previous sub-section, the retrogression began during D_2 in the southeast of the area, and probably began after D_2 in the north. In both areas it probably lasted until the early stages of D_3 because it does not have much, if any, kinematic relationship with ' S_3 ' (see Section 5).

The retrogression is manifested by the alteration of biotite, garnet and staurolite to muscovite and chlorite. Locally, large, randomly

oriented flakes of chlorite overprint the S_2 fabric. Not all biotite and garnet are altered, and not all thin sections show it: structurally deeper into the Omineca Belt (i.e. farther east), the retrogression is minor, or at least only localized. Locality 656, for example, is barely affected. [In agreement with this, Getsinger (1985) shows a "retrograded garnet zone" on the other side of Grain Creek - see Map 4.] Structurally higher, and towards the boundary with the Quesnel terrane, the replacement of biotite by chlorite is usually almost complete, where it can be inferred (biotite seems to be very vulnerable). Garnets in the Black Phyllite in the Seller Creek syncline are replaced by a mat of very fine chlorite, except perhaps for a small relict core (Fig. 3.8b; Plate 5E), and locally the garnet porphyroblasts can only be recognized by their distinctive shape. However, their former kinematic relationship with S_2 is still clearly discernible (they overgrow it). Garnet in the adjacent Snowshoe Group is better preserved, for some reason.

Retrogression in the southeast in the Spanish Lake anticline around Mt. Brew is quite conspicuous, where it occurs, because of the medium to coarse grain size of the schists and paragneisses. Because retrogression started during D_2 here, it may have been catalysed by the strain energy.

[3] 5. THIRD PHASE DEFORMATION: D_3

General features

This deformation produced crenulations (' S_3 ') of pre-existing foliations (Plate 5A), and flexural slip-like warps and corrugations of compositional layering and of F_2 folds on the scale of centimetres to kilometres. The axial surfaces are overturned to the southwest or

northeast but are never recumbent; they dip moderately to steeply (ca. 55° to 85°) and trend consistently northwest-southeast (Fig. 3.14). D_3 crenulations are usually southwesterly-verging in the Omineca Belt; to the southwest in the Quesnel terrane, northeasterly- and southwesterly-verging crenulations are about equally developed (Fig. 3.14).

This deformation was responsible for tightening-up some D_2 folds, and for warping D_2 axial surfaces. It was essentially post-metamorphic in the sense that there was not very much metamorphic recrystallization effecting the formation of F_3 crenulations: strain was ductile but appears to have taken the form of a plastic flattening of quartz and buckling of micas (see below).

D_3 minor structures

Third phase minor folds are more open than F_2 , being open to closed or locally tight (Fig. 3.5; Plate 1D,I). They fold a composite (S_{0-1-2}) of compositional layering (S_0), a foliation (S_1) and a crenulation foliation (S_2), and also F_2 axial surfaces (see cross-sections). In the axial surfaces of F_3 minor folds or of the pervasively-developed crenulations there is generally no new foliation (Fig. 3.8; Plate 5E), although for convenience in descriptions, these 'surfaces' are called ' S_3 '. No stretching or mineral lineation attributable to D_3 was recognized.

F_3 folds and crenulations are usually southwesterly-verging, on the overturned limbs of the major F_2 folds as well as on their right-way-up limbs; this proves that they are not consanguineous with the F_2 folds.

Northeasterly-verging minor folds and crenulations exhibiting otherwise identical properties are considered to belong to the same phase (F_3) since no cross-cutting relationships were found to establish superposition. In fact, in one locality both southwesterly-verging and

Fig. 3.14

Equal-area stereograms of 'S₃' crenulation 'foliation' and F₃ fold axes. Data is separated into two sets: (a) covers structural domains in the southwestern half of the thesis area and (b) covers domains in the northeastern half. Note that in the southwest (a), northeasterly-verging and southwesterly-verging crenulations are equally developed, whereas in the northeast (b) southwesterly-verging crenulations are markedly dominant.

northeasterly-verging crenulations are developed and they pass into one another, suggesting that they are consanguineous; it is possible that they are conjugate. Usually, however, an exposure has either southwesterly-verging crenulations or northeasterly-verging crenulations and not both.

F₃ 'major folds'

See Section 6.

D₃ metamorphic recrystallization

As previously mentioned, there was no major development of a foliation or mineral lineation in D₃. The fold mechanism was mainly flexural flow, most of the strain being effected by the bending of S₀₋₁₋₂ micas (Plate 5E) and by the plastic flattening of quartz. Micas are curved, and show undulose extinction or subgrain development; only in very tight F₃ microfolds where internal strain was high are they dynamically recrystallized. Rarely a few flakes of muscovite or chlorite occur parallel to the axial surfaces of the crenulations, but locally even these are kinked or buckled by 'S₃'.

Similarly, in moresiliceous metasediments, quartz exhibits undulose extinction, deformation bands and subgrains attributable to D₃ flattening, but shows limited recrystallization. Quartz in pelitic rocks does not show this strain so much because most of the flattening or shear was taken up by mechanisms involving micas.

These observations imply that the metamorphic grade was decreasing during D₃, and certainly had decreased since D₂ (see Chapter 4.2 and Chapter 6.2).

[3] 6. GENERAL COMMENTS ON THE EFFECTS OF THE STRUCTURE
ON THE MAP PATTERN

KLIPPEN AND WINDOWS OF THE QUESNEL LAKE SHEAR ZONE.

No windows of the Snowshoe Group through the Quesnel Lake Shear Zone were found within the Quesnel terrane, with the possible exception of an exposure at locality 145 of psammites and pelites very typical of the Snowshoe Group and unlike anything in the Quesnel terrane. These rocks are surrounded by a few metres of Antler Formation and then by Black Phyllite, and it is suggested that they could be lower plate rocks that were squeezed up in the core of the tight F_2 and (or) F_3 anticline (the Blackbear antiform - see below) in this vicinity.

Several klippen of the Antler Formation of the Quesnel terrane occur within the Snowshoe Group. The largest is in the core of the F_2 Badger Peak syncline, a parasitic fold on the inverted limb of the Seller Creek syncline (see Map 3 and cross-sections). F_2 minor folds systematically change vergence all around this fold (Fig. 3.15), proving that it is an infolded klippe. [Incidentally, the folded Quesnel Lake Shear Zone is not conspicuous here because the underlying Snowshoe is mainly pelitic, but there are some finely foliated quartzites with annealed quartz ribbons, and Antler Formation rocks are porphyroclastic.]

A possible exposure of Antler Formation on the slope south of Browntop Mountain is at about the same structural level as the Badger Peak syncline, and could be a continuation of this eroded and folded F_2 klippe. Incidentally, connecting the hinge lines of these two klippen in a strike section provides an accurate measure of the local F_2 plunge, which is about 2° to the northwest.

Small klippen of Antler Formation occur on the ridge south of Mt.

Fig. 3.15

Detailed cross-section through Badger Peak showing Badger Peak syncline with Antler Formation (unpatterned) in core. Dots = Snowshoe Group, crosses = Western Quesnel Lake Gneiss. Dotted line = topographic surface. Solid lines = S₀-1, dashed lines = S₂. Shape of F₂ minor folds shown by 'S', 'Z' and 'M'.

F₂ minor structures and S₀-1 - S₂ asymmetry were observed all around the fold, indicating F₂ vergence. Boundary of Antler Formation is Quesnel Lake Shear Zone; it is marked locally by recrystallized mylonitic quartzite, but generally adjacent Snowshoe is pelitic and not obviously highly strained.

Note the thin sheet of Snowshoe Group within the Antler Formation: it probably represents either an F₁ fold or a thin imbricate slice incorporated into the Antler Formation during D₁ shearing.

Brew (Map 1 and cross-section K-K'-K''). These are erosional remnants of the Quesnel terrane, preserved in small, probably F_3 synforms. Smaller, isolated lenses of meta-ultramafic rocks in the Mt. Brew area are interpreted to be sheared and infolded remnants of the basal Antler Formation (see Chapter 6.2 for an explanation of their formation).

Cariboo Lake ?klippe

One tantalizing question concerns the significance of the rocks on the western shore of the southern part of Cariboo Lake, immediately south of Keithley Creek. The black graphitic siltstones and phyllites here closely resemble unit uTrBP, the Black Phyllite. Directly across the lake, locality 357 is indistinguishable from the Black Phyllite. These outcrops are quite isolated from the Black Phyllite of the Quesnel terrane, being completely surrounded by Snowshoe Group metasediments and the Western Quesnel Lake Gneiss.

These rocks are interpreted, though somewhat reservedly, as an infolded and probably faulted klippe of Quesnel terrane.

Evidence:

* Most importantly, these rocks contain one exposure of highly sheared chloritic greenstone and meta-ultramafic schist. Both rock-types are very similar to those found within the Antler Formation, and they are tentatively identified as such. (No meta-ultramafic rocks have been found by the author in the Snowshoe Group, and none have been reported by other workers, to the author's knowledge.)

* While black graphitic siltstones and phyllites do occur within the Snowshoe Group e.g. subunit HPSH, the Cariboo Lake rocks are much more similar to unit uTrBP. They also contain completely altered

(?carbonatized) ?basaltic rocks virtually identical to those found locally within unit uTrBP and unit TrJT.

* Snowshoe metasediments immediately adjacent to these rocks are particularly strongly deformed, as those at the Quesnel Lake Shear Zone.

* F_2 minor folds in this stretch of rocks have variable asymmetry and vergence, and some are symmetrical. These features suggest an F_2 hinge zone.

According to the structural reconstructions, an F_2 syncline (see cross-section C-C'-C'') does indeed occur in this vicinity (actually the northwestern continuation of the Badger Peak syncline). However, the hinge zone of the fold here contains lower plate rocks, not upper plate rocks; there doesn't seem to be any practical way of infolding the Antler Formation and Black Phyllite to the extent needed to account for their postulated existence on south Cariboo Lake. It remains an enigma.

REGIONAL Z-SHAPED FOLD

The large Z-fold that dominates the map pattern is the result of the major en echelon northwestward plunging, southwesterly-verging sub-recumbent F_2 folds, the Seller Creek syncline and the Spanish Lake anticline. On the upward-facing common limb between these two folds are corrugations caused by F_2 parasitic folds, the largest being the Pits anticline.

The form of the terrane boundary is also affected by open F_3 folds, the main one being the Blackbear antiform which buckles and accentuates the crest of the Spanish Lake anticline northeast of its hinge zone (Map 3 and cross-sections). As a result, the Antler Formation southwest of

Mt. Brew has a gentle, undulose dip and northwesterly plunge, and occupies an unusually large outcrop area (although there may be windows or outliers of bounding rock units inside it which were not found during mapping). Furthermore, bedding and S_2 foliations have a southwestward dip on the southwest flank of this F_3 structure as a result of the flexure; normally they would be flat or dip northeastwards.

It appears that the Pits anticline is also buckled or accentuated somewhat by a superimposed F_3 fold (see cross-section J-J'-J").

F_3 folding warps bedding, S_2 and F_2 axial surfaces in a few other areas, such as in an antiform-synform pair on the ridge north of Goose Peak, so that they dip gently southwestwards instead of northeastwards.

The Eastern Quesnel Lake Gneiss appears to be in the core of the Spanish Lake anticline.

[3] 7. FAULTING

No major brittle faults have been mapped in the thesis area, although there are numerous small faults. Most of them trend north to northeast; their effects can be traced for up to about 3 km on the surface, from offsets or topographic expression. Some fault dips are known to be subvertical from topographic relief; the remainder are assumed to be steep also, from their straight outcrops.

The throws on most of the faults could not be determined, though the two faults just west of Mt. Brew (Map 1) appear to be a small graben structure. The amount of displacement on them is inferred to be small; there is certainly no marked metamorphic discontinuity associated with this faulting, nor with any other faults.

The two northeasterly-trending faults in the Cariboo River area are hypothetical, and are inferred in order to explain the offset of the Quesnel Lake Shear Zone, and the poor link between the geology on either side of the river. There is not much independent evidence for these faults: exposure there is poor, especially on the west bank. At one locality (112) on their trend, however, Snowshoe Group metasediments are fairly deeply altered, reminiscent of fault-zone alteration.

[3] 8. SUMMARY OF DEFORMATION AND METAMORPHIC RECRYSTALLIZATION

The deformation in the thesis area is dominated by two major consecutive phases. The first phase (D_1) was the tectonic emplacement of the Quesnel terrane allochthon onto the Snowshoe Group and Western Quesnel Lake Gneiss, forming the Quesnel Lake Shear Zone, which is marked by high strain and mylonitic fabrics (S_1 and L_1). Various D_1 kinematic indicators demonstrate a northeastward or eastward sense of shear (depending on the local orientation of L_1). Larger-scale D_1 structures such as folds were not positively identified.

Available evidence on mineral assemblages and nature of feldspar deformation indicates that D_1 metamorphic grade was middle to upper greenschist facies in and below the Quesnel Lake Shear Zone.

The second major phase (D_2) deformed the now coupled Quesnel terrane and Snowshoe Group into large-scale, syn-(peak) metamorphic, southwesterly-verging sub-recumbent folds, which plunge to the northwest. They produce the prominent Z-shape of the terrane boundary in the regional map pattern. D_2 deformation unequivocally post-dates the emplacement of the Quesnel terrane, and coincides with the main

phase of metamorphism (M_2) and recrystallization (S_2) in the region. A stretching lineation (L_2) occurs parallel to F_2 fold axes and to the S_0 - L_2 intersection (northwest-southeast).

The relationship between the porphyroblasts and S_2 varies across the thesis area. Essentially, the porphyroblasts are idioblastic and post-kinematic in the north around the Seller Creek syncline, and the garnet first-appearance isograd cuts through the syncline at a high angle. In the southeast in the Spanish Lake anticline, porphyroblasts are non-idioblastic and syn-kinematic, and are pulled-apart in L_2 and retrograded. A coarse quartz texture is dynamically recrystallized but not annealed. Thus, with respect to S_2 and D_2 strain, peak metamorphism in the southeast precedes that in the north.

Third phase deformation (D_3) produced northwesterly-trending crenulations (' S_3 ') and open to moderate folds which are overturned to the northeast or usually to the southwest. D_3 was essentially 'post-metamorphic' because very little axial planar foliation or recrystallization is developed.

No major brittle faults have been mapped. Minor faults trend north or northeast; they are probably steep and appear to have relatively little displacement.

CHAPTER FOUR

METAMORPHIC CONDITIONS IN SPACE AND TIME

[4] 1. INTRODUCTION

PREAMBLE

Regional metamorphism in orogenic belts is extremely complex because the geothermal (P-T) regime continually changes in time (it is dynamic), especially in response to a major phase of deformation. The peak metamorphic assemblage (p.m.a.) in a rock represents only one point or short segment in what might be overall a very complex P-T-t (pressure-temperature-time) path for that rock. A map of p.m.a.'s is only representative of a real geothermal regime if all the p.m.a.'s formed, effectively, instantaneously. A p.m.a. map of a large area, especially one with structural discontinuities or a regional plunge such that a range of crustal levels and tectonic packages are represented, does not necessarily represent a systematic P-T array (geothermal regime) because of the likely polychroneity of 'quenching'. It would be a map of a P-T-t array. P.m.a maps of smaller areas may not possess such a complex P-T-t array, but not necessarily: the processes that affect quenching can vary over small distances, as will be shown in this thesis.

The point is that it's as important to interpret an area in terms of 'isochronous' mineral assemblages, whether they are all metamorphic maxima or not, as it is to consider a map solely in terms of peak mineral assemblages which are certainly not guaranteed to be isochronous; an isochronous metamorphic map is a closer approximation to a real geothermal regime and thus a more faithful guide to fossil thermal gradients, "metamorphic facies series", and P-T-t paths. An isograd map constructed without these considerations potentially represents a misleading pattern of P and T, one that never really existed as such.

One way to constrain the time interval through which mineral assemblages formed in a polychronic metamorphic terrane is to relate them to a structural fabric, if it can be assumed from geochronometry or tectonic constraints that the deformation was more or less isochronous or relatively short-lived. This is usually done as a matter of routine, but it is worth emphasizing. The geothermal regime will have evolved during that deformation, but textural and fabric analysis may allow it to be reconstructed (especially if geothermometric and geobarometric data is available as well).

METAMORPHIC MAP

As described in Chapter 3, M_2 was the most significant phase of metamorphic recrystallization in the thesis area. In the attempt to produce an isochronous M_2 map (Map 4 in pocket, Fig. 4.1) care was taken to distinguish M_2 minerals based on their association with S_2 , and to screen out any relict M_1 or overprinting M_3 minerals. (It was not practical to produce also 'isochronous' metamorphic maps for D_1 and D_3

Fig. 4.1

General metamorphic map of the thesis area. Isograds and zones in northeast from Getsinger (1985).

because the data is so sparse.

In more detail, however, it is apparent that even within this single phase of deformation (D_2), the geothermal regime was complex. It was described in Chapter 3.4 how the peak mineral assemblages vary in their relationship to the S_2 strain fabric. Where porphyroblastic growth of biotite, muscovite, garnet, and chloritoid (reaction recrystallization in response to T increase) apparently overprints the S_2 strain fabric (recrystallization related to shape change) in the north of the area around the Seller Creek syncline, it may be inferred that prograde metamorphism persisted after the folding and most of the D_2 strain was completed. In the southeast of the thesis area around the Spanish Lake anticline, however, garnet and staurolite (peak metamorphic minerals) were involved in strain-related recrystallization (they were rotated after growth, and have pull-aparts and pressure shadows associated with retrogression) and do not significantly overprint S_2 . That is, much S_2 foliation seems to have formed here after the peak-grade porphyroblasts grew, and at lower grade. It may be inferred, therefore, that here prograde metamorphism was prior to or synchronous with F_2 folding and L_2 axial extension, and that during or certainly by the end of this deformation, isotherms were relaxing (retreating) because the garnet and staurolite underwent retrogression to chlorite and muscovite during at least part of D_2 and certainly during the L_2 axial extension, and quartz is only weakly annealed.

This textural variation means either that the folding was isochronous and the metamorphic peak was diachronous (older in the south than in the north) or that peak metamorphism was isochronous and deformation was diachronous or migratory (younger in the south than in

(the north). The preferred alternative is that deformation was isochronous and metamorphic quenching was diachronous. This is because structures tend to form quite quickly and continuously, and isotherm migration is slower and so more likely to be a direct or indirect consequence of rock movements.

Assuming this is correct, then the M_2 metamorphic map is not 'isochronous' at all but is a P-T-t array and not strictly a representation of a real geothermal regime. The deviation from isochroneity is probably relatively small; this is not the central point. The most interesting question is: why was the metamorphic peak diachronous?

Another phenomenon that needs to be explained is the steep metamorphic gradient across the Intermontane-Omineca boundary, and the way it conforms to the shape of the boundary (Fig. 4.2).

The answers to both these questions are discussed in Chapter 6.2. The main point of this introduction has been to emphasize the importance of the time factor in the analysis of a metamorphic terrane - the constraining of the geothermal regime. We have seen that by doing this it has been demonstrated that the garnet isograd is diachronous.

Aim of this chapter

The rest of this chapter deals first (Section 2) with M_2 mineral assemblages, assessing the prevailing P and T conditions of their formation based on published or unpublished estimates of mineral equilibria. M_1 and M_3 metamorphic grade is briefly mentioned at the end. Section 3 briefly describes the isograd patterns on the regional scale.

Fig. 4.2

Isograd map of the Quesnel Lake region. Antler Formation in black, Quesnel Lake Shear Zone marking its northeastern side. Isograds (dashed lines) and trace of Antler Formation west of North Arm of Quesnel Lake based on this thesis and Getsinger (1985); east of North Arm taken from Campbell (1978).

[4] 2. P AND T OF METAMORPHIC FACIES

APPROACH

In the following, the P and T conditions of metamorphism are estimated by the application of observed mineral assemblages to a petrogenetic grid. The grid (Fig. 4.3; mineral abbreviations in Table 4-1) is one compiled largely from experimental results, as discussed by Winkler (1979). The objective is to derive a line or field on the grid which covers the P and T range represented by M_2 assemblages in the thesis area. This can then be used in tectonic reconstructions (see Chapter 6.2).

Geothermometric and geobarometric techniques were not investigated in this thesis, mainly because of the problem of retrogression: for example, Fe-Mg partitioning between garnet and biotite, a common method for pelitic rocks, cannot be used here because these minerals are generally not in textural equilibrium.

Rock systems

P and T conditions are assessed from mineral assemblages in three different, general whole-rock systems, which are widely distributed in the thesis area: pelitic rocks (Snowshoe Group and Black Phyllite), mafic rocks (Antler Formation and parts of the Black Phyllite and Takla Group), and ultramafic rocks (Antler Formation).

In the prevailing greenschist to amphibolite facies of metamorphism ("low to medium grade" of Winkler 1979), pelitic rocks are more sensitive than mafic and ultramafic rocks to variations in P and T, and so are the most useful here. Ultramafic rocks are very sensitive to the H_2O/CO_2 ratio in the fluid phase, and their mineral assemblage patterns

Fig. 4.3

Petrogenetic grid comprising reactions pertinent to mineral assemblages in the thesis area. Other reactions not referred to in text are for reference only. Grid is mainly for the general pelitic system, although reactions 1 to 4 at low temperatures, shown by double lines, are for the mafic rock (ca. metabasalt) system. Mineral abbreviations are given in Table 4-1. 100 MPa = 1 kilobar. Depth is calculated assuming a mean density of middle and upper crust of 2.75 g cm^{-3} (100 MPa = 3.7 km). Pressure is lithostatic pressure = fluid pressure which is assumed to be PH_2O .

Reactions were compiled from numerous sources, with the general assumption that $a_{\text{H}_2\text{O}} = 1$. Reactions 1, 3 & 4 from Winkler (1979), reaction 2 from Turner (1981), reactions 5, 9 & 11 from Archibald et al. (1983), reaction 6 from Brown (1969), reaction 7 & 10 from Hoesche (1969), reaction 8 from Carmichael (unpublished). Minimum melting curve in presence of water from Turner (1981). Facies boundaries shown by dashed lines (Turner 1981).

Stippled area is estimated P-T range represented by rocks in the thesis area, also called the 'field gradient' (see later).

TABLE 4-1: MINERAL ABBREVIATIONS

ac	actinolite
ad	andalusite
as	Al_2SiO_5
bi	biotite
cc	calcite
ch	chlorite
cr	chromite
ct	chloritoid
cz	clinozoisite
do	dolomite
ep	epidote
ga	garnet (almandine)
he	heulandite
il	ilmenite
Kf	potassium feldspar
ky	kyanite
lw	lawsonite
ma	magnetite
mg	magnesite
mu	muscovite
pc	plagioclase
pp	pyrophyllite
pr	prehnite
pu	pumpellyite
qz	quartz
se	serpentine
si	sillimanite
st	staurolite
tc	talc
ti	titanite
to	tourmaline
tr	tremolite
zo	zoisite

may reflect this more than P and T variations.

Mafic volcanic rocks of the Takla Group dominate the subgreenschist ("very low grade") part of the thesis area in the extreme south, which is convenient because they are the best indicators of metamorphism at this grade (Winkler 1979).

Where temperatures of mineral equilibria are given below, it is assumed that the activity of $H_2O = 1$, and that $P_{H_2O} = P_{total}$.

MINERAL ASSEMBLAGES IN PELITIC ROCKS

This rock group includes metasediments with at least 20% micas and aluminous minerals such as garnet and staurolite, and for convenience here, also includes psammitic rocks - those with about 5 to 20% of these minerals. These percentages are somewhat arbitrary.

The metamorphically significant minerals that have been found are staurolite, garnet, chloritoid, biotite, muscovite, chlorite, clinozoisite, epidote, and sodic plagioclase; no Al_2SiO_5 polymorph was found. The garnet is presumed to be almandine-rich. The Snowshoe garnets were not probed by the author, but analyses in other areas by Getsinger (1985: 28-37 wt. % FeO), Campbell (1971: 30%) and Montgomery (1985: 33-35%) all indicate that garnets are almandine-rich; Getsinger described garnet zoning in detail. Garnets in the Black Phyllite in the thesis area were probed, and also found to be almandine-rich (Table 4-2).

Typical pelitic assemblages observed are shown in Table 4-3, and graphically in Fig. 4.4. The assemblages are labelled P.1 to P.8, and are grouped into zones and facies; these form the basis of the metamorphic map, Map 4.

TABLE 4-2: MICROPROBE ANALYSES OF GARNETS IN THE BLACK PHYLLITE

Sample		FeO	MgO	MnO	Al ₂ O ₃	CaO	SiO ₂	TiO ₂	Cr ₂ O ₃	Total
Locality 707 Garnet 1	CORE	31.38	1.89	2.25	21.63	6.30	37.18	0.06	0.0	100.70
	MID.	31.38	2.05	1.97	21.80	6.38	36.93	0.07	0.02	100.60
	RIM	31.66	2.13	1.78	21.66	6.34	37.40	0.06	0.07	101.10
Locality 707 Garnet 2	CORE	30.99	1.47	2.94	21.65	6.90	37.08	0.07	0.01	101.10
	RIM	31.72	1.65	1.99	21.64	6.81	36.83	0.11	0.01	100.80
Locality 707 Garnet 3	SPOT 1	31.63	1.66	2.11	21.44	6.11	37.11	0.06	0.02	100.10
	SPOT 2	29.44	1.44	4.83	21.31	6.13	37.34	0.12	0.02	100.60
Locality 708 Garnet 1	CORE	29.27	1.27	5.24	21.69	6.45	36.14	0.10	0.0	100.20
	MID.	29.48	1.25	5.63	21.53	6.52	36.28	0.08	0.02	100.80
	RIM	30.10	1.39	4.75	21.82	6.48	36.50	0.06	0.03	101.10
Locality 708 Garnet 2	CORE	29.67	1.26	5.19	21.57	6.78	37.03	0.10	0.02	100.60
	RIM	32.35	1.70	2.22	21.54	5.48	37.12	0.08	0.02	100.50

Notes: Analysed on a Cambridge Mark V electron microprobe using wavelength spectrometers.
 (Accelerating potential = 15 kv, beam diameter = 2-3 microns, specimen current =
 50 nanoamperes, as measured on pure iron). Total Fe as FeO. Garnets are largely
 retrogressed; analyses made where possible on relict crystals.

TABLE 4-3: MINERAL ASSEMBLAGES IN PELITIC ROCKS

FACIES	ZONES	ASSEMBLAGE	PHASES										
subgreenschist/ lower greenschist	CHLORITE	P.1		ch	mu	qz	pc			il	cc†	to	
		P.2		cz	ch	mu	qz	pc	Kft	tr†	cc†		
middle greenschist	BIOTITE	P.3	bl	ch	mu	qz	pc	ch*	Kft			to	
		P.4	bl	cz	ep	ch	mu	qz	pc	Kft			
upper		P.5	ga	cz†	ch	mu	qz	pc	ch*	il		to	
greenschist	GARNET	P.6	ga	ct	ch	mu	qz	pc	ch*	il			
		P.7	ga	bl	cz†	ch	mu	qz	pc	ch*	il	cc†	to
lower amphibolite	STAUROLITE	P.8	st	ga	bl		mu†	qz	pc	ch*	mu*		to

Notes: Assemblages based on textural equilibrium observed in thin section.

* = retrograde mineral, † = not in all thin sections in this assemblage. Plagioclase where determined is sodic, with less than ca. 15% anorthite. Ilmenite (il) may be represented by its oxidized form leucoxene. Other minor minerals that may be present include apatite, titanite and zircon, and graphite in Black Phyllite pelites.

Fig. 4.4-

AFM diagrams for mineral assemblages P.3 to P.8 in pelitic rocks. A = Al₂O₃ - 3K₂O, F = FeO, M = MgO. Mineral compositions taken from Winkler (1979). Range of assemblages partly reflects range of bulk compositions.

The distinction between prograde and retrograde chlorite is not always easy, but it appears to have been stable when garnet was formed in these rocks. The absence of biotite in an assemblage may mean it never grew, or that it has all been retrogressed to chlorite; the alteration of biotite to chlorite seems to go to completion more readily than does the alteration of garnet. Because of this uncertainty, the biotite isograd is tentative in that it merely marks the preservation of the mineral (Table 4-3, Map 4), and not necessarily the original maximum extent of the biotite zone.

The only well-defined first-appearance isograd in the area is that of almandine garnet (see also Section 3), so there is only a limited indication of the spatial pattern of metamorphism. The reactions responsible for the appearance of almandine garnet are probably the continuous reactions:

(see Thompson & Norton 1968 and Chakraborty & Sen 1967, respectively). These reactions are bulk-composition dependent, but are estimated to occur around 450°C based on the general appearance of almandine garnet in the upper greenschist facies.

Another possible garnet-forming reaction, which has been roughly calibrated by Brown (1969; Fig. 4.3), might well be relevant to the garnet isograd separating assemblages P.4 and P.7 in the Mt. Brew area (Map 4):

Chloritoid was found in only one locality (1083; assemblage P.6) in the northeast of the thesis area, reflecting its restriction to rocks of a certain bulk composition. Chloritoid is normally characteristic of the greenschist facies and may form by the following reactions:

(Hoschek 1969). The first reaction is plotted in Fig. 4.3.

Metamorphism in the Mt. Brew area

M₂ metamorphic grade is highest in the southeast of the thesis area, southeast of Mt. Brew. The relatively high grade here (particularly around locality 153) is indicated by the coarse grain size of schists and paragneisses, by the local presence of staurolite, and by the presence of trondhjemitic 'migmatites' (see below and Chapters 2.1 and 3.3, 3.4).

The appearance of staurolite, which marks the beginning of the amphibolite facies or "medium grade" (Winkler 1979), may be due to the reaction:

(Carmichael 1970 and unpublished data), and (or) the higher temperature continuous reaction:

(Hoschek 1969; Fig. 4.3). The first reaction indicates a temperature of about 500°C at moderate pressures. Staurolite probably did not form

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from the breakdown of chloritoid in this area because chloritoid is absent in the adjacent rocks downgrade, probably because of compositional restrictions (see Winkler 1979). These very localized staurolite-bearing rocks (assemblage P.8) are probably the only rocks in the thesis area in which chlorite became unstable at some point during M_2 , because they also contain muscovite, biotite and garnet; that is, the garnet-chlorite tie-line would have been broken between assemblages P.7 and 8 (Fig. 4.4).

It is interesting that some pelites in this staurolite-zone locality consist of much chlorite and garnet, but no staurolite or muscovite. It was probably the lack of muscovite in these particular compositions that prevented the above reaction(s) from proceeding and producing staurolite.

Trondhjemitic migmatites. Some pelitic rocks in the Mt. Brew area are called trondhjemitic migmatites (*sensu lato*) because they contain concordant and discordant veins of quartz + plagioclase (see Chapter 2.1). The veins are syn-kinematic with respect to D_2 (see Chapter 3.3).

The origin of trondhjemitic veins is problematical because they are quite common in metamorphic terranes well downgrade of the appearance of granitic veins which represent minimum melting conditions in the upper amphibolite facies (see Winkler 1979). The most likely explanation is that greenschist and amphibolite facies trondhjemitic veins do not represent partial melting at all, but rather hydrothermal processes involving metasomatism (open system) and (or) metamorphic differentiation (open or closed system), in which sodic plagioclase accompanies quartz in metamorphic solutions (Yardley 1978, Ashworth 1985, Turner 1981, p.465). If the host-rock contains potassium feldspar, this may also be incorporated into the vein, but if not (as

these Snowshoe rocks) potassium feldspar will not form by this process because muscovite + quartz is still stable. Melanosomes and restites are uncommon (Bradbury 1979), possibly because the source of the trondhjemite in the veins is not necessarily the immediately surrounding host-rock or paleosome.

The evidence indicates that the trondhjemitic migmatites in the Mt. Brew area had this kind of subsolidus origin.

Evidence:

- * Anatexis is ruled out because the grade indicated by the paleosome is much too low (mu + qz, and even ch are stable), and because the veins do not have a granitic 'minimum melt' composition (no potassium feldspar).

- * Feldspars are poikiloblastic and do not appear to have crystallized from a melt. A most important point is that these 'poikiloblasts' are common in isolated spots or in trains, as well as in quartzofeldspathic veins and lenticles.

- * There is no known igneous source for the veins (by injection of apophyses).

It is concluded that these migmatites formed by a subsolidus, hydrothermal process, probably metamorphic segregation or differentiation.

Barr (1985) studied non-anatectic migmatites in Moinean pelites in Scotland, and made a correlation between the segregations and zones of high strain, suggesting that the strain enhanced fluid accessibility and promoted reaction [mineral growth], because a large surface area was exposed during "intense grain size reduction". A similar argument can be made for the Mt. Brew migmatites, as they are syn-kinematic and very

strongly stretched in L_2 , along which fluid transport and diffusion were probably concentrated. Thus, the veins may have had a basically linear form (parallel to L_2) to start with.

It is noted that the trondhjemitic veins appear in an area approximately at the transition between the upper greenschist facies and the lower amphibolite facies. It is suggested that excess fluid pressures may have been generated at this transition by dehydration reactions, particularly those involving the prograde breakdown of chlorite.

Metamorphism in pelites in the Three Ladies Mountain - Mt. Stevenson area of Getsinger (1985)

Getsinger (1985; see Chapter 1.2) has mapped pelitic rocks in the Three Ladies - Mt. Stevenson area from the biotite zone to the sillimanite zone (Fig. 4.2). The sillimanite zone is in the east-central part of this area, on the North Arm of Queenel Lake, and metamorphic grade decreases to the north, west and south. There is also a pressure increase from north to south, based on garnet-zoning profiles and microscopic textures, although this is not supported by geobarometry which indicates no gradient. Garnet in the southwest of this area, which approaches the Mt. Brew area of the present thesis, is retrograded, as it is in the Mt. Brew area. Getsinger also reported retrograde textures involving biotite and garnet altering to chlorite and sericite, in this zone.

Getsinger found no unequivocal evidence for more than one episode of prograde metamorphism involving garnet, staurolite or kyanite.

Using the grossular-anorthite geobarometer, Getsinger obtained pressure estimates ranging from 460 to 660 MPa. Due to the resolution

of the analytical procedures and other restrictions, this range could not be narrowed. She concluded that the average pressure was 550 ± 70 MPa.

This is also the pressure estimated for peak metamorphism by Montgomery (1985; see Chapter 1.2) from phase relations in the Isosceles Mountain area, southeast of Quesnel Lake. Similarly, Fillipone (1985; see Chapter 1.2) estimated that assemblages in the nearby Boss Mountain area were formed at between 500 and 600 MPa.

MINERAL ASSEMBLAGES IN MAFIC ROCKS.

The main mafic rocks under consideration are the metabasalts of the Antler Formation, which transect the thesis area from northwest to southeast. The mineral assemblages and the inferred metamorphic grade do not vary very much; what variation there is in mineral assemblages is most likely due mainly to bulk composition, and does not necessarily conform with the metamorphic zonation observable in adjacent pelitic rocks. Typical assemblages are shown in Table 4-4 and graphically in Fig. 4.5, and form the basis of Map 4.

All the assemblages in the Antler Formation (assemblages M.3 to 5, Fig. 4.5) are characteristic of the greenschist facies, or "low grade metamorphism" of Winkler (1979). Blue-green hornblende was identified very locally, but it is always rimmed by actinolite. (The metamorphic age of the hornblende is not known; it might be pre-D₂.)

Although the anorthite content of the plagioclase is often difficult to determine because of the fine grain sizes and lack of twinning, all determinations indicate that it is albite.

TABLE 4-4: MINERAL ASSEMBLAGES IN MAFIC ROCKS

FACIES.	ASSEMBLAGE	PHASES
prehnite - pumpellyite	M.1	ch pc pu pr cc
lower	M.2 ac	ch pc mu qz cc ma ti
&	M.3 cz	ch pc mu qz cc ma ti
middle	M.4	ep. bit ch pc mu qz cc ma
greenschist	M.5 ac cz ep bit ch. pc mu† qz	cc† ti

Notes: Assemblages based on textural equilibrium observed in thin section.
 † = not in all thin sections in this assemblage. Plagioclase where determined is sodic, with less than ca. 15% anorthite. Greenish-brown colour of biotite indicates a relatively high ferric Fe/Ti ratio. Relict hornblende (age uncertain) present very locally in assemblage M.5.

TABLE 4-5: MINERAL ASSEMBLAGES IN ULTRAMAFIC ROCKS

ASSEMBLAGE	PHASES
U.1 se	mg cr
U.2 se	tc tr mg cb cr
U.3 se	tc mg ma
U.4	cht tc mg crt†

Notes: † = not in all thin sections in this assemblage. cb = matrix carbonate in serpentinite breccia.

Fig. 4.5

ACF diagrams for mineral assemblages M.1 to M.5 in mafic rocks. A = $\text{Al}_2\text{O}_3 + \text{Fe}_2\text{O}_3$, C = CaO, F = $\text{MgO} + \text{FeO} + \text{MnO}$; excess SiO_2 , H_2O and CO_2 . Mineral compositions taken from Winkler (1979). Triangle represents an average composition of tholeiitic basaltic-andesitic rocks (from Winkler 1979).

Other mafic rocks

Mafic rocks also occur in the Black Phyllite, such as subunits uTrBPc and uTrBPf, and in the Takla Group, such as subunits TrJTv and TrJTg. These rocks are mainly volcanoclastic but at least some are probably close to basalt in general composition. They either contain metamorphic minerals indicative of the low greenschist facies as in the Antler Formation, or they possess only subgreenschist facies alteration ("very low grade" of Winkler 1979) of primary pyroxenes and plagioclase (see Chapter 2.5). These two facies may be separated by the reaction:

(Fig. 4.3; Winkler 1979, Turner 1981). In the subgreenschist rocks, one assemblage (M.1: Fig. 4.5) diagnostic of the lower temperature part of the prehnite-pumpellyite facies was identified in the extreme south of the thesis area in subunit TrJTb at locality 324 (Map 4). Amygdules in this metabasalt contain prehnite, pumpellyite, albite and chlorite (no actinolite), which indicates a T of between about 250° and 300°C (Fig. 4.3; Winkler 1979). The presence of prehnite and pumpellyite versus laumontite, and the absence of glaucophane-crossite brackets P at between 300 and about 600 MPa (Winkler 1979). Winkler (1979) also noted that the preservation of these Ca-Al silicates proves that the fluid was very H₂O-rich.

MINERAL ASSEMBLAGES IN ULTRAMAFIC ROCKS

The spatial distribution of ultramafic rocks of the Antler Formation is not as wide as that of its mafic constituents, but they do extend

from the extreme northwest of the thesis area southeastwards to the Mt. Brew area. Like the mafic rocks, there is not much variation in the observed mineral assemblages, which are shown in Table 4-5 and in Map 4.

Assemblages U.1, 3 and 4 can be considered in the ultramafic system, $\text{MgO-SiO}_2\text{-H}_2\text{O-CO}_2$.

Serpentine in the form of lizardite or antigorite is stable over a large temperature range but only under very low X_{CO_2} conditions: the $\text{CO}_2/(\text{H}_2\text{O} + \text{CO}_2)$ ratio cannot exceed a few per cent (see Winkler 1979). Serpentine reacts under higher X_{CO_2} conditions to form, for example, talc, magnesite, and tremolite (if Ca is present). The alteration of serpentine to talc + magnesite is observed to some degree in virtually all ultramafic rocks in the thesis area, indicating that everywhere the fluid phase contained some CO_2 . In fact, talc + magnesite rocks (steatite) are exclusive of serpentinites almost everywhere except in the northwest of the area, indicating that X_{CO_2} was relatively high during metamorphism. Steatite is effectively meta-serpentine; a likely reaction involved in its production is:

(Winkler 1979). This reaction cannot be used as a temperature indicator because its temperature is dependent on X_{CO_2} at a given pressure, though it does occur below about 550 to 575°C because that is where forsterite is first produced in the presence of talc (somewhat beyond the first appearance of staurolite in pelites: Winkler 1979) and indeed no forsterite has been found in the thesis area.

It appears that X_{CO_2} increased to a moderate value within the ultramafic rocks throughout most of the thesis area in M_2 , and the generally decreasing ratio between serpentinite and steatite from the

northwest to the southeast probably reflects a concomitant increase in X_{CO_2} . As metamorphic grade increases in this direction also, this is evidence that the partial fluid pressure of CO_2 tended to increase with metamorphic grade in these rocks. Furthermore, this raises the question of whether the activity of H_2O in the metamorphic fluid was in fact less than 1 over the whole thesis area, instead of approximately equal to 1 as has been assumed in the compilation of the petrogenetic grid. An $a_{\text{H}_2\text{O}} < 1$ would lower the temperatures of the reactions in the grid. However, because carbonate rocks are very rare in the Snowshoe Group and the Black Phyllite, it is inferred that there was generally only a very low concentration of CO_2 present in the fluid throughout metamorphism - not much would have been needed to promote the breakdown of serpentine.

The presence of tremolite in assemblage U.2 needs a comment, because normally tremolite does not form in Ca-bearing ultramafic rocks until very high temperatures, after the appearance of forsterite and diopside (Winkler 1979). Assemblage U.2 (note no quartz) occurs in a sedimentary serpentinite breccia (see Chapter 2.3) with much authigenic calcium carbonate, and so should be considered in the system $\text{CaO-MgO-SiO}_2\text{-H}_2\text{O-CO}_2$, or that of siliceous dolomitic limestones. Tremolite can form at much lower temperatures in this system, for example from the reaction:

This reaction is bivariant, but because the partial preservation of serpentine constrains X_{CO_2} to be less than 10%, the temperature range of the reaction at 500 MPa can be tentatively bracketed between about 460° and 500°C (see Winkler 1979, his Fig. 9-5; Turner 1981, p.168).

P-T CONDITIONS OF METAMORPHISM - DISCUSSION AND CONCLUSION

A range of M₂ metamorphic grade is present in the thesis area. This 'field gradient' (Turner 1981) is highest in the east and decreases to the northwest, and particularly to the southwest (Fig. 4.1, Map 4). The P-T position of the field gradient is estimated in the petrogenetic grid (Fig. 4.3). Only a general 'area' of P-T space for the field gradient can be inferred, rather than a well-defined erosion surface line. The inferred pressure of 500 ± 50 MPa is based on the conclusions of Getsinger (1985), mainly, and also Montgomery (1985) and Phillipone (1985). The pressure at the low T end of the field gradient can be constrained only between 300 and 600 MPa (from prehnite + pumpellyite - see above). A small convex curvature at this end is shown in Fig. 4.3: this is schematic in that it emulates the common shape of metamorphic field gradients, which probably reflects some degree of post-metamorphic differential uplift.

The lowest temperature of the field gradient is indicated in the extreme south where prehnite + pumpellyite indicates minimum temperatures of about 250°C. Metamorphic grade increases northeastwards into the lower greenschist facies with the formation of tremolite-actinolite + clinozoisite-epidote, indicating a minimum T of about 350°C. Most of the thesis area is in the lower and middle greenschist facies (approximately 350° to 450°C), indicated by the chlorite and tentative biotite zones in pelites (assemblages P.1 and 2), and in mafic rocks (assemblages M.2 to 5) by the presence of actinolite instead of hornblende, and clinozoisite-epidote + albite instead of more calcic plagioclase (hornblende + oligoclase would appear in approximately the upper greenschist facies: Winkler 1979). The evidence of ultramafic

rocks interspersed among these lower and middle greenschist facies assemblages is consistent: the ultramafic assemblages (although some are probably not in equilibrium) indicate temperatures of between 400° and 500°C.

The upper limit of the lower and middle greenschist part of the area is taken to be the almandine garnet first-appearance isograd, estimated at 450°C. Although this isograd may be composition-dependent, at least within the Snowshoe Group - to which it is almost confined - it is probably a reliable indicator of the lower limit of the upper greenschist facies.

Lower amphibolite facies conditions were reached in one locality, indicated by the staurolite-bearing assemblage P.8, which suggests a peak M_2 temperature for the area of approximately 500° to 525°C.

Thus, the metamorphic field gradient inferred from mineral assemblages in pelitic, mafic and ultramafic rocks ranges from about 400°C in the northwest to about 525°C in the southeast, and to about 250°C in the southwest, at a general pressure of about 500 ± 50 MPa. The southwest-northeast gradient is quite steep, and its tectonic significance is discussed in Chapter 6.2.

A NOTE ON RETROGRADE METAMORPHISM

The retrogression observed in many pelitic rocks in the thesis area, described in Chapter 3.4, post-dates the M_2 peak and probably pre-dates M_3 . Thus, it is part of the metamorphism continuum, but since it does not correspond to a recognized deformation phase, it is treated

separately here.

The only point to be made is that the grade of the retrogression was probably in the chlorite zone of the lower greenschist facies, or perhaps lower, since both garnet and biotite broke down to form chlorite. This implies a rapid drop in metamorphic grade during and immediately after D_2 , and this is discussed in Chapter 6.2.

M_1 AND M_3 METAMORPHIC CONDITIONS

The limited information available for M_1 (D_1) metamorphic conditions indicates that they were about the same as for M_2 (D_2), namely lower to upper greenschist facies (see Chapter 3.2), or roughly 350 to 500°C (see also Chapter 6.2, sub-section 14).

Metamorphic grade decreased during and after D_2 , into the realm of lower greenschist and subgreenschist conditions during D_3 (see above note on retrogression). This may have been accompanied by a decrease in P due to erosion (see Chapter 6.2).

[4] 3. ISOGRAD PATTERNS AND FACIES MAP

In the metamorphic map of the thesis area (Map 4, Fig. 4.1) the garnet first-appearance isograd is plotted, and also a tentative biotite zone (see Section 2). Because only one occurrence of staurolite is known, the staurolite isograd or the lower limit of the amphibolite facies could not be delineated beyond that locality. The prehnite-pumpellyite - greenschist facies boundary is constrained in the south but not in the

northwest, where its projection is tentative. That is the extent of the effective control on the metamorphic pattern in the thesis area.

The control on the metamorphic pattern improves farther northeast and southeast in the Quesnel Lake region. Isograds have been mapped for garnet, staurolite-kyanite and sillimanite by Campbell (1971), Campbell (1978), Getsinger (1985) and Fillipone (1985) (Fig. 4.2). For about 100 km along the length of the Intermontane-Omineca boundary, these isograds are roughly concordant with each other and with major structures which are outlined by the Antler Formation. The garnet isograd, and locally the staurolite-kyanite isograd, do transect the boundary (Fig. 4.2), but only in synformal folds (the author's F_2 , and inferred F_2 -equivalent folds elsewhere). The metamorphic pattern farther northeast of the boundary will not be discussed here, except to point out that the repetition of metamorphic zones across the North Arm of Quesnel Lake was attributed by Campbell & Campbell (1970) to an east-side down fault, and by Getsinger (1985) to displacement on the Little River fault.

Note also that close to the terrane boundary the isograds are very close together, implying steep lateral thermal gradients. In the extreme southeast of the thesis area, for example, the garnet isograd is only about 500 m north of the terrane boundary, and the tentative biotite zone appears to pinch out or become very thin, while within 1 or 2 km southwest of the boundary, Black Phyllite pelites are very fine grained and friable, and are possibly at only subgreenschist grade. This apparent attenuation of metamorphic zones is probably related to the other phenomena mentioned so far: the S_2 - M_2 quenching diachroneity, and the concordance of the isograds with the terrane boundary. These matters are synthesized and discussed in Chapter 6.2 (sub-sections 25 and 26).

CHAPTER FIVE

PALEOMAGNETISM

[5] 1. INTRODUCTION

A paleomagnetic study was made of Takla Group rocks from just south and west of the thesis area in an attempt to discover the paleopole of this locality, and to compare it with that of cratonic North America for the same time. Fig. 5.1 shows the tectonic setting of the Quesnel Lake locality, and of several other paleomagnetic studies which are discussed later in this chapter.

This work was done under the supervision of Dr. E. Irving at the Pacific Geoscience Centre, Sidney, B.C., who suggested this study. Irving has been prominent in paleomagnetic investigations in the Canadian Cordillera, mainly in geological terranes in western and central British Columbia, contributing to the evidence that these terranes were formed at latitudes farther south than the presently adjacent craton was situated at those times. Less is known about the Quesnel terrane, hence the interest in the Quesnel Lake locality, results from which could add to our knowledge of the tectonic history of the Cordillera.

Fig. 5.1

Tectonostratigraphic terranes of the southern Canadian Cordillera, based on Monger (1984). NRMT and SRMT, Northern and Southern Rocky Mountain Trench. Paleomagnetic localities: AX = Axelgold intrusion, CM = Copper Mountain intrusions, CS = Captain Cove-Stephens Island plutons, GB = Gyichon batholith, HT = Hazelton-Takla volcanics, KY = Karmutsen Formation, MS = Mount Stuart batholith, QL = Quesnel Lake, SP = Spuzzum-Porteau plutons, TP = Topley intrusions.

Part of the results of this study have already been published (Rees et al. 1985).

[5] 2. GENERAL GEOLOGY AND SAMPLE COLLECTION

Most of the sampling was done outside the area mapped in this thesis to minimize potential problems due to structural overturning and metamorphism. Samples were collected at exposures (called 'sites') of the Takla Group in the Horsefly area of Morton (1976) and the Morehead Lake area of Bailey (1978) to the south and west of the thesis area, respectively (Fig. 5.2). These rocks are very similar to those in the thesis area, and it is important to emphasize that there is no evidence for any major lithological or structural break between them. To illustrate this, the schematic cross-section of Bailey (1978) shows large, faulted folds which are overturned to the southwest (Fig. 5.3). His interpretation of the structure broadly conforms with the major D_2 southwesterly-verging structures mapped by the author. Combined with the evident stratigraphic continuity, it is clear that after the Quesnel terrane was emplaced (D_1) and folded (D_2) there has been no further relative movement or disruption at this tectonic boundary. (See also Price et al. 1985, page 3-38, on the "congruency" of structures at the Intermontane-Omineca Belt boundary.)

Small-scale faulting and tilting is ubiquitous in the Horsefly and Morehead Lake areas, so bedding attitudes cannot be reliably extrapolated between exposures. Therefore, sampling was limited to 12 sites where the paleohorizontal could be independently and accurately determined, namely bedded volcanoclastic rocks. The number of suitable

Fig. 5.2

Simplified geological map of the Quesnel Lake locality, adapted from Morton (1976) and Bailey (1978). BS = Bootjack Stock. The Shiko Lake Stock is immediately west of site 9. SW-NE = line of cross-section shown in Fig. 5.3.

The locations of the sampling sites (latitude N, longitude W) are as follows: site 1 = 52°35', 121°33', site 5 = 52°37', 121°40', site 6 = 52°36', 121°47', site 7 = 52°26', 121°18', site 8 = 52°26', 121°24', site 9 = 52°27', 121°29', site 10 = 52°21', 121°15', site 11 = 52°30', 121°42', site 12 = 52°39', 121°43'.

Fig. 5.3

Cross-section SW-NE (see Fig. 5.2) from Bailey (1978) above, and generalized cross-section from this thesis, below, which corresponds to the northeastern half of the upper section. Random dashes = Jurassic intrusion, black = mid-Cretaceous Bootjack Stock. That part of Bailey's section in which the paleomagnetic sampling was done is indicated.

Bailey chose to interpret the main structures as rotated fault blocks, but their general form conforms to the F2 folds in the thesis section. The point is that Bailey's area and the thesis area are structurally and stratigraphically continuous and there is no evidence of a tectonic break between them.

sites was further limited by poor exposure, and the weathered or fractured condition of many of the rocks.

One collection (site 6) was made in reddish-brown to grey, coarse to pebbly volcanoclastic sandstone, consisting of lithic and crystal detritus in a pervasive hematitic cement. The remaining collections were made in green to grey, interbedded volcanic wackes, siltstones and poorly sorted volcanic mudflow deposits. Typically, beds range from a few millimetres to a metre or so in thickness, and way-up indications such as graded bedding and truncations are usually present. Angular to sub-rounded volcanic rock and crystal fragments, typically pyroxene and plagioclase, range from sand to pebble size, in a matrix of finer rock and mineral detritus and clay. Quartz is notably uncommon. Some units are quite massive and very poorly sorted, containing subangular pebbles or cobbles of augite porphyry flow, and probably represent volcanic mud flows. Some clasts are reddish and oxidized, but most of these rocks are not.

The volcanic and sedimentary rocks in the area were intruded first by mid- to late Early Jurassic olivine gabbro, monzonite and syenodiorite, and later by small intermediate stocks and dykes of Early to mid-Cretaceous age, such as the Bootjack Stock nepheline syenite.

Bedding dips range from 22 to 70°, but there is no evidence of overturned bedding (also Morton, pers. comm. 1984) except in sites 2 and 3 which are in the thesis area. The age of the rocks sampled is quite well constrained by fossils recorded by Bailey (1978): sites 1, 11 and 12 are approximately Norian, sites 5, 6, 7, 8, and 9 are approximately Hettangian, and site 10 could be either. The total age range indicated here is about 20 Ma, between about 225 and 206 Ma (Harland et al. 1982).

[5] 3. ANALYTICAL PRINCIPLES AND METHOD OF STUDY

Rock magnetization in orogenic belts is generally complex, in that it typically consists of two or more magnetizations, the observed or measured magnetization being the vector sum of these. The problem is how to resolve the observed magnetization into its constituents, each of which must be evaluated separately. The procedure used is the progressive demagnetization of rock specimens in a series of steps, re-measuring the specimens after each step. Ideally, two remanent magnetizations acquired at different times would have stability ranges (resistance to heat or alternating magnetic fields) that do not overlap, so that first one and then the other is progressively removed or 'randomized' during the demagnetization treatment (Fig. 5.4a). The direction of their resultant, the observed magnetization, will migrate systematically during this treatment. Such behaviour is displayed in stereographic and graphical projections. The two magnetizations and their resultant must lie in a single plane (Fig. 5.4b). Hence, on a stereogram the migration of the resultant traces an arc of a great circle. The migration ceases at some point in the treatment, and the directions are stable, when only the higher stability magnetization remains; further treatment merely reduces its intensity without changing its direction. This is called a stable endpoint, and the magnetization of higher stability has been successfully isolated (McElhinny & Gough 1963).

The direction of the lower stability magnetization that is being gradually removed during the early stages of demagnetization is obtained by vector subtraction. The range of demagnetization over which this calculation should be made is determined on orthogonal plots or Zijderveld diagrams (Zijderveld 1967). The observed magnetization

Fig. 5.4

(a) Stability spectra of two superimposed magnetizations A, B. $dM/d\tilde{H}, T$ is the slope of the intensity versus treatment curve, and is thus a measure of the amount of magnetization being removed at a given $\tilde{H}$ or T. At point p, A has been randomized while B remains unaffected. (b) A, B and their resultants, NRM through M4, lie on a single plane, which corresponds to a great circle on a stereogram. Successive removed magnetization vectors are parallel to A until p, when only B remains to be removed. The RM after p and the resultant are the same, B, because only one vector component is present. (c) Fig. 5.3b can be put into a cartesian frame with point O at the origin, in which the coordinates of the resultant magnetizations are expressed in terms of the three axes. The decay of A (the trajectory of the resultants to point p) is a straight line that does not converge on O, but the decay of B does. These linear segments are apparent on any two of the three possible orthogonal plots. (d) Partial overlap of stability spectra of two magnetizations C, D. (e) C is preferentially removed at lower demagnetization levels up to point q, so the RM is virtually parallel to C. Thereafter, C and D decay simultaneously producing a curved resultant trajectory (it could be straight if C and D are removed in proportion). After r the RM is unidirectional and parallel to D. In this case, vector subtraction must be used with care; a more complete overlap of two magnetizations precludes this method of resolving them.

References: Dunlop (1979), Halls (1976).

intensity is expressed in terms of three cartesian coordinates (which may be normalized to the initial, untreated intensity), and any two of these three are used in orthogonal plots. If only one of the two constituent magnetizations is removed over a series of steps, the two chosen coordinates must diminish in a linear manner (cf. Fig. 5.4c). Thus, the removal of the magnetization of lower stability is represented, on these diagrams as a straight line that does not converge on the origin. It is legitimate to carry out vector subtraction for the range of this linear segment. Thereafter, the magnetization of higher stability diminishes in intensity along a different line which does converge on the origin. The latter decay line corresponds to the stable endpoint. Decay lines may also be inferred from numerical printouts when the removed vector is found to be approximately constant over a series of demagnetization steps.

More commonly, the coercivities (a measure of the resistance of a magnetization to an applied field of opposite sign) or the unblocking temperatures (the temperature at which thermal agitation destroys the alignment of magnetic domains) of two magnetizations overlap to some extent, so that both of them decay over the same demagnetization range (Fig. 5.4d). The resultant direction changes during treatment follow an irregular course, and decay lines on orthogonal plots are curved or jagged (Fig. 5.4e). This is especially likely if more than two magnetizations are involved. In such cases, not all magnetizations are resolvable, but often at least one can be identified by the presence of a straight line of remanence decay over a part of the demagnetization range, or by a roughly defined endpoint.

Instrumentation

Demagnetization treatment was done by the alternating field (AF) method and by the thermal method. An AC tumbling-specimen demagnetizer (Schonstedt model GSD-5) was used, which is capable of inducing alternating fields up to an intensity of 100 millitesla ($1 \text{ mT} = 10$ oersteds). Thermal demagnetization was done in a furnace (Schonstedt model TSD-1); specimens were heated for 15 minutes at each temperature and then cooled in zero magnetic field. After each cleaning step, the specimens were measured in a fluxgate spinner magnetometer (Schonstedt model SSM-1) and the measurements recorded on tape and manually. The data was then read into the mainframe computer (UNIVAC) and processed by the program SPAN, from which all computations and required plots were obtained. When not being measured or treated, specimens were kept in a Schonstedt zero-field demagnetized shield.

The magnetometer measures specimens in six different positions, providing four independent determinations of the magnetization vector. The average magnetization vector of these four determinations is computed and expressed in cartesian and polar coordinates, together with a measure of the homogeneity between them known as Fisher's estimate of precision, k (k_g = precision in specimens). The resultant vector is then corrected for paleohorizontal (bedding) and for present horizontal (the field orientation of the specimen). These two reference frames help to distinguish between magnetizations acquired before and after deformation, respectively. (The paleohorizontal correction is done by removing the dip by rotating about the strike.)

Procedure

A total of 174 specimens was obtained from 83 cores, drilled in the

laboratory from oriented hand samples from 9 out of the 12 sites collected (Fig. 5.2). Two of the 12 sites (2 and 3, within the thesis area) were in overturned strata and were rejected because of structural correction problems. Another site (4) was rejected because the hand samples were inadequately oriented due to a collection error.

The natural remanent magnetization (NRM) was measured for each specimen. Those which eventually yielded resolvable magnetizations are plotted in Fig. 5.5. One, two or three companion pairs (from one, two or three cores) of pilot specimens were selected from each site in such a way as to represent the hand sample NRM scatter as far as possible, for that site. Detailed AF demagnetization in 10 mT steps from 10 mT to 100 mT, and thermal demagnetization in 100, 50, or 25°C steps up to 700°C was carried out on these pairs of specimens in order to determine the coercivity or temperature of stable endpoints, and the characteristic range of any identifiable line of remanence decay.

Using this information, specimens from most of the other cores were treated in an attempt to determine the general properties of each site. For efficiency, this was done usually by AF treatment at (10) 20, 30, 50, 70 and 90 (100) mT steps. A few others were done by detailed thermal demagnetization. In all, 93 or 53% of the 174 specimens, including 64 or 77% of the 83 cores, received some degree of demagnetization treatment and assessment.

[5] 4. RESULTS: DESCRIPTION OF MAGNETIZATIONS

The results are described by reference to the complete demagnetization behaviour of a selection of specimens. Each example

Fig. 5.5

NRM directions with respect to present horizontal for the 93 specimens which yielded resolvable magnetizations. Following convention, solid circles = downward inclinations, open circles = upward inclinations.

includes the direction change of the observed magnetization with treatment, the decay of intensity with treatment, and two orthogonal plots, all normalized to the NRM value.

Thermal demagnetization in site 10

Between 200 and 500°C, the measured magnetization of specimen 105B from site 10 migrates along a great circle (Fig. 5.6). The magnetization vector that was removed over this range was computed and found to be directed northeast and downward. Its decay in intensity is clearly linear on the orthogonal plots, although it does not converge exactly on the origin, indicating that a higher blocking temperature magnetization remains. This other magnetization becomes apparent on the stereogram at 650°C where the direction abruptly changes and stabilizes at east and upward (southeast and upward with respect to paleohorizontal). Demagnetization treatment stopped at 700°C, before the intensity was reduced to zero, so the orthogonal plots do not record the final convergence to the origin.

The graph of normalized intensity versus treatment shows a steady decay slope, representing the removal of the northeast, downward magnetization. This kind of decay is characteristic of viscous partial thermoremanent magnetization (see Section 6).

Thermal demagnetization in site 8

From site 8, specimen 82B records at least two magnetizations which are clearly distinguished on the orthogonal plots by two linear segments (Fig. 5.7). There is an approximately linear decay between 100 and 500°C which is marked by a great circle migration on the stereogram.

Fig. 5.6

Demagnetization of specimen 105B. (a) Equal-area stereogram of direction changes. RM is removed magnetization vector, with range. (b) and (c) Orthogonal plots, with respect to paleohorizontal (bedding). (d) Change in intensity with treatment, normalized to the NRM value.

Fig. 5.7

Demagnetization of specimen 82B. (a) Equal-area stereogram of direction changes. RM is removed magnetization vector, with range. (b) and (c) Orthogonal plots, with respect to paleohorizontal (bedding). (d) Change in intensity with treatment, normalized to the NRM value.

The magnetization removed (RM) over this range is again northeast and downward. A higher temperature magnetization, present between 500 and 600°C at least, is apparent on the orthogonal plots where the decay makes a sharp turn towards the origin along another line, although it does not reach it. The direction of this magnetization migrates along a different great circle, becoming easterly and upward with respect to paleohorizontal. The fact that the observed magnetization direction does not stabilize after the removal of the lower temperature magnetization, and that there is no final convergence to the origin on the orthogonal plots, suggests that there is still an unresolved magnetization present at 700°C.

AF demagnetization in site 10

Specimen 105A was treated by the alternating field method. Although the direction change with treatment is not as clear as with the thermal companion specimen 105B, the observed magnetization migrates from the initial NRM direction along a roughly defined great circle to a southeasterly upward direction, with respect to present as well as paleohorizontal (Fig. 5.8). The removed magnetization was calculated for the 30-50 mT range, and is northeast and downward with respect to present horizontal. The intensity decreases rapidly with treatment, so the decay line appears to converge on the origin because the divergence at higher levels of treatment is too small to be appreciated.

AF demagnetization in site 9

Specimen 97A from site 9 (AF treatment) is an example of two magnetizations with familiar directions, but in this case one is difficult to define. A northeasterly downward direction persists until

Fig. 5.8

Demagnetization of specimen 105A. (a) Equal-area stereogram of direction changes. RM is removed magnetization vector, with range. (b) and (c) Orthogonal plots, with respect to paleohorizontal (bedding). (d) Change in intensity with treatment, normalized to the NRM value.

50 mT when enough intensity is removed to shift the resultant magnetization to scattered, southerly upward directions with respect to paleohorizontal (Fig. 5.9). The 30-50 mT vector is thus a good measure of the removed magnetization. The higher stability magnetization(s) does not stabilize, nor does it decay to the origin on the orthogonal plots. In fact, its intensity increases with treatment (Fig. 5.9), possibly due to the removal of an antiparallel component.

None of the analysed specimens from this site yielded a stable endpoint.

Demagnetization in site 7

All specimens from site 7 demagnetized by AF show minimal direction change with treatment. The sole magnetization recorded and removed in, for example, specimen 74A is steeply inclined downward to the north, between the present earth's field (PEF) direction and the Recent geocentric axial dipole (GAD) field direction (Fig. 5.10). The coercive force of this magnetization exceeds 100 mT, although the k_s values and the relative intensity are much reduced. On the orthogonal plots, there is a single, approximately linear decay towards the origin. However, at least one thermal specimen from this site, 74B, records on the stereogram the removal of this magnetization (not illustrated): the direction migrates along a wide great circle arc from north and downward to southeast and upward at 650°C, with respect to paleohorizontal.

Demagnetization in site 5

Site 5 is the only site with strong coherence (consistency) of stable endpoints. Fig. 5.11b shows the initial wide scatter of the NRM

Fig. 5.9 Demagnetization of specimen 97A. (a) Equal-area stereogram of direction changes. RM is removed magnetization vector, with range. (b) and (c) Orthogonal plots, with respect to paleohorizontal (bedding). (d) Change in intensity with treatment, normalized to the NRM value.

Fig. 5.10

Demagnetization of specimen 74A. (a) Equal-area stereonet of direction changes. RM is removed magnetization vector, with range. (b) and (c) Orthogonal plots, with respect to paleohorizontal (bedding). (d) Change in intensity with treatment, normalized to the NRM value.

Fig. 5.11

(a) Site 5 stable endpoints at 675°C or 70 mT, with respect to paleohorizontal (bedding) (all upward directed). $n = 19$. (b) Initial scatter of NRM directions.

of all 19 specimens; after 70 mT or 675°C of demagnetization, all the specimens reach stable endpoints, clustered in the northeast and upward, with respect to paleohorizontal (Fig. 5.11a). The precision k_g is in the order of 10^1 , less commonly 10^2 . However, the orthogonal plots show that the corresponding linear segment does not converge on the origin, indicating that harder magnetizations are present, although the cluster begins to dissipate on further demagnetization, suggesting that they are not systematic.

As an example, the observed magnetization of specimen 57A migrates from southeasterly and downward to northeast and upward, with respect to paleohorizontal (Fig. 5.12b-e). Unlike several other sites, the removed magnetizations are not systematic (Fig. 5.12a).

[5] 5. ANALYSIS OF RESULTS: RESOLUTION OF MAGNETIZATIONS

As an introductory statement, the magnetizations observed in this study are typical of orogenic belts in that all sites possess two or more magnetizations, and their resolution is not straightforward (cf. Section 3). Nevertheless, four magnetizations have been isolated from the Quésnel Lake locality, based on their presence in at least 5 out of the 9 sites, or by a high degree of coherence, or on circumstantial grounds. These magnetizations, called W, X, Y and Z, are listed by site in Table 5-1¹, and summarized in Table 5-2².

The paleopoles in Table 5-2 are obtained from the mean declination and inclination, applying the dipole formula $\tan I = 2 \tan \lambda$, where I is the mean inclination and λ is the paleolatitude.

Fig. 5.12

(a) Removed magnetization vectors (usually 0-50 mT) from site 5. (b) to (e) Demagnetization of specimen 57A. (b) Equal-area stereogram of direction changes. (c) and (d) Orthogonal plots, with respect to paleohorizontal (bedding). (e) Change in intensity with treatment, normalized to the NRM value.

TABLE 5-1: RESULTS: MAGNETIZATIONS BY SITES

SITE	MAG.	H(C)S	D _B [°] , I _B [°]	D _H [°] , I _H [°]	M(A m ⁻¹)	k _w	ag ₅ [°]	DEFINITION
1	*	3(3)5	S, d W, d	-	-	-	-	* No coherence of magnetizations among cores. (S, south; W, west; d, down)
5	W	4(8)19	22, -53	-	0.29	72	11	Stable endpoints: 675°C, 70 mT No systematic removed vectors
6	Z?	1(2)3	n.d	n.d	n.d	n.d	n.d	More than one mag., but no coherent endpoints, nor good decay lines. Porous rock vulnerable to CRM
7	X	1(1)1	142, -3	-	0.112·10 ⁻¹	-	-	Endpoint: 600°C
	Z	3(6)6	-	6, 68	0.151·10 ⁻²	29	13	Stable directions. Endpoints: 100 mT
	X	5(10)12	131, -12	-	0.246·10 ²	4	27	Stable endpoints: 650°C, 50, 70 or 90 mT
8	Y	3(4)4	-	23, 43	0.155·10 ²	16	24	Removed vectors: 100-500°C, 300-550°C, 30-50 mT, 10-30 mT
9	Y	3(4)5	-	12, 59	0.636·10 ⁻²	13	22	Removed vectors: 200-500°C, 30-50 mT
	X	2(4)5	139, -43	-	0.157·10 ¹	12	23	Endpoints stable at 650°C, and between 50 and 100 mT
10	Y	3(5)6	-	38, 53	0.156·10 ¹	32	12	Removed vectors: 200-500°C, 20 or 30-50 mT, 30-90 mT.
	X	3(4)5	128, -29	-	0.602	2	85	Endpoints at 650°C, low to moderate AP
11	Y	2(4)5	-	63, 61	0.653	8	28	Removed vectors: 200-500°C, 40-30 mT, 30-50 mT
	X	2(2)3	136, -39	-	0.236	24	26	Endpoints at 650°C, 70 to 90 mT ^o
12	Y	2(2)2	-	62, 71	0.148	67	31	Removed vectors: 300-600°C, 200-500°C

Notes: H(C)S = hand samples (cores) specimens, D_B[°], I_B[°], D_H[°], I_H[°] = declination (D) & inclination (I) with respect to bedding (B) & present horizontal (H), M(A m⁻¹) = NRM intensity in amperes per metre, k_w = Fisher's estimate of precision by site, ag₅[°] = radius of circle of confidence, n.d = not determined.

TABLE 5-2: SUMMARY OF RESULTS

Magnetization	Weighting	PALEOFIELD					PALEOPOLE	
		datum	D°, I°	k _b	ag5°	lat., long.	dm° / dp°	
W	1 site (4 samples)	B	22.5, -53.5	72*	10.9	1.2°N, 39.9°E or 1.2°S, 140.1°W	15.2 / 10.6	
	19 specimens	B	20.2, -53.2	52*	4.7	2.0°N, 41.6°E or 2.0°S, 138.4°W	6.5 / 4.5	
	5 sites	B	135.2, -25.2	20	17.3	37°N, 117.9°E or 37°S, 62.1°W	18.7 / 10	
X	26 specimens	B	133.2, -24.7	4	16.7	35.8°N, 119.6°E or 35.8°S, 60.4°W	17.9 / 9.6	
	5 sites	H	36.2, 58.9	27	14.9	62.2°N, 18.7°W or 62.2°S, 161.3°E	22.2 / 16.6	
	22 specimens	H	34.9, 57.4	12	9.2	61.9°N, 14.6°W or 61.9°S, 175.4°E	13.4 / 9.8	
Y	22 specimens	B	38.2, 53.4	2	25	57.2°N, 12.5°W or 57.2°S, 167.5°E	35.4 / 24.7	
	6 specimens	H	6.2, 67.8	29*	12.7	85.8°N, 10.8°E or 85.8°S, 169.2°W	21.2 / 17.7	

Notes: D°, I° = declination & inclination, k_b = Fisher's estimate of precision between sites, for X & Y, with unit weight given to sites or specimens. * = within-site precision (k_w) for W & Z, with unit weight given to samples or specimens. ag5° = radius of circle of confidence (P = 0.05) for sample, specimen or site mean. B = bedding, H = horizontal.

W MAGNETIZATION

The W magnetization is the best defined, and is represented by stable endpoints directed northeast and upward ($D = 22.5^\circ$, $I = -53.5^\circ$), with respect to paleohorizontal (Fig. 5.11a), but it occurs only in site 5 (see Section 4). However, the strong coherence ($k_w = 72$, where k_w is Fisher's estimate of precision within the site) and the high stability fields (675°C , 70 mT plus) of all the stable endpoints suggest that this magnetization is indeed significant.

None of the other magnetizations (X,Y,Z) appear to be present in this site in any systematic way.

RESOLUTION OF OTHER MAGNETIZATIONS

In all the remaining sites, endpoints were inferred from the

¹The a_{95} value referred to in Tables 5-1 and 5-2 and succeeding figures is the half-angle of the cone of confidence around the calculated mean direction, such that there is a 95% probability ($P = 0.05$) that the true mean lies within the projected small circle. a_{95} is related to k :

$$a_{95} = 140/(kN)^{1/2}$$

where N is the sample number.

²The a_{95} circle of confidence, related to errors in directions (dD° , dI°), is converted to an ellipse of 95% confidence when dealing with errors around the corresponding paleopole because dD varies with paleocolatitude, p , which is a function of I .

$$dD = a_{95}/\cos I \quad (dI \text{ is simply } a_{95})$$

The axes of the ellipse are

$$dP = 1/2 a_{95}(1 + 3\cos^2 p) \quad (\text{axis parallel to paleomeridian})$$

$$dM = a_{95} \sin p / \cos I \quad (\text{axis perpendicular to } dP)$$

tabulated or plotted results, where the direction of the apparently hardest (highest stability) magnetization was approximately stable over a series of steps. An average or typical direction of these steps was taken. Other criteria for endpoints were an intensity at least 2% of the initial NRM, and a k_g value of at least 2. Below these values, the directions are unreliable because of noise, or because the magnetization has been essentially randomized.

No coherent grouping of endpoints was recognized in sites 1, 6 or 9.

X MAGNETIZATION

In sites 7, 8, 10, 11 and 12, most or all of the specimens possessing stable endpoints have those endpoints directed east, south or southeast and upward, with respect to paleohorizontal (Fig. 5.13; see Section 4 for examples). No other grouping was common to these five sites. The majority of these specimens have k_g values at the endpoint in the order of 10^1 , with a few in the order of 10^2 or less than 10, but their directions are poorly grouped. The within-site precision k_w ranges from 24 to 4 (Table 5-1), and the combined specimen distribution has a precision (k_b) of only 4 (Table 5-2). However, the precision on the basis of the five site means is better ($k_b = 20$, Table 5-2), and it is concluded that this southeasterly upward grouping is a remanent magnetization, here called X. The mean direction by sites is $D = 135.2^\circ$, $I = -25.2^\circ$; the specimens mean is virtually the same.

In thermal demagnetization, the stable endpoints representing the X magnetization become apparent at 650°C , suggesting that the carrier is hematite; the intensity is about 10% of the initial NRM. In AF

Fig. 5.13

(a) Stable endpoints from sites 7, 8, 10, 11 and 12, defining the X magnetization. $n = 26$. (b) Site means, and between-sites specimens mean (S) and sites mean (E). (c) Site means in present horizontal reference frame, showing greater dispersion.

demagnetization, the coercive force of the X magnetization ranges from 50 to 100 mT, probably depending on mineralogy and grain size; the intensity is typically a few percent of the initial NRM.

Y MAGNETIZATION

The Y magnetization was isolated on the following basis. Graphical and tabulated results reveal that linear decay lines, corresponding to great circle migrations, are present in 22 specimens from sites 8, 9, 10, 11 and 12 (see Section 4). The range of these decay lines is typically 30 to 50 mT or 200 to 500°C, with some variation (see Table 5-1). Vector subtraction over this range yields directions that are generally northeast and downward, with respect to present horizontal (Fig. 5.14b). Other directions occur, but were not considered further if their k_g was less than 2, if they were unique or the same as their eventual endpoint (where that endpoint is not northeast and downward), or if the decay is clearly the removal of the Recent earth's field magnetization.

This northeast, downward removed magnetization is designated Y, and is the only one common to these five sites (Table 5-1). In all of them except site 9, it represents the magnetization that was removed during treatment after the removal of soft magnetizations, and prior to the stabilization of other magnetizations (usually X) at higher levels of treatment. Specimens 105A and B, and 82B described in Section 4 are good examples. In site 9 the Y magnetization appears to be the hardest identifiable magnetization: higher levels of demagnetization produce scattered endpoints and low k_g values, rather than any good stable endpoints (see 97A in Section 4).

Fig. 5.14

γ magnetization. (a) Site means, specimen mean and sites mean. Note that the PEF and GAD field directions both lie outside the 95% circles for the γ magnetization. (b) Individual specimens, $n = 22$.

It is emphasized that none of these removed magnetizations is antiparallel, namely southwest and upward, such as might be produced during a field reversal.

With respect to present horizontal, the Y magnetization has a Fisher's precision (k_b) of 27 where sites are given equal weight, and 12 giving specimens equal weight (Table 5-2). The sites mean ($D = 36.2^\circ$, $I = 58.9^\circ$) and combined specimens mean ($D = 34.9^\circ$, $I = 57.4^\circ$) are virtually coincident (Fig. 5.14a). The dispersion of the specimens greatly increases in the paleohorizontal reference frame, k_b dropping to 2.46 (Fig. 5.15a); the mean direction is close to that for present horizontal, but this is fortuitous and the errors are large.

Z-MAGNETIZATION

The steep northerly downward magnetization that dominates AF specimens throughout treatment in site 7 (see Section 4) is named the Z magnetization (Fig. 5.15b, Tables 5-1 and 5-2). It may be present in the odd specimen in other sites (e.g. site 6 and site 12) but there it is always removed by 30 mT or 200°C of demagnetization.

[5] 6. ORIGIN OF MAGNETIZATIONS

W AND X MAGNETIZATIONS

The W and X magnetizations are tentatively interpreted to be approximately primary (pre-deformation) because of their generally high remanent coercive force and unblocking temperatures, and because their

Fig. 5.15

(a) Y magnetization specimens plotted in paleohorizontal (bedding) reference frame, showing greater dispersion than in Fig. 5.14(b). (b) Z magnetization, from site 7. $n = 6$.

site and specimen grouping is better after correcting for Middle Jurassic deformation. The magnetizations could have been acquired during diagenesis of the sediments in the Hettangian, or more likely, slightly later by thermal magnetization related to the intrusion and cooling of late Early Jurassic plutons.

Y MAGNETIZATION

The Y magnetization is interpreted to be a secondary remanent magnetization which overprinted primary magnetizations to varying degrees. Because the Y grouping is considerably better with respect to present horizontal compared to paleohorizontal, it was probably acquired after most or all of the folding in the Quesnel Lake locality, namely after Middle to Late Jurassic deformation. Post-Y magnetization differential tilting between sites is unlikely because the precision by sites ($k_b = 27$) is quite good; also, the specimen dispersion in Fig. 5.14b reflects scatter within rather than between sites. The assumption is, then, that the present horizontal reference frame is a good approximation to the post-Jurassic paleohorizontal at the time the Y magnetization was acquired. The only possibly post-Jurassic bedded rocks in the sampling area to test this assumption (Bailey 1978, his unit 10), are separated from underlying, deformed Jurassic rocks by a marked angular unconformity, indicating that they post-date Jurassic deformation, uplift and erosion (Bailey 1978). These undeformed sandstones dip less than 10° , suggesting that post-Jurassic tilting was negligible (but the problem of the Y magnetization datum is not so easily dealt with: see Chapter 6.3). Post-Y magnetization uniform tilting on a

regional scale is possible, but unlikely to be large; the only large dips (5 to 20°) reported in post-Jurassic strata (excluding the aforementioned rocks) are those close to large faults, distant from the sampling locality (Campbell & Tipper 1971). The tilt question is addressed further in Chapter 6.3.

The main question for now is how to explain and date the Y magnetization. Thermal specimens bearing the Y magnetization show a steady reduction in normalized intensity (e.g. Fig. 5.6d) rather than a sharp 'elbow' at a particular unblocking temperature. This is typical of the decay of a viscous partial thermoremanent magnetization (PTRM; Irving & Opdyke 1965, Irving 1964 and pers. comm. 1984) which is characterized by a high unblocking temperature spectrum but produced by sustained heating at low to moderate temperatures, and subsequent cooling in a magnetic field. In contrast, low or near-surface temperature, viscous-remnant magnetization (VRM), capable of continually adjusting to the ambient earth's field, generally does not survive demagnetization above about 20 mT and 300°C. The higher stability of Y, and the observation that the PEF and GAD field directions both lie outside the 95 confidence circles for Y (Fig. 5.14a), effectively precludes a Recent VRM origin for Y, and instead suggests that it was acquired earlier, after the Middle to Late Jurassic and during an episode of regional heating and cooling which produced the partial thermoremanence.

The most likely cause of this partial thermoremanence is the mid-Cretaceous plutonism that is documented in the region. The only Cretaceous radiometric age determination from within the sampling locality is a K-Ar whole-rock age of 111 ± 4 Ma for the Bootjack Stock

nepheline syenite (Fig. 5.2; Bailey 1978). Morton (1976) and Bailey (1978) concluded that numerous other small stocks and dykes that intrude the area are also probably Cretaceous, from their geological relationships and closer similarities to known Cretaceous intrusions as opposed to the more alkalic Early Jurassic plutons. The Early Jurassic biotite-quartz diorite Takomkane Batholith 50 km to the southeast, has a K-Ar biotite age of 105 ± 4 Ma, attributed by White et al. (1968) and Soregaroli & Nelson (1976) to isotopic resetting related to the intrusion of the nearby Boss Mountain Stock. An altered andesite dyke and a biotite fragment in a vein breccia from the same place have K-Ar biotite ages of 98 ± 4 Ma and 104 ± 4 Ma, respectively. The east arm of the Quesnel Lake Gneiss, 50 km to the east of the Quesnel Lake locality also has a reset K-Ar biotite age of 108 ± 4 Ma. Within the Quesnel Lake sampling locality, near sites 8 and 9 (Fig. 5.2), the Early Jurassic Shiko Lake monzonite stock has a K-Ar biotite age of 154 ± 5 Ma (Schink 1974) which may indicate partial isotopic resetting. (In an analogous way, the Y magnetization is a partial magnetic overprint, not strong enough to obliterate the primary magnetizations, except perhaps in site 9 - see Section 5.)

The early Tertiary is known to be another period when the geothermal gradient was comparatively high in southeastern British Columbia, marked by isotopic resetting and magnetic overprinting. However, no physical evidence for such effects of Tertiary age have been found as far north and west as the Quesnel Lake locality (R. Parrish, Geological Survey of Canada, pers. comm. 1985).

Comparison with known mid-Cretaceous paleopoles

A test of whether or not the Y magnetization is a mid-Cretaceous PTRM overprint may be provided by a comparison with known mid-Cretaceous paleopoles. For the purpose of discussions that follow, the cratonic North American paleopole for the mid-Cretaceous that is used for reference here is that calculated by Irving et al. (1985), based on an average of six cratonic determinations with a mean age of 94 Ma. This cratonic reference paleopole, 67.7°N , 186.3°E ($k = 312$, $A_{95} = 3.8^{\circ}$) is known to be a paleo-North (geographic) Pole because the earth's field was virtually continuously NORMALLY polarized between 118 and 83 Ma (Harland et al. 1982), an interval known as the Cretaceous Normal Superchron. This reference paleopole is markedly different from the Y paleopole (61.9°N , 14.6°W ; Table 5-2), nor do they lie on a small circle centred on the Quesnel Lake locality (Fig. 5.16). Nor does the Y paleopole coincide with any part of the post-Middle Jurassic apparent polar wander (APW) path for the craton. However, it is clearly compatible with a group of (north) paleopoles derived from the mid-Cretaceous Spuzzum and Porteau plutons, and the Captain Cove and Stephens Island plutons of the Coast Plutonic Complex, and the 125 Ma Axelgold anorthosite-gabbro from the Cache Creek terrane (Fig. 5.1). The paleomagnetism of the Spuzzum and Porteau plutons was studied by Irving et al. (1985) and they considered that these magnetizations are primary and were acquired during the cooling of these intrusions within the interval 105 to 90 Ma, which is the same interval indicated by the plutonism and K-Ar resetting recorded within and near the Quesnel Lake locality. Like the Y magnetization, no reversals are recorded in the known mid-Cretaceous magnetizations (all directions are downwardly inclined), re-inforcing the interpretation that they were all formed

Fig. 5.16

Polar map projection showing the present positions of paleomagnetic localities in western North America, and their corresponding paleopoles and standard error circles ($P = 0.63$). Abbreviations as in Fig. 5.1. Also shown is the APW path for North America, including the paleopole (NA) for a mean age of 94 Ma.

Note: The standard error (a_{63}) is the half-angle of the cone of confidence around the calculated paleopole, such that there is a 63% probability ($P = 0.37$) that the true paleopole lies within the projected small circle. a_{63} is related to Fisher's precision, k

$$a_{63} = 81/(kN)^{1/2}$$

where N is the sample number.

around 100 Ma (± 15 Ma) during the Cretaceous Normal Superchron.

On this basis, it is tentatively concluded that the Y magnetization is an approximately mid-Cretaceous PTRM overprint. A fairly long cooling interval would prolong the acquisition of remanence, and this could partly explain the poorer within-site precision of Y compared to that of the sites mean.

Z MAGNETIZATION

The Z magnetization is probably a viscous or chemical remanent magnetization recording the Recent earth's field. The reason for its high stability in site 7 is not known. A possible explanation is that localized, low temperature weathering alteration affected the rocks at this site (E. Irving, pers. comm. 1985).

[5] 7. INTERPRETATION OF MAGNETIZATIONS

W AND X MAGNETIZATIONS

The paleopoles of the W and X magnetizations are listed in Table 5-2 and plotted in Fig. 5.17. They are quite different. If they are both primary, this means either that site 5 was not in its present relative position at the time of the formation of W and X, or much more likely, the W and X magnetizations were acquired at different times in the Early Jurassic, between which there was apparent polar wander.

There is more serious uncertainty about the significance of the W and X paleofields with respect to the cratonic reference frame. First, the ages of the magnetizations are not precisely known, although if they

Fig. 5.17

Polar map projection showing paleopoles for the W and X magnetizations and for several Late Triassic and Jurassic localities from British Columbia (general position marked by asterisk). See text. Solid circles = northern hemisphere, open circle = southern hemisphere.

Key: H1 = Hazelton Group, mid-Sinemurian, H2 = Hazelton Group, Toarcian, TK = Takla Group, Late Triassic (H1, H2, TK, from Monger & Irving, 1980), TP = Topley intrusives, 142 Ma (Symons, 1983b), NA = North America, 190 Ma.

are primary, they are both probably Hettangian or somewhat younger, roughly 205 to 184 Ma. Second, the Late Triassic-Early Jurassic was a period of frequent geomagnetic field reversal, so the sign of the paleolatitude cannot be determined. Third, the North American reference paleopole for this interval is currently under review because of revised radiometric dating of the reference localities and of the Triassic-Jurassic boundary (see May & Butler 1986).

Given these problems, and the rather poor resolution of the **X** magnetization, no relative paleolatitudinal displacement or rotation was computed for **X**. However, this was done for the **W** magnetization, assuming an age for the magnetization of 190 Ma which yields an expected cratonic paleopole of 66°N, 93°E (Irving & Irving 1982) (which may or may not prove to be realistic), and an expected paleolatitude for site 5 of $31.4 \pm 5^\circ$ N. The observed (north) paleopole for site 5 from the **W** magnetization is 1.2°N, 39.9°E, and the paleolatitude is 34.1°N or S $\pm 10.6^\circ$. Using the northern (paleolatitude) option, which requires a reversed field, the result indicates a statistically insignificant relative displacement of $2.7 \pm 11.7^\circ$ to the south, but a very large anticlockwise rotation of $141.8 \pm 14.1^\circ$ about a vertical axis relative to the paleomeridian. The southern option (normal field) requires a geologically more accountable clockwise relative rotation of $38.2 \pm 14.1^\circ$, but a huge latitudinal shift of $65.5 \pm 11.7^\circ$ northwards (about 7000 km), relative to the craton.

Fig. 5.17 shows the paleopoles of the **W** and **X** magnetizations in relation to the paleopoles from the Late Triassic-Early Jurassic Takla Group and Hazelton Groups (Monger & Irving 1980) and from the Topley intrusions (Symons 1983b) of the Stikine terrane. They are fairly

scattered, but some can be linked on small circles, implying relative rotations between these localities. While most of these paleopoles are generally discordant from the cratonic paleopole (taken at 190 Ma) by roughly 10-20°, the uncertainty of the reference paleopole and of the ages of the magnetizations represented here make more elaborate comparisons and interpretations difficult. At present, the significance of the W and X magnetizations cannot be resolved.

Y MAGNETIZATION

Table 5-3 lists the relative latitudinal displacements and rotations for the Y magnetization and for known Early to mid-Cretaceous magnetizations from British Columbia and northern Washington (adapted from Irving et al. 1985). All directions are down to the northeast. The corresponding paleopoles are plotted in Fig. 5.16. The Y magnetization from the Quesnel Lake locality will hereafter be referred to as QL.

Other magnetizations possibly equivalent to QL

Following Irving et al. (1985), other magnetizations (KY, CM, GB and GBB), are included in Table 5-3 and Fig. 5.16 because of the proximity of their paleopoles to the mid-Cretaceous group. KY is a heterogeneous, secondary remanent magnetization that was isolated from the Upper Triassic Karmutsen Formation of Vancouver Island (Wrangell terrane; Fig. 5.1), considered by Yule & Irving (1980) to be a chemical remanent magnetization, or a viscous PTRM at least partly of Cretaceous age.

CM, GB and GBB are magnetizations obtained from two Early Jurassic intrusions (both 198 Ma) from the southern part of the Quesnel terrane

TABLE 5-3: PALEOFIELDS, DISPLACEMENTS AND ROTATIONS OF MID-CRETACEOUS MAGNETIZATIONS

SYMBOL	LOCALITY (MAGNETIZATION)	D°, I°	A ₉₅ °	λ°	RPD°	RR°	REFERENCES
QL	Quesnel Lake (Y, *)	35, 57	9	38±10	23±10	-73±15	this thesis
SP	Spuzzum and Porteau plutons	30, 57	5	37±5	22±6	-66±8	Irving et al. 1985
AX1	Axelgold intrusion (layering datum)	25, 69	5	52±6	14±6	-66±12	Monger & Irving 1980, Irving et al. 1985
AX2	(horizontal datum)	32, 60	5	41±4	25±5	-75±10	
KY	Karmutsen Formation (Y overprint, *)	25, 57	10	38±8	23±9	-60±13	Yole & Irving 1980
CM	Copper Mountain batholith (* overprint?)	25, 45	5	26	ca. 35	n.d	this thesis, Symons & Litalien 1984
GB	Guichon batholith (* overprint?)	20, 51	11	31±7	28±8	-57±11	Symons 1971, Irving et al. 1985
GBB	as above	28, 36	7	20±5	ca. 41	n.d	Symons 1983a, inter- pretation, this thesis
MS	Mount Stuart batholith	10, 45	5	27±3	31±4	-42±6	Beck et al. 1981
CS	Captain Cove & Stephens Island plutons	18, 58	5	39±5	26±6	-56±10	Symons 1977

Notes: D°, I° = declination & inclination, A₉₅° = radius of circle of confidence (P = 0.05) of paleopole, λ° = paleolatitude, RPD° = relative paleolatitudinal displacement, RR° = relative rotation (negative = clockwise), * = presumed mid-Cretaceous, n.d = not determined.

(Fig. 5.1). The Copper Mountain intrusions (CM) paleopole was calculated by the author from the data of Symons & Litalien (1984), incorporating some sites he had deemed unreliable. The result (11°E , 60.2°N ; Table 5-3) is slightly removed from the mid-Cretaceous group, but it is included in Table 5-3. The Copper Mountain intrusions are indeed cut by the 102 ± 4 Ma Verde Creek quartz monzonite (Symons & Litalien 1984), although Early Jurassic biotites do not have their K-Ar ages reset (Preto et al. 1971). In any case, Symons attached no significance to the Cretaceous intrusion and presumed that the magnetization was primary and Early Jurassic.

GB and GBB are paleopoles obtained by Symons from the Guichon Batholith. Originally, Symons (1971) explained his result (GB) solely by an in situ rotation of about 40° . He re-examined his collection (Symons 1983a) in the light of Monger & Irving's (1980) contention that the Stikine and Cache Creek terranes had moved northwards relative to the craton by $13-14^{\circ}$ since the Late Cretaceous or early Tertiary. His later analysis (GBB) derived a mean inclination that was 14° shallower than his 1971 result, which in turn indicated a 14° more southerly paleolatitude, and enabled him to conclude that the Quesnel terrane had experienced the same translations as the Stikine and Cache Creek terranes (Symons 1983a,b). As with the Copper Mountain intrusions, Symons presumed that the Guichon Batholith magnetization was primary and Early Jurassic.

An alternative hypothesis to Symons (1971, 1983a, 1984) interpretations, and the reason that CM, GB and GBB are included in Table 5-3, is that these magnetizations are not primary and Early Jurassic, but Cretaceous overprints or partial overprints (Irving, pers.

comm. 1984) which Symons (1983a, 1984) erroneously compared to an Early Jurassic cratonic reference frame, and derived aberrant paleolatitudes which fortuitously agreed with the determinations of Monger & Irving (1980). Although the Copper Mountain and Guichon Batholith paleopoles are indeed distinctly different from those that Monger & Irving (1980) obtained from the Late Triassic-Early Jurassic Hazelton and Takla volcanics of the Stikine terrane, it should be pointed out that they do lie approximately on small circles, and so could be accounted for by in situ rotations; this was the explanation offered by Symons (1983a).

Paleolatitudes

Assuming an average age of 94 Ma for all the magnetizations in Table 5-3, the corresponding paleolatitudes can be projected on a paleogeographic map of North America derived from the cratonic reference paleopole for this time (Figs. 5.18, 5.19). The paleolatitude of QL is $38 \pm 9.8^\circ\text{N}$, compared with an expected paleolatitude of $61 \pm 3.8^\circ\text{N}$; the observed rotation is $34.9 \pm 12.5^\circ$ clockwise, compared with an expected rotation of $38.3 \pm 7.9^\circ$ anticlockwise. Paleolongitude is indeterminate. A relative latitudinal anomaly of $23 \pm 10.5^\circ$, and a relative clockwise rotation of $73.2 \pm 14.7^\circ$ are indicated by this result (Fig. 5.18). These anomalies are statistically identical to those indicated by several of the other results from British Columbia (Fig. 5.19), a consequence of the 'far-sided, right-handed' grouping of their paleopoles (Fig. 5.16). There is no northern-southern hemisphere ambiguity because the paleofield had normal polarity.

Tectonic implications

While uncertainties remain about the origin of some of these

Fig. 5.18

Paleogeographic map of North America for the mid-Cretaceous; present coastline projected for convenience. The expected declination for QL is 38° anti-clockwise of present north, while the observed declination is 35° clockwise (error is shown stippled). The total rotation is therefore the angle between the mid-Cretaceous meridian and the observed meridian: $38^\circ + 35^\circ = 73^\circ$ clockwise. The observed inclination indicates a paleolatitude (OP) of $38 \pm 10^\circ\text{N}$, compared with an expected paleolatitude of 61°N . Longitude is indeterminate.

Fig. 5.19

Paleolatitudes of mid-Cretaceous or presumed mid-Cretaceous magnetizations from B.C. and northern Washington, plotted with reference to the North and Central American coastline. Abbreviations as in Fig. 5.1. GBB is omitted: it plots slightly south of the map area. Paleolatitude and error band shown for QL. Longitude is indeterminate.

anomalous mid-Cretaceous magnetizations, the similarity of their paleolatitudes and the grouping of their paleopoles, far away from the APW path for cratonic North America (Fig. 5.16), is a compelling observation. The explanation is either that there has been uniform tilting of the whole region they represent, or that they were on a different plate to North America, and relative latitudinal displacements (2000 to 2500 km) and rotations (60 to 70° clockwise) have occurred. Of course, a combination of both processes is also possible (see Chapter 6.3). Whether tilting or plate movement is the explanation, it must have occurred between the mid-Cretaceous (ca. 100 Ma) and the mid-Tertiary (Irving et al. 1985), because the Oligocene Chilliwack batholith is paleomagnetically concordant with North America.

(Relative motions between the Quesnel terrane and North America between the Early Jurassic and the mid-Cretaceous, if any, is not constrained by the W and X magnetizations of this thesis, so will not be mentioned further. The question is reviewed in Irving & Wynne, in prep.)

That the anomalous localities were approximately in their present relative positions in the mid-Cretaceous is supported by geological evidence that there was indeed continuity between these terranes at this time. The Slide Mountain, Quesnel (QL, CM, and GB localities), Stikine and Cache Creek (AX locality) terranes were conterminous by the end of the Triassic, forming the Intermontane superterrane, based on the reworking and mixing of diagnostic rock types, and by overlapping deposits (Monger 1984). The Coast Plutonic Complex (SP and CS localities) is a large magmatic arc which was intruded along the boundary between the Insular (KY locality) and Intermontane

superterrane after their amalgamation in the Early to mid-Cretaceous (Monger 1984). Thus, allochthonous oceanic terranes from the Wrangell terrane eastwards to the Slide Mountain and Quesnel terranes are geologically and paleomagnetically linked by the mid-Cretaceous. Furthermore, because the Slide Mountain and Quesnel terranes were structurally tied to the Omineca Belt by the Middle Jurassic (this thesis), the implication is that at least the western part of the Omineca Belt, which is generally regarded as part of the North American (paleo-)continental margin, was also on this aberrant plate or 'microplate'.

There is, however, apparently equally good geological evidence that southern Omineca Belt strata are continuous with undoubted North American miogeoclinal strata in the Foreland Belt to the east. The problem is, therefore, that paleomagnetic and geological evidence disagree on whether or not the southern Omineca Belt was on the North American plate or on another, distant plate to the south. There is no obvious structural discontinuity between the aberrant mid-Cretaceous microplate and North America which could accommodate tectonic displacements of this magnitude and age, and thus reconcile the two lines of evidence. Such a structural discontinuity occurs in the northern Canadian Cordillera in the form of the Tintina Trench and Northern Rocky Mountain Trench dextral strike-slip fault system, but this is only about 50 or 25% (depending on estimates of offsets) of the amount of displacement indicated by paleomagnetism, and in any case, strike-slip displacements decrease southwards towards the southern Canadian Cordillera to virtually zero (Price & Carmichael 1986).

This delightful problem is discussed in Chapter 6.3.

[5] 8. SUMMARY OF PALEOMAGNETISM

Rees et al. (1985) may also serve as a summary of this chapter.

A paleomagnetic study of zeolite facies Takla Group rocks from southwest of the thesis area has identified four remanent magnetizations (W, X, Y, and Z) by thermal and alternating field step demagnetization. Presently only X and Y are significant. Stable endpoint X directions are southeast and upward with respect to paleohorizontal but are poorly grouped. It is probably a primary or Early Jurassic magnetization, but for various reasons its significance cannot be interpreted further.

The Y magnetization has intermediate coercivities and unblocking temperatures, identified by linear decay segments. Directions are northeast and downward with respect to present horizontal. None are antiparallel, i.e. southeast and upward, which would have indicated geomagnetic field reversal. The Y magnetization is interpreted to be a secondary viscous partial thermoremanent magnetization, acquired after Jurassic folding during an episode of regional heating, most likely mid-Cretaceous plutonism. A mid-Cretaceous age is supported by the absence of reversals, indicative of the Cretaceous Normal Superchron (118 to 83 Ma), and by the location of its paleopole (QL: 61.9°N, 14.6°W) among others from British Columbia and northern Washington which are argued to be of that age. These paleopoles do not coincide with any part of the APW path for North America.

The paleolatitude and rotation of QL is $38 \pm 9.8^\circ$ N, $34.9 \pm 12.5^\circ$ clockwise, compared with expected (North American) values of $61 \pm 3.8^\circ$ N, $38.3 \pm 7.9^\circ$ anticlockwise. Paleolongitude is indeterminate. The result indicates a relative latitudinal anomaly of $23 \pm 10.5^\circ$, and a relative clockwise rotation of $73.2 \pm 14.7^\circ$.

The explanation for QL and affiliated paleopoles is either that there has been uniform tilting of the whole region they represent, or that they were on a different plate to North America and 2000 to 2500 km of northward translation and 60 to 70° of clockwise rotation of that plate has occurred since the mid-Cretaceous. At least part of the Omineca Belt was attached to the discordant plate because they are linked by Middle Jurassic structures and mid-Cretaceous plutons. The problem is that there are no known structural or stratigraphic discontinuities which could fully account for tectonic displacements of the required magnitude and age between the discordant microplate and the cratonic miogeocline.

CHAPTER SIX

TECTONIC HISTORY

This chapter discusses the tectonic history of the thesis area, which involves the interpretation and synthesis of many aspects of the geology, local and regional (Section 2). A starting point (Section 1) is a discussion of the origin of the three main tectonostratigraphic 'assemblages': the Snowshoe Group of the Omineca Belt, the Antler Formation, and the Quesnel Belt of the Quesnel terrane, as this helps to constrain the scale and geometry of the pre-deformation setting. Section 3 compares and contrasts other parts of the Intermontane-Omineca Belt boundary in the Canadian Cordillera and its possible analogues in the conterminous United States, and also discusses the large-scale Mesozoic and Cenozoic tectonic implications of the paleomagnetic results given in Chapter Five.

The numerous sub-sections in each section are numbered for easier cross-referencing.

[6] 1. THE ORIGIN OF EACH OF THE MAIN TECTONIC
ASSEMBLAGES IN THE THESIS AREA

SNOWSHOE GROUP

(1) Depositional environment

The (meta-)sedimentary rocks of most of the Omineca Belt and the Foreland Belt are generally regarded as being derived from a west-facing continental margin sedimentary prism (Monger & Price 1979). Continental shelf facies occur in the east (Foreland Belt), while to the west (Omineca Belt) a thick succession of Upper Proterozoic and Paleozoic rocks probably represent continental slope and rise facies. Given its present position, the Snowshoe Group would appear to represent the most westerly part of the sedimentary prism, consistent with the tentative interpretation (Chapter 2.1) that the Snowshoe Group was deposited in the outer part of a submarine fan complex. This is the interpretation adopted in this discussion; however, it is admitted that the stratigraphic continuity between the Snowshoe Group and the Proterozoic and Paleozoic stratigraphy to its east has not been demonstrated satisfactorily (see Chapter 2.1), and until it is the paleogeography of the continental margin cannot be reconstructed with confidence.

(2) Primary thickness of the Snowshoe Group and its basement

Assuming the Snowshoe Group does represent the distal part of a continental margin sedimentary prism, it was probably deposited on attenuated and rifted continental or transitional crust. In the next sub-section on the origin of the Antler Formation, it will be concluded that the North American margin was rifted in the upper Paleozoic, in addition to what rifting it sustained in the Proterozoic prior to and

during the deposition of the Windermere Supergroup (Fig. 6.1a,b). Which phase of rifting and subsidence the Snowshoe represents, if not both, may only be deduced by further stratigraphic dating and facies analysis. For now, it is concluded (Fig. 6.2) that the Snowshoe Group represents a clastic sedimentary sequence no more than 10 km thick that was deposited on a thin basement disrupted by numerous listric normal faults, predominantly west-side-down, produced by one (Proterozoic) or two (plus Paleozoic) phases of attenuation and rifting.

ANTLER FORMATION AND OTHER MEMBERS OF THE SLIDE MOUNTAIN TERRANE

(3) The evidence for oceanic crust

The author's observations on the Antler Formation in the thesis area corroborate the interpretation of other workers on correlated rocks elsewhere that this unit represents metamorphosed and deformed oceanic crust. This is based partly on the rock association, namely metabasaltic greenstone, (layered) metagabbro, and meta-ultramafic rocks. These are the main rock types found in (metamorphosed) ophiolite complexes. Missing here are radiolarian cherts and thick accumulations of pillow lavas, but these are present in abundance in the type-Antler Formation in the Black Stuart synclinorium and in the Fennell Formation (Chapter 2.3). The interbedded radiolarian chert, argillite and basalt in the type-Antler Formation represent a condensed succession, diagnostic of pelagic sedimentation. About 100 Ma of conodont-bearing stratigraphy is recorded in less than 300 m (Struik & Orchard 1985).

It is true that nowhere in this belt is there preserved a sheeted dyke complex (but see Montgomery 1978 and Chapter 1.2), a feature of

Fig. 6.1

Cartoons showing evolution of western North American plate margin from Late Proterozoic to Early Jurassic (cf. Tempelman-Kluit 1979, his Figure 20). As explained in text, this model is largely based on previous work such as Davis et al. (1978), Monger (1984), Price et al. (1985). Crosses = continental crust basement, dots = continental margin sedimentary prism, black = oceanic crust, diagonal lines = subduction complex, H = Harper Ranch Group, Q = Quesnel Belt.

(a) LATE PROTEROZOIC TO LOWER PALEOZOIC. Generally passive continental margin sedimentation, westerly-prograding.

(b) DEVONIAN-MISSISSIPPIAN. Subduction begins in Devonian, dipping towards continent. Arc volcanism (Harper Ranch Group) begins on edge of continental margin. Extension within or behind arc leads to crustal attenuation. Guyet Formation conglomerate (ascribed a western source by Struik 1980) may indicate rifting.

(c) UPPER PALEOZOIC. Back-arc basin ("Antler back-arc basin" in text) opens up, generating oceanic crust presently represented by Antler and Fennell Formations and Kaslo Group. Kaslo-Milford depositional relationship in Kootenay Arc represents ocean-continent interface. Note hypothetical piece of continental crust in the basement of arc.

(d) PERMIAN TO EARLY TRIASSIC. Back-arc spreading ceases, as probably does arc volcanism and subduction at plate margin. This is approximately the time of the Sonoma orogeny in the United States, and Permian thrusting in the Kaslo Group may be a record of this compression event. Width of back-arc basin not known (indicated by ?) but was probably at least 150 km (see text, later).

(e) LATE TRIASSIC. Subduction resumes west of defunct Harper Ranch Group arc, either at the old subduction zone or at a new one. Generates calc-alkaline volcanism (Nicola and Takla groups of the Quesnel Belt) on top of the Harper Ranch Group. Together these form the Quesnel terrane. Subduction complex develops west of Quesnel Belt arc, forming the Cache Creek Group.

This time, there is no back-arc extension and oceanic crust formation (as in b,c) probably because entire plate margin is under compression. Black Phyllite mud is deposited in this "(Antler) marginal basin", disconformably on Antler Formation. Overlapped by volcanoclastic fan of the basinward-prograding Quesnel Belt (see Fig. 2.10).

Stikine terrane approaches from west (or northwest/southwest?) on subducting plate. Potassic (shoshonitic?) alkalic volcanism appears in Quesnel Belt.

(f) LATEST TRIASSIC TO EARLY JURASSIC. Stikine terrane collides with Cache Creek terrane (subduction complex) and Quesnel terrane, completing assembly of Intermontane superterrane (which also includes Antler-Fennell-Kaslo rocks of Slide Mountain terrane). Block faulting and uplift in Quesnel terrane may be due to collision, and may be related to potassic volcanism. This cross-section corresponds to Fig. 6.2.

Stage is set for collapse and obduction of Antler marginal basin onto North America in latest Early Jurassic (Fig. 6.6).

(a) Late Proterozoic to lower Paleozoic

North American
continental margin

(b) Devonian - Mississippian

Guyot conglomerate
western source?

Back-arc extension

(c) upper Paleozoic

Harper Ranch Group

Antler-Fennell-Kaslo (Slide Mountain
terrane to be)

Rifting and spreading

(d) Permian to Early Triassic

Early thrusting in Kaslo? (= Sonoma Orogeny?)

Spreading ceases
- basin width unknown

(e) Late Triassic

Stikine

Cache Creek

Quesnel Belt

Black Phyllite

(f) Latest Triassic to Early Jurassic

Stikine

Cache Creek

Quesnel

Slide Mountain

North
America

Fig. 6.2 Tectonic setting at the western edge of the North American plate in the late Early Jurassic (ca. 190 Ma), just prior to the D₁ obduction of the Antler marginal basin and the eastern edge of the Quesnel terrane arc. v = Quesnel Belt, dashes = Black Phyllite, black = Antler Formation, + = basement, unpatterned = sedimentary cover: Snowshoe Group in southwest, Kaza and Cariboo Groups in northeast, M = M-discontinuity. X-X' cross-section is approximately southwest-northeast; lower two sections are enlargements. Gap in oceanic crust in X-X' signifies uncertainty of width of basin (see text). Dashed line within or at base of black layer is projection of obduction detachment (Quesnel Lake Shear Zone, Eureka thrust, Pundata thrust).

some ophiolites, but this is not universally developed in modern oceanic crust anyway (Coleman 1977). In the Antler-Fennell-Kaslo belt (Slide Mountain terrane), the Black Riders Complex of Montgomery (1978) is the most complete and best documented ophiolitic sequence (Chapter 1.2).

The meta-ultramafic constituents of the Antler Formation could represent the basal ultramafic cumulates of oceanic crust, or the underlying serpentized peridotite of the upper mantle. However, the limited thickness of the Antler Formation and the properties described in the Black Riders Complex (Montgomery 1978) suggest that it is mainly or entirely oceanic crust.

The geochemical evidence for an ocean floor origin for the thesis-area Antler Formation was given in Chapter 2.3; again, this trend is consistent throughout the region (op. cit.). The interpretation of Aggarwal et al. (1984) is not considered to be a significant deviation from the consensus because their evidence is so restricted. Anyway, their identification of alkalic oceanic island-type metabasalt is certainly not out of character with an ocean-floor setting: ocean-floor seamounts are typically composed of alkalic basalt.

(4) Mid-ocean spreading centre or back-arc basin?

Accepting the oceanic crust interpretation, the next question is whether the Antler-Fennell-Kaslo rocks formed at a completely exotic mid-ocean spreading centre which was transported en masse to the North American margin (this is how Harms (1986) interpreted the Sylvester allochthon in northern British Columbia - see Section 3), or whether it formed 'in situ' in a peri-continental back-arc basin. At present, there is no reliable geochemical or petrological criterion to help

answer this question. However, the apparent absence of a sheeted dyke complex in the Slide Mountain terrane may be more indicative of crustal formation in a back-arc basin rather than in a mid-ocean ridge, according to Upadhyay & Neale (1979). They pointed to modern analogues and suggested that back-arc ocean basins widen by spreading in highly diffuse zones without the major development of sheeted dykes.

Another criterion favouring the back-arc basin alternative arises from the fact that such small basins tend to have shallow calcite compensation depths (CCD) so that the first major deposit on the basaltic crust is siliceous ooze (Jenkyns 1978). Conversely, the evidence from modern major oceans which have deeper CCD's is that the crust is overlain first by biogenic carbonate, followed by siliceous ooze after it sinks below the CCD. The fact that the Antler Formation is interbedded with radiolarian cherts, with only minor limestone, suggests that it represents a back-arc basin environment.

Two kinds of 'back-arc basins' are known: those created by extension and sea-floor spreading above a subduction zone behind an active island arc, and those that were formerly edges of large ocean basins which were later cut off by the development of an intra-oceanic subduction zone (Tarney & Windley 1981). Regional geological evidence indicates a back-arc basin origin of the first kind. [Note that the proximity of the Quesnel Belt (see below) is not relevant here, as these volcanics are Triassic-Jurassic.] The Antler-Fennell-Kaslo rocks in the basin are approximately contemporaneous with the mafic to felsic volcanic and sedimentary rocks of the upper Paleozoic (Upper Devonian to Permian) Harper Ranch Group of the Quesnel terrane (Price et al. 1981, 1985; cf. Monger 1977, Davis et al. 1978) which Monger & Ross (1971) suggested

represents an upper Paleozoic island arc. The nature of the contact between the Harper Ranch Group and the Slide Mountain terrane is not known, but assuming that it is not tectonic (that is, that they were formed adjacent to one another) it is quite plausible that the Slide Mountain terrane was formed by back-arc extension in the upper Paleozoic as the 'Harper Ranch Group ?arc' on the edge of the North American (paleo-)continental margin rifted and spread away to the west, creating a small oceanic basin between the arc and the continent (Fig. 6.1b,c,d; Davis et al. 1978, also Price et al. 1981, 1985; compare the model of Tempelman-Kluit 1979, his Figure 20). This suggests that the Harper Ranch Group is floored by a piece of North American crust. The width of the basin is not known, but palinspastic reconstructions explained in Section 2 suggest it was at least 150 km.

Beyond this circumstantial evidence, there is no physical evidence for the alleged rifting in the thesis area because the transition has been buried by the obducted Quesnel terrane. The critical evidence comes from the Kootenay Arc, where Klepacki (1983) and Klepacki & Wheeler (1985) have demonstrated a stratigraphic link between the Kaslo Group and the Omineca Belt. ?Lower Permian tholeiitic volcanic rocks at the base of the Kaslo Group are in depositional contact with the McHardy Assemblage of the suspect but presumed 'North American' Mississippian-Pennsylvanian Milford Group. The McHardy Assemblage is intruded by dykes and sills which fed the volcanic rocks. This primary relationship is provisionally extrapolated northwards to the Quesnel Lake area and is incorporated into the tectonic modelling in Section 2. The ultimate implication of it is that the Intermontane-Omineca Belt boundary, at least in central and southern British Columbia, is not a major plate suture. One possible connection with Devonian-Mississippian rifting is the

intrusion of the Quesnel Lake Gneiss at that time, although there is no petrologic property of the gneiss to implicate its formation with rifting.

The tectonic history in Section 2 really begins with the formation of the Quesnel Belt in the Late Triassic. To emphasize that the Antler Formation oceanic crust is not genetically related to the Quesnel Belt, the basin it represented in the Mesozoic (see below) will hereafter be called the Antler marginal basin rather than the Antler back-arc basin, its inferred original upper Paleozoic setting.

TAKLA GROUP AND BLACK PHYLLITE

As explained in Chapter 2.5, the Takla Group in the thesis area forms part of the Quesnel Belt of the Quesnel terrane, a belt of Upper Triassic to Lower Jurassic volcanic and volcanoclastic rocks and comagmatic intrusions. The 'basement' of the Quesnel Belt, which forms the remainder of the Quesnel terrane, is locally exposed in the south of the terrane, and consists of the Harper Ranch Group and possibly the gneisses of the Mt. Lytton Complex (Monger in Price et al. 1985). Precambrian continental crust may also be present (Petö & Armstrong 1976; see above).

Relevant to this thesis is the tectonic significance of the Quesnel Belt. Only a small representation of the Quesnel Belt outcrops in the thesis area, but the field observations and interpretations may be applied to the question on a regional scale.

The Black Phyllite, unit uTrBP, is herein incorporated as part of the Quesnel Belt (following Price et al. 1985) because it is gradationally overlain by Takla Group volcanoclastics.

(5) Depositional environment

The depositional environments of the Black Phyllite and the Takla Group were discussed in Chapters 2.4 and 2.5, respectively. The general Late Triassic setting appears to be a starved oceanic basin (cf. Read & Okulitch 1977) between a continental margin (presumed to be North America, to the east), and a mainly submergent volcanic complex to the west (Fig. 6.1e,f). This 'marginal basin' (Davis et al. 1978 called it a "back-arc basin") was floored by Mississippian to Early Permian oceanic crust represented by the Antler Formation and equivalents (Slide Mountain terrane). The centre of the basin was deep enough and flat enough to be below the CCD and to be virtually unaffected by gravity-driven bottom currents (rare in marginal basins: more common in 'open' oceans: Mitchell & Reading 1978), and it was filled with hemipelagic carbonaceous mud and silt (Black Phyllite), probably derived from a western source.

The basin shallowed to the west where the crust was building with the formation of basaltic and possibly andesitic volcanics, plutonic rocks and cogenetic sedimentary rocks of the Quesnel Belt. That part of the Quesnel Belt presently around latitude 53°N was about 50 km wide (it was probably much wider to the north and especially to the south). The Upper Triassic stratigraphy in the thesis area coarsens upwards from hemipelagic mud (Black Phyllite), through tuffaceous rocks, volcanic wackes, shallow water debris flow conglomerates, to basaltic pillow lavas. This succession indicates the progradation of a volcanic and volcanoclastic apron into the marginal basin (cf. Karig & Moore 1975, Fisher & Schminke 1984). The primary thickness of the Upper Triassic

rocks, based on the palinspastic cross-sections in this thesis, was roughly 15 km. An assemblage of this composition and thickness is particularly characteristic of marginal basins, but can also occur in linear chains of intraplate volcanoes, especially considering the alkalic composition of the volcanics; so was the Quesnel Belt an island arc or a volcanic chain of some other origin? These alternatives may be resolved by the petrology and geochemistry of the Quesnel Belt.

(6) Petrology and geochemistry

Schau (1968) studied the Nicola Group in south-central British Columbia, where basaltic andesite and andesitic volcanic rocks have a generally calc-alkaline geochemical trend (Table 6-1, column 5; see also Mortimer 1986, Spence 1985). He was probably the first to suggest that the Nicola Group represents the altered and "broken" remnants of a "volcanic island archipelago", presumably referring to what in modern terms is an island arc.

However, much of the Quesnel Belt is alkalic rather than calc-alkaline (Barr et al. 1976, Fox 1977). The major element composition of these mildly potassic alkalic basalts resembles very closely that of alkali basalts in oceanic islands associated with the Mid-Atlantic Ridge, such as St. Helena, Gough Island, and Tristan da Cunha. This led Morton (1976) and, up to a point, Preto (1977) to propose that the Quesnel Belt was in fact related to rifting rather than to subduction. Lefebvre (1976) also noted the resemblance between the Nicola Group and the Atlantic oceanic islands, but he was ambivalent in his interpretation, pointing out that there was a significant discrepancy in the TiO_2 contents, shown clearly in Table 6-1: modern oceanic island alkali basalts contain 3 to 4% TiO_2 whereas nearly all analysed basalts from the

TABLE 6-1: REPRESENTATIVE CHEMICAL ANALYSES OF BASALTS FROM THE QUESNEL BELT, ISLAND ARCS, AND OCEANIC ISLANDS

	1	2	3	4	5	6	7	8	9	10	11	12	13
SiO ₂	48.06	49.40	48.50	45.86	48.50	46.83	48.37	47.63	48.87	47.70	45.54	42.43	43.10
TiO ₂	0.98	0.76	1.08	0.88	0.77	0.85	0.58	0.88	1.23	3.20	3.28	4.11	4.10
Al ₂ O ₃	13.47	14.30	15.01	13.81	15.60	12.25	12.45	17.71	15.44	15.20	15.78	14.15	13.10
Fe ₂ O ₃	2.74	2.66	5.38	8.48	10.10*	4.96	4.89	2.15	10.59*	2.30	2.50	5.94	5.50
FeO	8.06	7.54	5.17	1.49		5.29	5.33	6.74		9.70	9.56	8.48	8.50
MgO	8.41	6.49	5.86	8.19	6.38	8.96	10.53	9.98	9.07	9.70	6.33	6.71	9.00
CaO	10.18	9.20	8.78	10.21	8.57	13.30	11.65	11.03	10.10	8.90	9.33	11.91	12.40
Na ₂ O	3.01	3.46	3.41	2.40	3.21	2.80	2.59	2.74	3.74	2.70	3.59	2.77	2.70
K ₂ O	2.04	2.48	1.95	2.61	2.59	1.90	1.75	0.72	0.53	1.60	1.33	2.04	1.60
P ₂ O ₅	0.34	0.44	0.40	0.23	0.38	0.44	0.34	0.20	0.30	0.17	0.55	0.58	-
H ₂ O ⁺	2.44	loss 2.94	2.49	2.30	2.94	1.43	loss 1.83	-	-	0.67	1.85	0.34	-
H ₂ O ⁻	-	-	0.22	0.41		0.72	-	-	-	0.50	0.25	0.44	-
Total	99.73	99.67	98.25	96.87	99.04	99.73	100.31	99.78	99.87	101.34	99.89	99.8	100.00
K ₂ O/Na ₂ O	0.68	0.72	0.57	1.08	0.80	0.68	0.68	0.26	0.14	0.59	0.37	0.74	0.59
Sr	995	-	-	-	-	-	972	607	550	-	525	1000	-
Zr	59	-	-	-	-	-	34	95	80	-	190	200	-
Y	21	-	-	-	-	-	14	21	12	-	43	15	-
Cr	181	-	-	-	-	-	360	668	508	-	#	65	-

Notes: Values of major elements in weight per cent. Columns 1-4 = alkali basalts from the Quesnel Belt, column 5 = calc-alkaline basalt from the Quesnel Belt, columns 6-9 = island arc alkalic basalts, columns 10-13 = oceanic island alkali basalts.

Sr, Zr, Y, Cr given in parts per million.

* = Total Fe.

= Cr highly variable between 15 and 280 ppm.

TABLE 6-1 (concluded):

Key to chemical analyses

1. Average of 11 alkali olivine basalts, alkali basalts, analcime basalt, nepheline basalt, Horsefly Group (calculated by author from data in Morton 1976).
2. Average of 35 alkali olivine basalts, alkali basalts, Morehead Lake (calculated by author from data in Bailey 1978).
3. Average of 8 alkali basalts, Nicola Group (calculated by author from data in Preto 1977).
4. Alkali basalt, potassic series, Nicola Group (Lefebure 1976).
5. Average of 8 altered calc-alkaline basalts, Nicola Group (Schau 1968).
6. Average of 2 potassium nephelinites, potassic basanite, potassic transitional basalt, New Ireland (Johnson et al. 1976).
7. "Shoshonitic" basalt, Viti Levu, Fiji (Gill 1970).
8. Alkali olivine basalt, Grenada, Lesser Antilles (Sigurdsson et al. 1973).
9. Alkali basalt, Penguin Island, Scotia Arc (Tarney et al. 1977).
10. Average olivine basalt, Gough Island (Le Maitre 1962).
11. Average of 5 basalts, St. Helena (Baker 1969).
12. Alkali olivine basalt, Tristan da Cunha, (Baker et al. 1964).
13. Average alkali basalt, Tristan da Cunha (Baker et al. 1964).

Quesnel Belt contain averages of less than 1.1% TiO_2 (Fig. 6.3). The TiO_2 factor is important because it has been shown empirically that basalts formed at island arcs above subduction zones consistently have TiO_2 values less than about 1.2% (e.g. Chayes 1964, Perfit et al. 1980), whereas those of oceanic islands are consistently higher. TiO_2 appears to be, in general, a rather good discriminant, and Bailey (1978) used it to infer that the Takla Group rocks in the Morehead Lake area represented island arc volcanism, in variance to Morton's (1976) interpretation of the neighbouring area.

There is, thus, an ambiguity regarding the tectonic significance of the Quesnel Belt, Morton (1976) and Preto (1977) favouring a rift-related origin, versus Schau (1968) and Bailey (1978) favouring an island arc setting. The author investigated whether trace elements such as Zr and Y could discriminate between these two possibilities, following the method of Pearce & Cann (1973). Unfortunately, of the above-mentioned studies only Morton (1976) presented analyses of these trace elements. His data clearly falls into the fields affiliated with arc basalts, in terms of trace element abundances (Table 6-1) as well as their ratios (Figs. 6.4 and 6.5). This limited but important trace element data, together with the widely regarded TiO_2 discriminant and the calc-alkaline rocks (which statistically are diagnostic of convergent plate boundaries), lead the author to prefer an island arc origin for the Quesnel Belt, as has been depicted in Fig. 6.1e,f.

Recent petrographic and geochemical work by Mortimer (1986) corroborates this interpretation. He obtained low values of Ti, Nb, Zr, and Y from Nicola Group basalts (plus variable enrichment in K, Ba and Rb), inferring that the volcanics were generated by subduction.

Fig. 6.3

TiO_2 - SiO_2 diagram for alkalic basalts and more differentiated rocks (with higher SiO_2) from the Quesnel Belt, compared with modern island arc and oceanic island basalts. Diagram is from Bailey (1978) with the addition of mean values from four modern island arcs: New Ireland, Viti Levu, Scotia Arc and Grenada. The close correlation between the latter group and the Quesnel Belt strongly supports an island arc origin for the Quesnel Belt.

Fig. 6.4

TiO₂-Zr diagram for alkalic basalts from the Quesnel Belt, modern island arcs and oceanic islands. Fields from Pearce (1980). MORB = mid-ocean ridge or ocean floor basalts. Symbols as for Fig. 6.3 except that solid circles = Quesnel Belt in the Horsefly area only (data from Morton 1976), and oceanic islands are represented by mean values from Tristan da Cunha (open circle) and St. Helena (open triangle). Note restriction of Quesnel Belt to field of arc lavas.

Fig. 6.5

Ti-Y-Zr diagram for alkaline basalts from the Quesnel Belt (in the Horsefly area only (Horsefly Group): data from Morton 1976), modern island arcs and oceanic islands. Fields from Pearce & Cann (1973). Note that the fields of arc basalts (fields B and D) actually completely overlap field C (ocean floor basalt) where most of the Quesnel Belt data lies, such that field C also represents arc basalts (see Pearce & Cann 1973). (This feature is also partially true for Fig. 6.4.)

Finally, a destructive-plate margin origin was also postulated by Beddoe-Stephens & Lambert (1981) for the Early Jurassic Rossland Group (Fig. 2.3), which although isolated from them, is most likely a direct correlative of the Takla and Nicola Groups of the Quesnel Belt. From Ti, Y, and Zr, and isotopic characteristics they concluded that these alkali basalts were island arc-generated, and not contaminated by continental crust. These authors further suggested that the assemblage has shoshonitic affinities (another characteristic of magmatic arcs: see below) and was probably formed in a localized oceanic basin, after the end of a phase of subduction, in a faulted, tensional environment.

(7) Potassic alkali basalts versus shoshonites, and their tectonic significance

One thorny problem remains to be discussed: alkalic basalts are not particularly widespread or abundant in modern island arcs. Where they occur they tend to be very restricted in space and time. A variety of potassic alkalic basalt that is not uncommon in modern island arcs is the shoshonite suite. The author tentatively concludes from a review of the major element chemical compositions of representative Quesnel Belt alkali basalts (see Table 6-1, columns 1 to 4), that they are not shoshonites *sensu stricto* (see Joplin 1968): shoshonites are mainly hypersthene-normative, and have higher SiO_2 , Al_2O_3 , and $\text{K}_2\text{O}/\text{Na}_2\text{O}$, and lower Fe-enrichment than Quesnel Belt basalts. The latter may be transitional between alkali basalt and shoshonites, however.

While non-shoshonite analogues of the Quesnel Belt are rare in modern island arcs, they do occur. Examples include the potassic alkali basalts in New Ireland (Papua New Guinea: Fox 1977), Viti Levu (Fiji), and the Scotia Arc (Table 6-1, columns 6,7,9). Note that they are all low in TiO_2 .

However, Mortimer (1986) and Spence (1985) did interpret Quesnel Belt alkali basalts as shoshonites (as did Beddoe-Stephens & Lambert 1981, for the Rossland Group). Mortimer recognized a high K- to shoshonitic series in the Nicola Group, as well as calc-alkaline and low-K tholeiitic series. He also found a crude space-time distribution in which the calc-alkaline rocks are generally (but not exclusively) west of and younger than the shoshonitic rocks, suggesting a west-facing and evolving arc.

Similarly, Spence (1985) reviewed several Upper Triassic and Lower Jurassic alkaline potassic volcanic suites from the Quesnel Belt (and ~~other~~ parts of British Columbia, including some from the Stikine terrane) and classified them as part of the shoshonite association. She emphasized that shoshonitic rocks with low TiO_2 are characteristic of plate collision and magmatic arcs, and echoed the point being made in this thesis that the potassic nature of these basalts does not necessarily mean that they were generated by rifting.

The reason for alkalic volcanism in island arcs is not well understood. The question is beyond the scope of this thesis, but it is pertinent to point out the following.

(a) Gill (1976, 1982) has linked shoshonitic volcanism in Fiji to post-subduction stabilization and uplift of the arc, with the rifting of the arc to form the Lau Basin. [This was followed by extrusion of alkali olivine basalt of oceanic island type, which illustrates how complex tectonics and volcanism can be, even over short periods of time.]

Morrison (1977) noted that "shoshonitic magmatism is commonly associated with block faulting and uplift, either following a subduction

phase or during a transition between two subduction regimes of different orientation".

(b) De Long et al. (1975) noted a striking coincidence between alkalic volcanism in modern arcs and hinge faulting, or the subduction of fracture zones or other major lineaments. They speculated that these features directly or indirectly create access to sub-lithospheric mantle not normally tapped in subduction zones, allowing the extraction of alkaline magmas from deep levels.

(c) In variance to early ideas (e.g. the K_2O -depth correlation of Dickinson & Hatherton 1967), the occurrence of shoshonitic magmas is apparently not necessarily related to the lithospheric thickness above the subduction zone, nor does it always follow in time tholeiitic and calc-alkaline volcanism (Arculus & Johnson 1978). Arculus & Johnson (1978) further suggested that alkalic volcanism may be the result of the partial melting of a mantle wedge chemically modified by a long history of subduction, and (or) tectonic events not directly related to subduction (e.g. (b), above).

Implications for the Quesnel Belt. The Quesnel terrane is bounded along its western side by the Cache Creek terrane, a melange of ocean floor rocks interpreted to represent a subduction complex (Monger et al. 1982; Fig. 6.1e). According to Monger (1981, 1984; also Travers 1978), the Quesnel Belt was generated by the eastward subduction of an oceanic plate (bearing the Stikine terrane?), the Cache Creek terrane representing the (forearc) subduction complex. This subduction and the related magmatism presumably declined with the emplacement of the Stikine terrane against the Cache Creek terrane at the end of the Triassic, or

the beginning of the Jurassic (Fig. 6.1f; Monger & Price 1979; see also Brown et al. 1986, Fig. 4). Perhaps the change in the chemistry of the volcanism from calc-alkaline to alkaline (cf. Gill 1982 and point (a) above, Morrison 1977) at this time was related to the stabilizing of the Quesnel terrane, possibly involving block faulting and uplift, as it adjusted to this new terrane configuration. Preto (1977) in fact inferred that centres of Nicola Group volcanism were controlled by the distribution of large high-angle faults (compare the model of Spence 1985).

It is possible to speculate further that if Monger & Ross (1971) are correct in interpreting the Harper Ranch Group as an arc assemblage, subduction-related magmatism operated in the Quesnel terrane from the Pennsylvanian to the Late Permian (the range of the Harper Ranch Group) and again in the Late Triassic to Early Jurassic (Quesnel Belt), a considerable span of time during which the subjacent upper mantle would have sustained chemical modification possibly leading to potassic volcanism (cf. Arculus & Johnson 1978 and point (c), above). Another possible source of the K_2O enrichment is the postulated continental crust in the Quesnel terrane, although this would be refuted by Beddoe-Stephens & Lambert's (1981) interpretation of their isotopic data.

The point of these speculations is to defend the interpretation of the Quesnel Belt as a Triassic-Jurassic island arc, which needs to be done in spite of the calc-alkaline rocks and the compelling TiO_2 criterion, because of the uncharacteristic alkalic nature of the remainder of the volcanics. In summary, therefore, the Quesnel Belt probably formed at a convergent plate boundary, above an east-dipping subduction zone. Subduction and mainly calc-alkaline magmatism with low TiO_2 contents gradually ceased with the emplacement of the Stikine

terrane from the west. The arc stabilized and was uplifted and faulted, with the end-production of potassic, alkalic volcanics (still with low TiO_2) and plutons, possibly in part a result of partial melting in an upper mantle chemically modified by a prolonged history of mantle-slab interactions.

[6] 2. TECTONIC MODEL

The following tectonic model is built around a series of palinspastic cross-sections which are proposed for the Early Jurassic, the early Middle Jurassic, and the late Middle Jurassic. The cross-sections are 'centred' on the thesis area but encompass the full thickness of the crust and extend far into the (paleo-)continental margin so that they and the accompanying explanations are forced to be compatible with regional data rather than be ad hoc.

EARLY JURASSIC

At this stage (Fig. 6.2, in Section 1), the rocks now exposed in the thesis area formed the eastern edge of an island arc complex (Quesnel Belt of the Quesnel terrane), and part of a marginal basin floored by oceanic crust (Antler Formation) which was contiguous with the continental margin to the east (the Snowshoe Group of the Omineca Belt). In this scenario, the Quesnel terrane marks the western margin of the North American plate.

(8) Dimensions of the components of the cross-section in Fig. 6.2

The basement of the continental margin is assumed to be attenuated continental or transitional crust (Monger & Price 1979), between 5 and 15 km thick (this would thicken northeastwards beyond the cross-section towards a 'standard' thickness of about 30 km). The thickness of the sedimentary cover is conjectural but the 10 km thickness shown, tapering southwestwards to zero, is an actualistic estimate. The Quesnel terrane arc is about 25 to 30 km thick. Probably the main uncertainty is the width of the intervening marginal basin. A minimum upper plate-lower plate cut-off resulting from the collapse of this basin via obduction⁸ (see below, and Fig. 6.6) yields a minimum width of the basin of 150 km, but as neither actual cut-offs are observable, the basin may have been much wider (the question is raised again in sub-section 11).

LATEST EARLY JURASSIC TO EARLY MIDDLE JURASSIC

(9) Timing of onset of D₁ obduction

Marine sedimentary rocks were still being deposited in the Sinemurian (ca. 200 Ma) in the Morehead Lake area (Bailey 1978), and marine Toarcian (ca. 190 Ma) rocks are known from elsewhere in the Quesnel terrane (Tipper 1984). If the end of the marine record marks the time of uplift in the Quesnel terrane in response to crustal shortening and underplating by the continental margin, then collapse of the marginal basin and obduction of it and the arc began at approximately the end of the Early Jurassic, about 190 to 185 Ma. (Thrusting may have begun earlier farther east in the marginal basin, however.) The D₁ Quesnel Lake Shear Zone is the main expression of this thrusting (Fig. 6.6). Note that this timing is a maximum estimate; the thrusting and

Fig. 6.6

Tectonic reconstruction for the early Middle Jurassic (ca. 180 Ma), immediately after the telescoping and obduction of the Antler marginal basin and the Quesnel terrane arc onto the continental margin. Symbols as for Fig. 6.2. Quesnel terrane, (Omineca Belt) cover, and (Omineca Belt) basement are each area-balanced with their counterparts in Fig. 6.2; plane strain is assumed. Basement and cover are 240 km wide compared with 280 km in Fig. 6.2; this inferred arbitrary -15% (plane) strain has been incorporated into this D_1 cross-section, in addition to the collapse of the basin, to represent some shortening and thickening due to reverse movement on pre-existing rift-related listric faults in the basement during obduction, forming F_1 folds in the cover (see sub-section 13). The Quesnel terrane is about 13% shorter and thicker than in Fig. 6.2.

obduction may not have started until the Middle Jurassic (ca. 185 to 180 Ma); the apparent uplift of the Quesnel terrane in the late Early Jurassic might have been caused by something else.

These timing arguments are compatible with Klepacki (1984) who dated the thrusting of the "Eastern Assemblage" and "Quesnellia" onto North America in the Kootenay Arc at "between the early Toarcian (193 Ma) and ca. 173 Ma" (U-Pb zircon age of cross-cutting intrusion, the Kuskanax batholith).

The late Early Jurassic was approximately the time that the north-central Atlantic ocean opened, and this has been linked with the onset of major deformation in the Cordillera (Wheeler 1970, Monger & Price 1979). If the North American plate was rigid, it is indeed plausible that crustal shortening was initiated at its weakest point: the Antler- and related marginal basin(s) lining its western edge (cf. Coney et al. 1980). Quite different relative plate motions in the 'proto-Pacific' ocean to the west could have triggered this tectonism independent of Atlantic spreading, of course, but this coincidence in timing strongly suggests a direct cause and effect relationship.

(10) Vergence of thrusting

The fact that the Quesnel terrane stratigraphy youngs in a direction away from the Quesnel Lake Shear Zone, and the sense of shear determined from kinematic indicators described in Chapter 3.2, combine to produce a geometry which requires that the Quesnel terrane was carried on a thrust which cut up-section towards the continental margin to the east or northeast, placing the Quesnel terrane on top of the Omineca Belt. The Antler Formation qualifies as an obducted ophiolite complex (Montgomery

1978).

A northeastward vergence for the Quesnel terrane is supported by the east-facing paleoslope implied by the chert clasts (from the Cache Creek Group?) in the conglomerate in unit 1mJ (Chapter 2.6). This paleoslope, facing a foredeep in front of the allochthon, is depicted in Fig. 6.6.

The direction of emplacement relative to the 'strike' of the Omineca Belt may be judged from the D_1 stretching lineations in mylonitic rocks in the Quesnel Lake Shear Zone. However, the azimuths of these lineations have been rotated by later deformations, particularly by the F_2 folding and the L_2 axial extension. Given the apparent influence of the L_2 northwest-southeast extension, and there being no other known post- D_1 stretching, then the preservation of a sub-domain of northeasterly-trending L_1 stretching lineations in the Western Quesnel Lake Gneiss north of Sellar Creek, on the overturned limb of the D_2 Sellar Creek syncline (Fig. 3.1), suggests that the L_2 axial extension was minimal or non-existent there. The lineations are at a very high angle to the axis of this fold, so unfolding the fold should conserve the displacement vector. This L_1 trend is therefore the closest thing to a pristine D_1 relative displacement vector, and indicates that it was virtually orthogonal to the strike (most probably also the 'paleotrend') of the Omineca Belt, and not particularly oblique to it. Fig. 6.6 is parallel to this displacement vector.

(11) Extent of allochthon - magnitude of obduction

The tectonic contact between the Quesnel terrane allochthon and the Omineca Belt is folded by D_2 deformation (Fig. 6.9). Unfolding this contact into a planar line without compensating for internal strain

yields a width of about 150 km (Fig. 6.6). The leading edge of the allochthon, that is the upper plate cut-off, is not preserved in the Cariboo Mountains to the northeast, and the lower plate cut-off is buried by the Quesnel terrane to the southwest, so the full extent of the allochthon is not known. Furthermore, the total relative displacement is this minimum 150 km plus the amount involved in any thrust duplication in the allochthon. No imbricate D_1 thrusting has been recognized in the thesis area, but there is significant thrust duplication in the Antler Formation in the Black Stuart synclinalorium (Struik 1982b, Struik & Orchard 1985; see sub-section 15). In the absence of better constraints, the magnitude of obduction is postulated to be at least 150 km.

(12) Rate, duration and sedimentary record of obduction and D_1

It was suggested above that the collapse and obduction of the Antler marginal basin began at the end of the Early Jurassic, between about 190 and 185 Ma. Evidence of the rate of relative movement during obduction is not available, but the very early half-spreading rate of the North American plate (at ca. 180 Ma) has been estimated by Sclater et al. (1980, from the area of 180 Ma-old oceanic crust) to have been quite slow, comparable to the modern rate which is about 2 cm a^{-1} .

By graphing relative displacement rate against magnitude of obduction (Fig. 6.7), an estimate of the duration of obduction is obtained, namely between 7 and 10 Ma. Subtracting this from 190 to 185 Ma suggests that the obduction and D_1 deformation in the thesis area was completed by about 180 Ma.

A problem emerges when this timing is related to the age and paleo-environment of unit 1mJ (see Chapter 2.6). These rocks were interpreted

Fig. 6.7

Relative displacement between obducted Quesnel terrane and Omineca Belt (ordinate) versus relative convergence velocity (abscissa) based on plate reconstructions (see text). Shaded field of values is considered to be most representative of D_1 obduction: a displacement of 150 km at about 2 cm a^{-1} would take between 7 and 10 Ma.

to have formed, at least partly, in a terrestrial environment in the early Bajocian (ca. 183 to 180 Ma). The unit might represent a non-marine to shallow marine ?foredeep sedimentary basin in the Intermontane superterrane, flanked by an easterly-facing paleoslope supplying clasts (in subunit 1mJc) from the Cache Creek terrane situated to the west. Unit 1mJ overlies Norian rocks (208 Ma) of unit TrJT, which were buried to a depth of at least 10 km during obduction (see sub-section 14), so unit TrJT would have had to be exhumed that much before the early Bajocian (ca. 183 Ma). This uplift and erosion must have post-dated D_1 movement on the Quesnel Lake Shear Zone, or the shear zone would have become brittle, which it didn't. So would there have been enough time to complete obduction, beginning at 190 Ma, and 10 km of erosion of the upper plate before 183 Ma? At a substantial erosion rate of 1 mm a^{-1} , 10 Ma would be needed for erosion in addition to the 7 to 10 Ma needed for obduction, which is more than double the time that is available (190-183 Ma). Starting the obduction earlier than 190 Ma would help but in any case, the thesis area would have to be re-buried quickly during D_2 to re-attain the estimated pressures of M_2 (ca. 20 km). The unavoidable implication is that unit 1mJ, if it is correctly dated, does not simply reflect a stratigraphic unconformity.

Thus, the existence of 183 Ma sediments among Takla Group rocks which at that time should have been well-buried and at a few hundred degrees, creates problems. One solution would be that unit 1mJ is completely fault-bounded and has been dropped down onto unit TrJT by normal faulting in the Late Jurassic or later. Further work on this problem should check the identification and dating of the fossils; if they are in fact significantly older or younger, the matter might be resolvable.

Notwithstanding this problem, the end of D_1 obduction is provisionally interpreted to be about 180 Ma. It cannot be much younger because D_2 and D_3 deformation occupies the time between about 175 and 165 Ma (see sub-section 19).

(13) D_1 structures

The shear and flattening related to the thrusting of the Quesnel terrane generated northeasterly-verging folds and penetrative foliation (S_1) in the lower plate of the Quesnel Lake Shear Zone and in the deeper, ductile rocks of the upper plate (Fig. 6.6). No definite F_1 folds or thrust imbrication have been found in the thesis area but easterly-verging first phase structures have been identified by Elsby (1985), Montgomery (1978) and Murphy (1985) (Table 6-3). Bedding and S_1 were generally transposed parallel to each other (S_{0-1}) and to the shear zone. Northeasterly-verging structures in the lower plate could also have been induced by the proposed mode of basement contraction, namely reverse faulting on pre-existing westward-dipping rift-related faults. A plane strain shortening and thickening of about 15% in each plate, compared with their Early Jurassic dimensions, has been incorporated into Fig. 6.6. to represent the internal strain caused by obduction.

The lenses of steatite that occur within the Snowshoe Group within a few hundred metres of the Quesnel Lake Shear Zone (Chapter 2.3) are remnants of basal Antler Formation meta-ultramafic rocks. They were incorporated there by folding and (or) imbrication of the upper plate during the thrusting, in the manner depicted schematically in Fig. 6.8.

(14) P-T conditions in the Quesnel Lake Shear Zone

Temperatures must have been at least about 400° to 450°C in the

Fig. 6.8

Schematic diagram explaining possible mode of incorporation of Antler Formation steatite lenses (subunit PPA1) within the structurally highest part of the Snowshoe Group beneath the Quesnel Lake Shear Zone during D₁ shearing. Initially a fold or imbricate thrust forms in the shear zone and becomes attenuated and cut off from the upper plate. Embedded in Snowshoe Group pelites, it is further attenuated and dismembered into discrete lenses.

Quesnel Lake Shear Zone to produce the mylonitization in the Western Quesnel Lake Gneiss and the Snowshoe Group (Chapter 3.2). However, it was probably not much hotter than this because of the transitional brittle-ductile deformation of the feldspar (notwithstanding the reservations outlined in Chapter 3.2), and the absence of relict amphibolite facies mineral assemblages in S_{0-1} or S_1 .

The pressure in the Quesnel Lake Shear Zone is estimated at 450 to 500 MPa, based on a 17 km overburden thickness (Fig. 6.6) with an average density of 2.85 g cm^{-3} . Based on these figures, the (linear) geothermal gradient during D_1 was about 26° km^{-1} ($450^\circ/17 \text{ km}$) which is a reasonable value.

However, if the overridden part of the Omineca Belt in the immediate lower plate of the shear zone in the thesis area was close to the earth's surface prior to obduction (Fig. 6.2), and not then under a high geothermal gradient, then it must be explained how it came to be at about 450°C during obduction. There are two main possibilities to consider.

(a) Thermal recovery in the lower plate (upward migration of isotherms) was rapid because heat was transferred from the hot base of the upper plate, namely the Antler Formation oceanic crust. The best examples of such basal "dynamothermal aureoles" occur beneath ophiolites that include thick, very hot upper mantle peridotite, and they commonly have an inverted gradient. This process does not apply to the thesis area because first, there is no major upper mantle component (if any) in the Antler Formation; second, even if there was, it would not have been very hot - there was at least 50 Ma between the formation of the Antler Formation and its obduction, during which time it would have cooled; and third, there is no evidence of an inverted metamorphic gradient.

Nevertheless, the base of the upper plate must have had a temperature of several hundred degrees, and this would have provided an effective thermal blanket to whatever thermal recovery ensued in the lower plate, expediting the attainment of thermal equilibrium between them. In this way, a temperature of ~~about~~ 450°C in the lower plate could have been achieved quite quickly.

(b) Another possibility is that the lower plate wasn't near the surface at all when obduction began: it was deeper and hotter. This is interesting because it implies that there was some extra cover to the Snowshoe Group which was underplated to the overriding Quesnel terrane as a 'horse' or slice and removed, exposing (a ramp of) formerly deeper, warmer rocks to the shear zone. This idea is additionally attractive because it provides a (tectonic) way of 'unroofing' the Quesnel Lake Gneiss, obviously a pluton which must have been at some depth in the Snowshoe Group prior to obduction. (It could also have been exhumed by prior erosion, however.) The poor knowledge of regional Snowshoe Group stratigraphy makes this option difficult to test.

In conclusion, the simplest explanation for the rapid thermal recovery in the lower plate is the heat transfer, or at least the insulation, provided by the 'warm' upper plate (a, above). Possibility (b) may also be important, but is presently difficult to demonstrate.

(15) Origin of the Quesnel Lake Shear Zone

What caused the Antler marginal basin to collapse where it did?

Modern marginal basins commonly have high heat flow values and high geothermal gradients, effecting a shallow brittle-ductile transition.

4

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The thinner the 'elastic' lithosphere is, the lower is its bulk strength. If the Antler marginal basin had a high geothermal gradient by virtue of its subjacent subduction zone, and a shallow brittle-ductile transition, this would help to explain why it was a weak link in the westernmost North American plate.

How deep was the brittle-ductile transition and where was it located? The largest thickness estimated for the Antler Formation in the region is 3000 m (for the Black Riders Complex, Montgomery 1978). The formation in the thesis area is at most 1000 m, and usually less than 250 m thick, but it has probably been thinned here tectonically from a greater thickness. Considering these figures and the fact that there is apparently no large component of metamorphosed upper mantle associated with the Antler Formation (or equivalents), it seems that it was detached from the rest of the oceanic lithosphere at a high level within oceanic crust. Some part of this zone (dashed line in Fig. 6.2) must have coincided in the beginning with the brittle-ductile transition and the initial detachment, from which it propagated. It is thought likely that the detachment initiated at a deep level, probably in the uppermost mantle about 25 km beneath the arc in the southwest, and propagated upwards within the oceanic crust towards the continental margin. In discussing subduction-obduction inception, Casey & Dewey (1984) regarded the interface between an island arc and oceanic lithosphere along a back-arc margin as being:

"a site in which flexural and torsional rigidity of the lithosphere is extremely low stress levels required to initiate rupture at these types of margins are probably smaller [than at Atlantic- and Andean-type margins] and may initially lead to rapid thermal resorption of the lithosphere within the hot underlying mantle [so that] mantle drag forces resisting subduction would therefore be initially minimal".

What controlled the polarity of the obduction-subduction? Beyond the reasons given above, this is difficult to answer. One possibility is that hypothetical basinward-dipping listric faults in the westernmost Omineca Belt continental or transitional crust basement (formed by rifting in the Upper Proterozoic or upper Paleozoic: see Section 1) were reactivated in the reverse sense, perhaps simultaneously with the oceanic thrusting, establishing an upper plate to the northeast sense of crustal shear (cf. the earlier comments on the vergence of F_1 folding). A series of these stress guides would tend to force the continental margin down and under as opposed to up and over the converging Antler marginal basin. Also, a warm and elevated marginal basin versus a cool and sediment-burdened continental margin may have promoted the partial underthrusting of the latter under the former.

How was the upper plate strong enough to have been transported as a rigid thrust sheet? The strength of a thrust sheet is partly a function of its tapered shape (Davis et al. 1983). Treating it as 'thrust sheet', the upper plate of the Quesnel Lake Shear Zone in the thesis area, that is towards the rear of the 'taper', was thick enough (ca. 17 km) to behave rigidly and move en masse, and it is apparently not imbricated by thrusts. Also, the Quesnel Lake Shear Zone at these depths was plastic, ~~so~~ had a comparatively low creep strength. The reconstruction in Fig. 6.6 shows that this thickness tapers northeastwards to a thin toe as the Triassic-Jurassic cover dwindles to nothing. If this reconstruction is correct, then in the northeast the toe of the upper plate consisted of only perhaps 2 to 4 km of oceanic crust at the most, and the basal detachment must have been brittle. In order to be thrust over the continental margin, either the yield stress of the detachment would have

to be extremely low at the toe, or the leading edge of the wedge would have had to be thickened by thrusting or folding to maintain its internal strength and its critical taper (see Davis et al. 1983).

The most northeasterly part of the allochthon that is preserved to help resolve this issue is in the Black Stuart synclorium (Fig. 1.2), where the 1100 m-thick Antler Formation is separated from the Omineca Belt by "mylonite" (Struik 1981a,b). It is interpreted that this is a klippe or salient of the obducted Antler marginal basin, following Monger (1977) and Monger et al. (1982). The 1100 m is a structural thickness: Struik & Orchard (1985) have documented at least three or four "imbricated layer-parallel thrust sheets", based on the repetition of conodont-bearing strata. These authors implied that these thrusts formed prior to or synchronous with the emplacement of the Antler Formation. Therefore, it is suggested by the author that they represent the mechanism by which the allochthon was thickened during transport to achieve its critical taper, and made capable of being transported as a whole along its basal detachment. Incidentally, it is not clear if this part of the Antler Formation was near the thin leading edge of the allochthon: if Struik's (1981a,b) identification of mylonite is accurate, there must have been a substantial thickness to the allochthon here. It would follow that the allochthon did extend much farther east before it was eroded away.

Another way to thicken and strengthen the upper plate is to underplate slices or horses of the Omineca Belt to it, as was speculated above in sub-section 14.

(16) Pundata thrust, Pleasant Valley thrust, and Eureka thrust

Pundata thrust is the name given by Struik (1981b) to the thrust

that separates the Antler Formation of the Slide Mountain terrane in the Black Stuart synclinorium from the underlying Snowshoe Group, Cariboo Group, Black Stuart Group, and Kaza Group of the Omineca Belt (Fig. 1.2). He further separated the Snowshoe Group from the last three (collectively called the Cariboo terrane: Struik 1985a, 1986b) by the westerly directed Pleasant Valley thrust (Fig. 1.2; Struik 1983a; see Chapter 2.1). The Pleasant Valley thrust is apparently truncated by the Pundata thrust.

The Eureka thrust is the name given by Struik (1983b) to the thrust southeast of Prince George (Figs. 1.2, 2.3) separating the Quesnel terrane from the Snowshoe Group, and is thus equivalent to the author's Quesnel Lake Shear Zone into which it passes to the southeast.

As stated in the previous sub-section, the author would intuitively link the Eureka thrust (= Quesnel Lake Shear Zone) with the Pundata thrust (Rees 1981); the Antler Formation in the upper plate of both thrusts is thus a single sheet which was emplaced northeastwards and later folded by D_2 and D_3 , probably syn-kinematic with the Pleasant Valley thrust since that is also (south)westerly-verging (cf. Murphy 1985). At present, it is not clear if the model of Struik (1986b) is compatible with the author because it is ambivalent on some crucial points:

* (p. 1050) whether or not the "flat Pundata thrust" and its upper plate are involved in the regional southwesterly-verging folding (which the Eureka-Quesnel Lake Shear Zone is deformed by). If they are not, then the Pundata thrust must post-date the folding and therefore the obduction of the Quesnel terrane. Therefore they are not the same allochthon (indeed, Struik 1986b, p. 1053, notes that they "are not laterally continuous"); so where and when did the Pundata thrust sheet.

come from?

Alternatively, if the Pundata thrust is folded, it could be the same allochthon and the same emplacement event as the obducted Quesnel terrane (this thesis; Struik 1986b, p. 1050), but that means that D_1 obduction post-dates the Pleasant Valley thrust because the Pundata thrust truncates it. This in turn would mean that the Pleasant Valley thrust has no kinematic relationship with the similarly-verging regional folding.

* Unfortunately, Struik (1986b) has no conclusion on the relative timing between the folding and the Pleasant Valley thrust. But his comments on p. 1059 strongly imply that they are syn-kinematic.

The outcome of this is that at present the author cannot reconcile his model with that of Struik (1986b) because there are crucial uncertainties in the latter, particularly concerning the age and tectonic significance of the Pleasant Valley thrust.

(17) Magmatism generated by the obduction-subduction

There was no subduction-zone magmatism related to obduction presumably because the rocks that were subducted beneath the Quesnel terrane did not approach the depth necessary for partial melting (ca. 100 km). (Compare the tectonic model of Tempelman-Kluit (1979) in this regard - see Section 3.)

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(18) Post-obduction deformation: D_2 and D_3 deformation

After D_1 obduction, the coupled Quesnel terrane and the Omineca Belt were folded by southwesterly-verging folds of the second and third

deformation phases, D_2 and D_3 . Fig. 6.9 is an interpretation of the crustal structure resulting from D_2 and D_3 . Each of the main components of the cross-section, the Quesnel terrane (including the Antler Formation), the sedimentary cover, and the basement, is area-balanced with its antecedent counterpart in Fig. 6.6. Plane strain is assumed (but see later sub-section 21). Fig. 6.9 was built up by projecting the cross-sections of this thesis and of other workers (Bailey 1978, Campbell 1971, Fletcher 1972, Struik 1982b, Murphy 1985) into it, and interpolating and extrapolating the remaining structure. An average plunge of 10° to the northwest was used; this is at the low end of the range of reported plunges of fold hinges. The folds at the Intermontane-Omineca boundary southeast of the thesis area are probably somewhat conical, so some modification of their style was applied in their projection.

At high structural levels such as in the Black Stuart synclinorium (Struik 1982b, 1986b), D_2 crustal shortening by southwesterly-verging deformation took the form of reverse faulting and thrusting in brittle and quasi-ductile rocks. In deeper, ductile rocks, such as in the thesis area, overturned and recumbent folds and nappes formed. The northwestward plunge of F_2 folds could have been produced at the same time as the folds, or later as a result of post- D_2 differential uplift.

The rocks in the thesis area were at a maximum depth of about 20 km in D_2 time, based on a pressure of 550 MPa obtained from metamorphic rocks (interpreted by the author to have 'equilibrated' in D_2 time) in the Mt. Stevenson area (Getsinger 1985; see Chapter 4.2).

Mt. Borland anticlinorium. This is an informal name, introduced herein, for the major overturned F_2 anticlinorium structure that is inferred to exist in the northeast of the thesis area and beyond. This

Fig. 6.9

Hypothetical cross-section for the western Omineca Belt and structurally overlapping Quesnel terrane at about 165 Ma (latest Middle Jurassic), showing southwesterly-verging D₂ (and D₃) structures. Symbols as for Fig. 6.2, 6.6, plus: squiggles = Cache Creek terrane, x = Jurassic intrusion; RMT = approximate position of west side of present Southern Rocky Mountain Trench. Basement, sedimentary cover, and Quesnel terrane are each area-balanced with their counterparts in Fig. 6.6; thus, plane strain is assumed here for D₂. The length of the Quesnel Lake Shear Zone equals its unfolded length in Fig. 6.6. Basement is 150 km wide compared with 240 km in Fig. 6.6; thus D₂ horizontal shortening is depicted as being approximately 37.5%.

Upper part of section is constrained by presently exposed structures, utilizing up and down plunge projection. Form lines in heavier ink are taken from existing cross-sections projected into section; form lines in lighter ink are interpolations by the author. Surface is constrained by estimated pressures of metamorphic rocks at the level of the thesis area (roughly in centre of cross-section; cf. Fig. 2.2), which were quenched at this time.

Lower part of section, depicting mode of basement shortening, is conjectural. It is assumed that basement contracted telescopically by reverse movement on ?rift-related ductile faults; the deeper basement shortened and thickened quasi-homogeneously. Although most of these faults originally dipped westwards, they were steepened during D₁ such that the potential for them to generate northeastward thrusting into the cover during D₂ was significantly reduced.

A quasi-decollement (really just a high strain zone) developed near the basement-cover contact; this partial ungluing allowed the basement and cover to deform consanguineously but via contrasting styles (cf. Jackson 1980). Thickening of the basement and cover created an uplift component which, - added to the continuing convergence between the Quesnel terrane and Omineca Belt - produced the southwestward rotational strain in the cover; northeastward dipping basement faults may have helped to establish this sense of shear (cf. Murphy 1985). Admittedly, the same conditions might have promoted an overturning to the northeast; the deciding factor was the downward deflection of the Quesnel terrane.

The degree of intertonguing between basement and cover has been minimized in order to keep a tight rein on the area balance between this cross-section and that in Fig. 6.6. The intertonguing may be much more pronounced than shown: possibly the larger nappes are basement-cored. The number of nappes at deep levels in the cover in the northeast is speculative: there could be fewer, larger nappes (approaching the scale of the Scrip nappe - Raeside & Simony 1983) or more, smaller folds.

fold shows up clearly in Fig. 6.9; it dominates the structure of the northwestern Cariboo Mountains, in a position similar to the Lightning Creek anticlinorium (Fig. 1.2; see below). The location of its axial surface trace is not known but it must occur northeast of the Mt. Borland area (Fig. 2.1), the most northeastern locality in the thesis area where a stack of downward-facing F_2 structures were observed; they begin immediately north of the Badger Peak syncline (cross-section I-I'-I''). The existence of such a large apparent (horizontal) thickness for the overturned limb of the anticlinorium (at least 7 km north-northeastwards from Badger Peak: Map 1) is supported by the fact that the Antler Formation does not re-appear in the thesis area, nor in fact does it re-appear until the Black Stuart synclinorium.

This major F_2 anticlinorium is not the same thing as the Lightning Creek anticlinorium (Fig. 1.2), as described by Campbell et al. (1973), nor is it a re-interpretation of it. The Lightning Creek anticlinorium, and related structures like the Black Stuart synclinorium, the Lanezi Arch and the Isaac Lake synclinorium, appear to be fairly open, upright warps, and as such are probably F_3 . [Struik 1985b, in his Figure 8, shows a Lightning Creek anticlinorium "third phase fold" deforming both westerly-verging F_2 folds (= author's F_2 also?) and an earlier S_1 "bedding-parallel cleavage" (= author's S_1 also?).] But it is not clear how much of their form and surface traces they owe to older, quasi-coaxial recumbent F_2 structures like the Mt. Borland anticlinorium that they may have nucleated upon.

(19) Timing of D_2 and D_3 deformation and metamorphism

The available data in the Quesnel Lake area is summarized as follows.

(a) D_2 deformation post-dated the obduction of the Quesnel terrane which was completed by about 180 Ma (early Middle Jurassic; see subsection 12).

(b) Granodiorite stocks in the Cariboo Mountains 80 km to the east of the thesis area were intruded between deformation phases F_2 and F_3 of Pigage (1977) which are interpreted to be equivalent to the author's D_3 and post- D_3 (Table 6-3). The stocks are dated at 163 ± 7 Ma by the Rb-Sr whole-rock method (Pigage 1977). If the correlation of deformations is correct, this date provides a minimum age for D_2 and D_3 (163 Ma is latest Middle Jurassic).

(c) Mineral ages of 174 ± 4 Ma and 161 ± 15 Ma from the Quesnel Lake Gneiss were interpreted by Mortensen et al. (in press) and Armstrong (pers. comm.) to mark the time of resetting in the gneiss due to metamorphism (see Chapter 2.2), which was syn-kinematic with respect to D_2 .

Additional data from farther southeast in the Omineca Belt may be cited here on the ages of intrusions which are syn-tectonic with regional metamorphism and deformation which are likely equivalents of D_2 in the Quesnel Lake area.

(d) Metamorphic zircons from the Adamant pluton in the Selkirk Mountains have been dated at 169 ± 4 Ma (Shaw 1980). The pluton suffered its peak metamorphism between the local D_2 and D_3 deformation phases which, on the bases of structural similarity and palinspastic reconstructions given in Brown et al. (1986), are probably approximately equivalent to D_2 and D_3 in the thesis area and the Cariboo Mountains.

(e) In the northern Kootenay Arc, the Kuskanax batholith (Fig. 2.3)

intrudes Paleozoic Omineca Belt rocks, and the Quesnel terrane. It was emplaced prior to or during the second phase of deformation (was the first phase related to the emplacement of the Quesnel terrane?) and has a U-Pb zircon age of 173 ± 5 Ma (Parrish & Wheeler 1983).

(f) Also in the Kootenay Arc, intrusions which were syn- to late-kinematic with ductile deformation involving both the Omineca Belt and the Quesnel terrane (e.g. Nelson batholith, Fig. 2.3) are dated at between 175 and 165 Ma, with metamorphism continuing to at least 156 Ma (Archibald et al. 1983).

Accepting various assumptions about this data, such as that deformation, metamorphic and intrusive events were not significantly diachronous throughout the southeastern Omineca Belt, then D_2 probably began sometime before 175 Ma, but after the Quesnel terrane allochthon was fully in place, at roughly 180 Ma.

It will be postulated later that there was no major hiatus, if any, between D_2 and D_3 , the main distinction between them in the thesis area being the transition between prograde metamorphism and the onset of cooling. Most workers in the Quesnel Lake region have agreed that D_3 (i.e. their designated equivalent phases) post-dated the metamorphic peak, although locally metamorphic recrystallization and plutonism dated at about 165 Ma (see above) continued into D_3 and beyond.

In summary, D_2 deformation probably began between 180 and 175 Ma, probably less than 5 Ma after the completion of obduction. Deformation probably continued with little or no break through D_3 to about 160 Ma.

(20) Origin of the L_2 axial extension

One way to model the formation of a thick pile of overturned folds by horizontal compression, in which the foliation and fold axial surfaces suffer rotation by progressive deformation, is as a heterogeneous plane strain simple shear deformation of the original stratigraphy. As a first approximation, this model may be applied to D_2 ductile deformation of the Omineca Belt and overlying Quesnel terrane. The axial surfaces of E_2 folds and the S_2 foliation would represent approximately the local XY plane of finite strain; fold axes would be parallel to the Y axis (subhorizontal, northwesterly), and the transport or vergence direction of the folds would lie close to X in the vertical XZ plane.

This model has been applied by Murphy (1985) in the eastern Cariboo Mountains to the east of the thesis area (Fig 1.2), where the axial surfaces of D_2 -equivalent folds change from open and near-vertical in shallow structural levels to northeasterly-dipping, and isoclinal and recumbent at deep structural levels. Murphy (1985) explained this as the result of a depth (temperature) controlled gradient in the simple shear component of the strain path, the deeper rocks undergoing a greater amount of southwesterly directed shear and flattening because of their greater ductility.

While the role of large-scale simple shear seems self-evident, the deformation was more complicated than that. For example, referring to the eastern Cariboo Mountains again, Murphy (1985) had to invoke a simultaneous component of pure shear flattening (coaxial horizontal shortening with a vertical X direction) to account for a large amount of thickening of his stratigraphy, and to help account for the steepness of F_2 axial surfaces at shallow crustal levels.

The complication in the author's area, however, is the L_2 axial extension, the principal finite extension in quartzofeldspathic rocks (excluding the Quesnel Lake Gneiss) wherever one is observed (Fig. 6.10). The strain geometry and the magnitude of the extension were not quantified by the author but this has been analysed by Struik (1980). From stretched clasts in conglomerates he found that most strain ellipsoids, including the calculated mean, are of the constrictional type with the X axis parallel to the regional strike and the fold axes, assuming no volume loss; if a volume loss is invoked, which Struik inferred is reasonable, the constrictional character is enhanced. Struik (1980) estimated an extension of 40% in the conglomerate. From another grain-shape analysis in the Kootenay Arc, Ellis (1986) calculated a 120% hinge-line extension.

In an ideal simple shear model, the principal finite extension should be perpendicular to fold axes (except after very high strains, when they may become parallel), not parallel to them, as is L_2 . This is a problem which is encountered in fold belts all over the world (see Hobbs et al. 1976, p. 283); the reason for it may be a combination of many different processes. Two general approaches to the problem are discussed below: (A) and (B) involve the kinematics of the bulk strain regime, and (C) concentrates on folding mechanisms.

(A) A combination of simultaneous simple shear (ss) and pure shear (ps) operated, such that they are partially coaxial (they share one axis of incremental strain; Fig. 6.11).

The plane strain simple shear incremental extension (X_{ss}) is normal to the fold axes, and there is no extension in the Y_{ss} direction (northwest-southeast), parallel to fold axes. As will be discussed

Fig. 6.10

Sketch diagram of general D_2 finite strain fabric (a), and schematic geometry (b). The question is how to explain and reconcile these properties in terms of a kinematic regime, and ultimately in terms of a tectonic model.

Fig. 6.11

Model (A): Pure shear and simple shear components of the instantaneous strain regime. Diagram is schematic; axial ratios are qualitative only. Note that the pure shear and simple shear are partially coaxial in that X_{ps} and Y_{ss} , and Y_{Zps} and X_{Zss} are coincident. This is not transposition.

later, this component of the strain regime results from some mode of underthrusting from the southwest.

The pure shear incremental extension (X_{ps}) is horizontal and parallel to the fold axes (northwest-southeast) and to Y_{ss} ; the incremental Y_{ps} is assumed to be vertical. The XY_{ps} plane is thus vertical, which agrees with the model of Murphy (1985) except that he inferred that X_{ps} was vertical. The incremental Z_{ps} axis is therefore related to a layer parallel sub-horizontal shortening from northeast to southwest. The difference between the incremental Y_{ps} and Z_{ps} is inferred to be small, i.e. the K factor is high and the incremental strain is of the constriction type. Note that even without the accompanying simple shear component, folds would form under this pure shear component with their axes northwest-southeast, parallel to the X_{ps} axis - it is simply a function of the initial attitude of the layering with respect to the strain axes.

The essence of the matter now is the resultant of these two simultaneous components of strain, which is mainly a function of the ratio of their strain rates. It is possible to model physically, mathematically or by computer the geometry and kinematics of the strain path resulting from a range of simultaneous simple shear and pure shear strain rates (e.g. Ramberg 1975, Ghosh & Ramberg 1976, Coward & Kim 1981, Sanderson 1982). This was not done in this thesis (incidentally, the modelling studies referred to are not directly applicable to the present problem, because they rarely venture away from two-dimensional treatments). The author's conception of it is very qualitative and is visualized as a 'general rotational' strain imposed on initially horizontal layering, with the X axis horizontal, trending northwest-

southeast, and the incremental flattening plane ($XY_{ps/ss}$) lying between vertical and a 45° angle, dipping northeast. Folds form with their axes parallel to X_{ps} . The non-coaxial (simple shear) component induces a rotation of the finite $XY_{ps/ss}$ plane (S_2 foliation, F_2 axial surfaces) about Y_{ss} ($= X_{ps}$), causing the folds to flatten further and tighten and overturn towards the southwest. The YZ_{ps} plane corresponds to the XZ_{ss} plane. Although there is an instantaneous simple shear extension (X_{ss}) at 90° to the pure shear extension (X_{ps}), it does not develop because it is constantly 'buffered' by the shortening due to the pure shear quasi-constriction. Similarly, rigid body rotation of competent fold limbs causes them to roll through a range of incremental strain fields. Only the X_{ps} ($= Y_{ss}$) direction is continuously additive. [Pure shear is a much more efficient method of building finite strain than is simple shear, which is probably the least efficient method (Pfiffner & Ramsay 1982); this is probably why the overall finite X corresponds to that of the pure shear component (X_{ps}) - the simple shear component is primarily responsible for the rotation of the XY plane - that is, the overturning of the folds.]

Conversely, the pure shear extension (X_{ps}), is not affected by the simple shear component because it is parallel to a direction in simple shear of no incremental longitudinal strain (Y_{ss}); in other words, it is always additive. Sanderson & Marchini (1984) pointed out that the finite extension under combined simple shear and pure shear (transpression, in their case) can "swap" from one orientation to one 90° away, as the 'dominance' of one regime is superseded by the other regime with progressive deformation. They went on (p. 451): "the axis swapping is very interesting since it indicates that the maximum finite stretch may be normal to the maximum incremental stretch". In the case

under discussion here, the finite extension at early stages of the deformation may indeed have been in XZ_{ss} , but with rotation (passive or active) it got closer to the maximum shortening direction of the pure shear component (Z_{ps} , subhorizontal), and was gradually subtracted. Eventually (probably quite quickly) the extension parallel to X_{ps} emerged as the finite maximum. A corollary is that there is no simple correlation between finite strain and the amount of work done in deforming a rock, as remarked by Sanderson (1982).

Could the simple shear and pure shear components be superimposed rather than simultaneous strains? This is rejected because the results of two superimposed strains are quite different from a single simultaneous combination. For example, stretching lineations produced by the first strain would be re-oriented and folded by the second strain. There would also be, probably, abundant evidence of superposition in the form of crenulations, or two distinct fabrics.

(B) Oblique convergence or transpression. This is a modification of (A) above, in which the simple shear component is oblique to the pure shear quasi-constrictional component (Fig. 6.12), instead of transverse to it (as in (A)). They share no axes of incremental strain. This is analogous to transpression (Harland 1971, Sanderson & Marchini 1984).

The sense of obliquity is such that the D_2 underthrusting generating the southwesterly-verging simple shear was more northerly directed than northeasterly directed. This sense of obliquity is justified below.

Another way of describing this, which might be helpful, is as the addition of a second component of simple shear to the strain regime of (A). This simple shear has its Y axis horizontal and normal to X_{ps} and

Fig. 6.12

Model (B): Instantaneous strain regime comprising model (A) pure shear and simple shear (above - from Fig. 6.11), plus another component of horizontal simple shear (below, left); top to the southeast. This overall strain regime corresponds to a quasi-constrictional flattening of the sedimentary cover, combined with an oblique, southwesterly 'subduction' shear; this is depicted in Fig. 6.13.

the fold axes; its sense of shear is upper surface towards the southeast. All we have done is resolve the oblique subduction or underthrusting of model (B) into a transverse southwesterly-verging simple shear (as model (A)) and a southeasterly-verging simple shear (Fig. 6.12).

The sense of obliquity is inferred for two main reasons. First, there is a conspicuous en echelon trend to the axial surface traces of the major D_2 structures at the Intermontane-Omineca boundary between latitudes 52° and 53°N (Fig. 1.1). The en echelon pattern implies an oblique, more northerly-trending underthrusting (Harland's 1971, Figure 3 illustrates the development of en echelon folds via transpression). Second, the vergence of the syn-kinematic microfabrics in L_2 (see Chapter 3.3) - not widely observed, but consistent - indicates an 'upper surface to the southeast' sense of shear.

It is further suggested that this additional simple shear enhanced the tendency for the finite extension lineation (L_2) to 'stabilize' in the observed direction.

Ellis (1986) has also noted the parallelism of stretching lineations and syn-metamorphic (Jurassic) fold axes in the southeastern Canadian Cordillera. He concluded from this that the formative kinematics were oblique to the orogen, and that oblique (dextral) subduction of the North American plate beneath allochthonous terranes was responsible for the strain geometry (cf. Fig. 6.13). He didn't explain why the folds do not verge in the same direction that they were stretched, as has been the main concern of the present author in this sub-section.

(C) Stretching parallel to fold axes may be due to non-cylindrical folding. As culminations and (or) 'surges' grow out of cylindrical

Fig. 6.13

Summary diagram of the postulated overall D_2 strain regime for model (B), namely a combined horizontal flattening and oblique shear, known as transpression. Lower diagram shows a hypothetical oblique subduction situation which might induce transpression in the overriding plate (although the author is not proposing this plate tectonic situation - it is just an analogy).

folds, the original axial length of the fold is bound to increase. [By 'surge' it is meant that the fold is distended in the subhorizontal direction, approximately the direction of vergence; this corresponds to the incipient stage of a sheath fold. It is also known as 'arcuation'. A culmination is a vertical distension of the fold.] Thrust sheets and folds can also deform internally, independent of the boundary conditions, by moving over an irregular surface such as a lateral ramp (Coward & Kim 1981). As a result of these potential modes of ductile extension, a stretching lineation would form, especially in competent quartzofeldspathic rocks. It would be confined to the layering and would remain in the finite flattening plane.

To demonstrate this hypothesis, it would be necessary to show that the observed amount of axial extension is accounted for by the mode and scale of distension that is invoked to explain it. The existence of a strong axial extension in a region of cylindrical folds might be hard to explain by this hypothesis. However, in support of it, it was mentioned in Chapter 3 that in the southeast of the thesis area, the F_2 northwestward plunge is the steepest that was encountered (usually 15 to 40°), while the L_2 axial extension is also more prominent here than elsewhere. This coincidence suggests a genetic relationship, that the rocks were stretched parallel to the F_2 axis when the plunge culmination in the Mt. Brew area formed.

Conclusion. It is concluded that the L_2 axial extension could represent a bulk longitudinal pure shear extension as a component of the overall D_2 strain regime which also included transverse or oblique simple shear; however, it could have been produced simply by more local stretching due to the non-cylindrical distension of folds. The relative

importance of these two possible mechanisms should be investigated further, as it has important tectonic implications (see sub-section 21).

Lastly, a few more interpretations concerning L_2 . The processes responsible for L_2 , discussed above, affected all the rock units in the thesis area, but had limited effect on L_1 in the Western Quesnel Lake Gneiss. L_1 in the gneiss was at least partly re-oriented, but in general the competent gneiss appears to have resisted L_2 extension, especially in the gneiss north of Sellar Creek where a pristine L_1 is preserved at about 90° to the regional L_2 .

One reason for the resistance and survival of L_1 in the Western Quesnel Lake Gneiss is that it was more difficult to dynamically recrystallize (during D_2) a megacryst alignment than it was to dynamically recrystallize deformed (D_1) quartz elasts or an L_1 mineral lineation.

The several northeast-trending F_2 minor fold axes observed in the Badger Peak area (Fig. 3.6d) deserve a comment. It is suggested that prior to D_2 , S_{0-1} in this region was deformed by high strain into northeasterly-trending F_1 folds with their axes parallel to L_1 in the (then) suprajacent Western Quesnel Lake Gneiss (cf. Sanderson 1973). Some domains of these folds would thus have vertical strata striking northeast-southwest. With F_2 recumbent folding about northwest-southeast axes, this package was flattened and folded so that F_2 minor folds formed with their axes still northeasterly-trending. The L_2 axial extension wasn't as effective in re-orienting these F_2 folds into a northwest-southeast trend as it was in the rest of the thesis area because of the competence of the (by now) subjacent Western Quesnel Lake

Gneiss, and many of these odd minor folds are preserved.

(21) The large-scale significance of a possible longitudinal extension of the D₂ rockpile

The conclusion of sub-section 20 was that L₂ could have been produced at least partly by an absolute extension of the D₂ rockpile via a component of quasi-constrictional shortening. Stretching lineations parallel to F₂ fold axes are widely documented in the southern Omineca Belt, so the implication is that unless there was a way of absorbing this stretching internally, the entire region sustained a finite longitudinal extension. In other words, when the continental margin was shortened in the order of 100 km (see sub-section 18) only part of the resulting extension was taken up by vertical thickening, and the remainder was squeezed out longitudinally. Presently there is no structural evidence for a complementary longitudinal shortening, such as northeast-southwest cross-folds with vertical fabrics.

In invoking this longitudinal extension, this model differs from most established ones which assume that crustal shortening approached plane strain. Plane strain may be warranted in the case of deformation characterized by systematic faulting such as piggyback thrusting where there are well defined displacement vectors (stretching lineations in the direction of fold, nappe or thrust transport); otherwise it is only assumed as a convenience in balancing and restoring structural sections. In the case of D₂, plane strain is probably not justified because it is not characterized by simple shear-like discontinuities. Any component of the strain regime which was oblique to the finite strike of the orogen (e.g. model B in the preceding sub-section; Ellis 1986) would automatically nullify the premise of plane strain. In conclusion, while

plane strain may be a useful simplifying assumption, it should not be rigidly applied to D_2 on a large scale in the southern Omineca Belt because of the existence of hinge-parallel stretching lineations.

L_2 extension versus the assumption of plane strain in Fig. 6.9. The postulate that there was extension parallel to F_2 fold axes throughout D_2 conflicts with the assumption of plane strain utilized in the balancing of Fig. 6.6 with Fig. 6.9. However, all it means is that some extra basement and sedimentary cover must be added to the Omineca Belt in Fig. 6.6 to compensate for material 'extruded' out of the plane of the cross-section in Fig. 6.9 in D_2 . Since it is not necessary to constrain the width of the Omineca Belt in Fig. 6.6, an arbitrary increase is not prohibited.

Further implications. The implication that the southern Omineca Belt is somewhat longer than it was before D_2 should not be unacceptable because we don't know how long it used to be - or what shape it was. Perhaps the (meta-)sedimentary prism formed some kind of salient which was not particularly 'confined' at its northwestern or southeastern end: in D_2 it underwent shortening combined with one or two components of shearing (sub-section 20) and was extended longitudinally (or 'laterally extruded') as well as vertically thickened.

A crude analogy is drawn with the Himalayas: it is now known that east-west 'lateral extrusion' of parts of the Himalayas has occurred and is continuing as India inexorably ploughs its way northwards into Asia (Tapponier et al. 1986); presently it absorbs more than half the convergence while underthrusting absorbs the remainder. The analogy is not pressed too far because geometrically the Himalayan longitudinal extension is effectively strike slip rather than constrictional.

(22) Tectonic model for D_2 : Origin of southwestward vergence

D_2 axial surfaces and foliations have low dips which, assuming horizontal compression, implies a non-coaxial component to the strain path. This simple shear component of the D_2 kinematic regime was produced by a relative underthrusting of rocks that lay to the southwest. In simple words, first the Quesnel terrane was overthrust relative to the Omineca Belt (D_1), and subsequently partially underthrust relative to the Omineca Belt (D_2). The questions that arise are (Fig. 6.14):

(a) How are D_2 southwesterly-verging structures accommodated at depth: was the basement deformed synthetically, or was there another process?

(b) Does the answer to (a) require that the tectonic vergence between the Quesnel terrane and the continental margin ultimately changed between D_1 and D_2 , and if so, why?

This is a complex issue, and only preliminary and tentative comments are made here.

The D_1 obduction of the Antler marginal basin and the Quesnel terrane stopped at about 180 Ma when that particular locus of convergence locked up. The 'locking-up' may have been caused by the main mass of the Quesnel terrane, with its thick volcano-plutonic root and (or) basement, abutting against the leading edge of the Omineca Belt continental or transitional crust basement, which refused to be deflected downwards any more (Fig. 6.6). Alternatively or additionally, the lower plate was thickened during D_1 , such as by reverse faulting on reactivated basement faults and by northeasterly-verging folding (Fig. 6.6), to an extent that decreased its mean density, and increased its

Fig. 6.14

Schematic cartoons showing the possible ways of forming southwesterly-verging D_2 structures. For explanation of (i) and (ii), see text. v = obducted Quesnel terrane, + = basement, unpatterned = sedimentary cover.

buoyancy and its resistance to further subduction beneath the Quesnel terrane (this is similar to the model of Archibald et al. 1983, p. 1911).

Whatever the reason(s), horizontal shortening could no longer continue by obduction after about 180 Ma. So did convergence continue via northeasterly-verging thrusting on another surface(s) propagating beneath the allochthon, carrying the allochthon piggyback-style, and thus inducing that vergence to continue through the remainder of the Mesozoic? If so, the answer to question (b) above would be "No", there was no D_1 - D_2 vergence reversal. Indeed, a major easterly-verging detachment, the Monashee decollement, was active in the Middle to Late Jurassic (165 to 155 Ma) farther south (Read & Brown 1981, Brown & Read 1983, Journeay & Brown 1986) and again in the Late Cretaceous when a basement duplex began to form, feeding into the folds and thrusts of the Foreland Belt (Brown et al. 1986, Journeay & Brown 1986). Although this basement duplex is not exposed in the Cariboo Mountains, it is probably present in some form at depth, since it is this thrusting that is responsible for ca. 125 km of thrusting in the adjacent Foreland Belt (Murphy 1985, Brown et al. 1986; it is not shown in Fig. 6.9).

From this review of the circumstantial evidence, it would seem that shortening of the North American margin was characterized by (north)easterly-verging thrusting, rooting at deep levels, in the crust or upper mantle to the (south)west, from Early-Middle Jurassic obduction through to the early Tertiary, progressively stepping towards the continent. Question (a) above is really asking whether (i) the D_2 southwesterly-verging deformation can be explained in the context of this dynamic system, in which case it is antithetic to basement shortening, or whether (ii) D_2 is an anomalous episode in which basement

shortening was synthetic to it, and unrelated to the predominant sense of crustal shear during the Mesozoic. Alternatives (i) and (ii) will now be discussed.

(i) Tectonic wedging or delamination. In this scenario, a deep-rooted thrust or flat fault begins to propagate eastwards within the Omineca Belt, well below the obducted Quesnel terrane. Because it is still blind, its upper plate cannot be carried passively because the front end is pinned, and a simultaneous counterflow must occur in the upper plate to balance thrust motion. If the counterflow takes the form of a discrete, but oppositely verging thrust emanating from the toe of the blind basal thrust, a kinematic wedge is produced, analogous to the triangle zone at the toe of the Foreland Belt (Price 1986). If the counterflow is instead distributed throughout most of the upper plate, antithetic (south)westerly-verging folds would result. The upper plate acts as an indenter 'delaminating' the crust in its path by forcing the upper part of it to flow in the opposite sense to the basal thrust (Fig. 6.14c).

This idea has been proposed by Price (1986) to explain D_2 folds in terms of a sustained eastward propagation of initially blind deep-rooted flat faults. In this model, which corresponds to Model 1 of Murphy (1985), (south)westerly-verging deformation is not a deviation from (north)eastward vergence, but a by-product of it. The model could possibly explain D_2 folding of the Intermontane-Omineca boundary as such is represented in the thesis area. However, westerly-verging folds, interpreted to be the contemporaneous counterparts of the F_2 folds of this thesis, occur in the upper plate of the Monashee decollement (Selkirk allochthon); their formation not only pre-dates the early

motion on the decollement, but the shear zone that separates the allochthon from the lower plate has an upper plate to the east sense of shear (Brown & Murphy 1982) - opposite to the shear sense that would be required to form the folds by tectonic wedging. If these D_2 folds were formed by wedging, it must have occurred prior to the Monashee decollement, and far to the west of the folds' present position. This is feasible, but then the Monashee decollement would have to be a flat fault which cut through the already deformed and metamorphosed D_2 rockpile at a much higher level than the basal fault of the 'wedge' that originally formed the folds, and then carried them piggyback-style eastwards. This is not easy to imagine, and deviates from the predicted pattern of migration of deep-level faults towards the foreland.

In conclusion, D_2 folding in the thesis area might have been produced by tectonic wedging, but no (north)easterly-verging basal thrust is exposed to help demonstrate the process. Even if there was, it would have to be coeval with D_2 to be relevant here. The best-documented easterly-verging basement-rooted fault that is exposed in the Omineca Belt, the Monashee decollement, post-dates D_2 folds, and indeed no significant antithetic folds (D_2 or otherwise) can be attributed to its passage through the crust; during at least the later stages of its movement, it was probably not even blind. The Monashee decollement does little to support the concept of tectonic wedging.

(11) Basement shortening synthetic to D_2 structures. In this scenario, basement shortening feeds directly into (south)westerly-verging folds in the cover in a synthetic sense. This idea corresponds to Murphy's (1985) Model 2, in which he postulated the eastward subduction or underplating of the outer part of the continental or

transitional crust basement (cf. 'A-type subduction' of Bally 1981). This discontinuity propagates into the sedimentary cover producing southwesterly-verging rotational deformation. The style of deformation varies with depth due to gradients of temperature and of combined pure shear and simple shear, such that there is a transition between the suprastructure and infrastructure (Murphy 1985). Murphy concluded that bulk crustal shortening was greater at deep levels than at shallow levels, requiring that they be detached. He calculated that there was approximately 50-65 km of northeastward underthrusting of the infrastructure relative to the suprastructure beneath the Cariboo Mountains, inferring that the detachment between them was the Pleasant Valley thrust of Struik (1982a,b; Fig. 6.9). That this could be such a significant fault would account for the problem of stratigraphic correlation across it from northeast to southwest, as mentioned in Chapter 2.1.

Preferred model. A variation of (i) and (ii), i.e. Murphy's (1985) Models 1 and 2, is preferred by the author in which the basement contracts by reverse movement on numerous listric normal faults which were originally produced by rifting of the crust in the Proterozoic or upper Paleozoic (Fig. 6.9), instead of on just one major fault. Uniformly shortening and thickening the basement like this would cause a gradual isostatic rebound. Jackson (1980) invoked this process (in another context) to explain crustal shortening, avoiding the 'room problem' of duplicating a thick basement:

"The basement tends to return to its original thickness while the overlying sedimentary column, having been deposited on an extended basement, is forced to take up the shortening by folding."

Paraphrasing now:

"This process does not necessitate either abnormal thickening (uplift) or subduction of continental crust in the early stages of continental collision". (my parentheses and underlining)

This model for D_2 explains better the development of southwesterly-verging folds below as well as above the Pleasant Valley thrust (Fig. 6.9). In more detail, if the cross-section is imagined to be pinned in the northeast, then the contraction of the basement towards this pin would drag or underthrust the cover northeastwards with it, there being a high strain zone or decollement between them. (Jackson 1980 also invoked a "decoupling layer" between the contracted basement and the folded cover.) The remaining thing to explain is the sense of shear in the cover: the author infers that it was actually produced by the upward rebound of the continental margin (see above) such that the displacement vector of the Quesnel terrane relative to the Omineca Belt was deflected downwards after 180 Ma instead of upwards as before (obduction), and the sense of rotational deformation in the cover reversed from that of D_1 . Reverse faulting on any continentward-dipping basement faults would have helped to establish the (south)westward vergence; however, most basement faults would probably be steep or westerly-dipping, so while shearing in the basement effected its shortening it probably did not contribute directly to the vergence of the cover.

Incidentally, perhaps it was only after the basement re-achieved its pre-rifting thickness that it was strong enough to be duplicated en masse (cf. Jackson 1980); this might explain why the early phase of movement on the Monashee decollement (ca. 165 Ma: Journeay & Brown 1986) so closely follows D_2 .

Comparison is made with the Western Alps: the Helvetic nappes are a

series of decollement nappes which are digitations in one quasi-coherent thrust mass (Trumpy 1969). The basement massifs are not significantly involved in the nappes (unlike the Pennine nappes), but movement between them caused the shortening and partial detachment of the sedimentary cover.

In conclusion, the author prefers to explain D_2 deformation in terms of continued convergence of the Quesnel terrane towards a continental margin that was simultaneously rebounding due to its regenerated buoyancy. The basement contracted 'telescopically' by reverse movement on listric faults that originally attenuated it during one or two prior phases of crustal rifting, rather than by large-scale antithetic or synthetic basement shearing. This model forms the basis of Fig. 6.9, and is similar to that of Archibald et al. (1983). A simultaneous eastward propagation of blind, deep-level flat faults may have accompanied the process, effecting a form of kinematic wedge (Price 1986), but this process was probably not ultimately responsible for the bulk of southwesterly-verging deformation in the cover.

(23) Relation of this tectonic model to the origin of the L_2 axial extension

The preferred tectonic model in the previous sub-section is compatible with the explanation for L_2 given in sub-section 20 because neither involves any strong or localized southwesterly-verging shearing, which is desirable because of the lack of vergence parallel D_2 stretching lineations. Instead the basement contracts uniformly all along its formerly attenuated width with a dominant layer-parallel pure shear flattening - hence the XYZ_{ps} component invoked earlier. In contrast, the basement thrusting envisaged in alternative models (Fig.

6.14c,d) would be expected to generate a lot of simple shear, probably concentrated in discrete shear zones.

(24) D₃ deformation and recrystallization

Third phase deformation in the thesis area produced southwesterly- and northeasterly-verging crenulations, but very little recrystallization and foliation. It took place during thermal relaxation over the entire area, when the rocks were cooling but still ductile.

Given the similar geometry, vergence and near-coaxiality of F_3 and F_2 , it is considered that there was very little hiatus between D_2 and D_3 , and perhaps none (see sub-section 26). Thus, it is probable that the dynamic processes responsible for southwesterly-verging deformation (see sub-section 22) continued unchanged through from the time the F_2 folds formed when the rocks were hot and prograding, into the phase of cooling (D_3) when strain-related recrystallization subsided and shortening proceeded predominantly by buckling and crenulation of existing S-surfaces. Being cooler the rocks were harder and D_3 shortening was limited to roughly 30%; this figure is based on the typical wavelength and amplitude of F_3 mesoscopic and microscopic crenulations. Field evidence supports this notion of a dynamic continuum between D_2 and D_3 , the essential difference between them being the drop in the attendant metamorphic grade. Montgomery (1985) made the same interpretation with the corresponding deformation history in the area southeast of Quesnel Lake. [This proposal may not apply elsewhere in the Omineca Belt where D_2 and D_3 structures are not kinematically similar.]

Two characteristics of D_3 support the contention that the D_2 - D_3 kinematic regime had a prominent layer-parallel pure shear shortening

component (see sub-section 20 on the L_2 axial extension): one is the presence of both northeasterly- and southwesterly-verging, possibly conjugate crenulations and minor folds, and the other is the steepness of the F_3 - S_3 flattening plane (ca. 60° to 85°).

As the rocks became cooler and harder during D_3 , the rotational or shear component of the strain path diminished and the strain approached a more irrotational path, such that northeasterly-verging crenulations formed as well as southwesterly-verging crenulations.

(25) Regional metamorphism and recrystallization in space and time

General metamorphic processes. The onset of large-scale D_2 folding of the Omineca Belt sedimentary prism (Fig. 6.9) produced unstable thermal inversions and convolutions (cf. Oxburgh & Turcotte 1974) which immediately but slowly began to be smoothed out by the recovery (upward migration) or relaxation (downward migration) of isotherms into a more stable disposition, adjusting to the new arrangement of thermal gradients. This is the "dynamic geothermal regime", referred to in Chapter 4.1, which gives rise to prograde and retrograde regional metamorphism, the details of which depend on the P-T-t path of a particular rock. Eventually it would result in a steady-state pattern of heat flow in which the geothermal gradient would remain more or less constant in a given vertical column.

Metamorphic rocks can quench long before a steady-state isotherm and isograd topography is re-attained. This can happen, for example, due to rapid uplift and erosion (tectonic or surficial) which has the effect of suppressing isotherms, forcing assemblages to quench in whatever 'disarray' they happened to be in, whether deformation had stopped or not. The potential for this is high in upper and middle crustal levels

because the effects of erosion and cooling are felt more quickly there, giving rise to undulose, steeply-dipping, or upside-down isogradic surfaces, and apparently steep thermal gradients. [In deeper crustal levels, the chilling effect of the down-cutting erosion surface is delayed (see England & Richardson 1977, Thompson & England 1984); isotherms do not relax until much later - probably after deformation has finished - and a generally regular and gentle isotherm and isograd topography may be achieved before the rocks quench.] Another way rocks can quench 'prematurely' is if they are rapidly uprooted and transported on a thrust fault, cutting them off from their deep heat sources, and placed onto cooler rocks which extract their heat from them.

The point about isotherm suppression due to erosion, or to rapid cooling due to the juxtaposition of cooler rocks, is relevant to a discussion of the questions posed in Chapter 4.1 and 4.3: why was the M_2 peak diachronous, and what is the explanation for the form and steepness of the metamorphic gradient at the Intermontane-Omineca boundary?

Why was the M_2 peak diachronous? From the relationships between M_2 peak metamorphic minerals and the S_2 strain fabric described in Chapters 3.4 and 4.1, it seems that something shut off thermal recovery and prograde metamorphism in the core of the Spanish Lake anticline during the formation of that fold. In contrast, in and north of the Seller Creek syncline, thermal recovery was not shut off until after that fold formed. The garnet isograd is therefore diachronous with respect to the formation of these two major folds.

What was responsible for this spatial variation in the time of quenching? As mentioned earlier, rapid erosion causes isotherm

suppression, curtailing thermal recovery. Invoking this, perhaps erosion and cooling of the rockpile vertically above the Spanish Lake anticline was very rapid - more rapid than above the Seller Creek syncline. As a result, prograde metamorphism was shut off at an earlier stage in the former than the latter. There are two problems with this explanation: one is that it is unlikely that the erosion rate should vary so much over only about 10 km, unless there was a northeasterly-dipping denudation (extensional) fault between, which can be ruled out here. The second problem is that erosion plays an important role only if it is very rapid (Thompson & England 1984, p. 947), or if the thermal profile achieves a steady-state condition very quickly, making it vulnerable to any enhancement of heat loss by erosion. This is also unlikely: normally thermal steady-state is not approached until a comparatively late stage in an orogenic phase.

The preferred explanation for the quenching diachroneity is essentially the same as that hypothesized by Sleep (1979), and is as follows (Fig. 6.15). Prior to D_2 , isotherms and isograds were probably subparallel to the gently dipping Quesnel Lake Shear Zone. On F_2 folding, isotherms in the Spanish Lake anticline were folded into an antiform and were immediately overlain and flanked laterally to the southwest by relatively cool, mainly basaltic volcanic and volcaniclastic rocks of the Quesnel terrane. In contrast, the Seller Creek syncline isotherms were synformally folded and were overlain (by the Mt. Borland anticlinorium) and flanked to the northeast by a thick pile of pelitic and psammitic metasediments. It is suggested that heat was extracted rapidly from the thermal antiform in the Spanish Lake anticline by the cooler Quesnel terrane that almost enveloped it because the temperature gradient was comparatively high (see later). This

Fig. 6.15

Simplified thermal and structural evolution of the thesis area. Cross-sections cover segment of Intermontane-Omineca boundary represented in thesis area; section in t7 corresponds with mid-crustal segment in Fig. 6.9, and compare also Fig. 2.2. Solid line = Quesnel Lake Shear Zone, dashed lines = isotherms, SCS = Seller Creek syncline, SLA = Spanish Lake anticline, ga = garnet. Cf. Fig. 6.16; see text for explanation.

Note that isotherms are for reference only and do not represent a particular temperature; nor do they necessarily represent the same temperature in each time frame. Arrow pointing up means isotherm is recovering and rock is prograding at that place and time. Small bar across isotherm means it is fixed w.r.t. rock and assemblage is quenching. Arrow pointing down means isotherm is relaxing and rock is cooling. The thing to note is the migration of the bars from SLA to SCS with time, representing diachronous quenching.

curtailed isotherm recovery, and syn-kinematic porphyroblastic minerals did not continue to grow post-kinematically. This is also demonstrated by the retrogression of the peak M_2 assemblages, which began during D_2 in this domain (see Chapters 3.4 and 4.1).

In contrast, the quartzofeldspathic metasedimentary rockpile around and above the Seller Creek syncline provided better insulation (quartz has a relatively low thermal conductivity), being nearly as hot as the rocks beneath it. The heat supply from below was not quickly cut off. Heat flowed laterally southwestwards towards the Quesnel terrane due to the thermal gradient, but because that gradient was lower than in the Spanish Lake anticline, cooling was slower: isotherms did recover somewhat as heat flowed into the thermal synform from the sides, and prograde metamorphism outlasted F_2 folding; syn-kinematic M_2 porphyroblasts continued to overgrow S_2 post-kinematically. Retrogression due to the cooling did not begin until even later, although it was certainly underway by D_3 .

Fillipone (1985) also has suggested that the "rapid temperature drop-off" across the terrane boundary reflects a "heat sink effect" between the "basement and the cover" (= Omineca Belt and Quesnel terrane). He noted that staurolite grade rocks only 600 m structurally below the tectonic boundary in the Boss Mountain area are flanked by chlorite-muscovite assemblages in volcanic and sedimentary rocks above it, indicating "that metamorphic isograds were compressed, and nearly paralleled the margin".

Chamberlain (1986) has described a similar phenomenon from a polydeformed terrane in New Hampshire, where the highest prograde assemblages occur in synclinal basins while the lowest grade assemblages

occur in the anticlinal domes. Anticlines refolded by synclines show cooling overprinted by heating, while synclines refolded by anticlines show heating followed by retrogression. The analysis was based on geothermometry, isograd patterns, textural relationships and garnet-zoning profiles.

If there was so much heat transfer from the Omineca Belt to the Quesnel terrane, how did the steep metamorphic gradient at the boundary survive? The proposed answer to this question follows on from the preceding explanation of the metamorphic diachroneity.

The high heat flow across the folded boundary did not induce much prograde metamorphism in the Quesnel terrane because, it is proposed, these mafic rocks had a significantly higher specific heat capacity (C_p) than the adjacent metasediments by virtue of their higher water content. Table 6-2 shows that the C_p of water is five times greater than that of typical silicate minerals (note that the mean C_p of the Quesnel terrane and Omineca Belt are otherwise comparable). Thus the hydrated secondary mineralogy of the metavolcanics provided an effective heat sink, and there was not enough surplus heat to drive prograde reactions in the mafic rocks. Eventually the highly folded isotherms did flatten out, producing a gentle gradient across the boundary, but that didn't happen until there was a net cooling of the whole 'system', quenching it (see Fig. 6.15, t_7). Presumably this was because the heat supply from below to the westernmost Omineca rocks - especially into 'protrusions' like the Spanish Lake anticline - was not great enough to replenish the heat extracted by the Quesnel terrane, and so smooth out the metamorphic transition at the time when the rocks were still close to their maximum T. Uplift and erosion may have effected another 'front'

TABLE 6-2: SPECIFIC HEAT CAPACITY OF SOME COMMON MINERALS

	mineral	specific heat ($\text{J Kg}^{-1}\text{K}^{-1}$)
"mean continental crust" = Omineca Belt	quartz	745.4
	nepheline	756.8
"mean oceanic crust" = Quesnel terrane	orthopyroxene	792.4
	olivine	847.2
	diopside	721.3
	nepheline	756.8
	water	4188

Notes: This is the only data available for relevant minerals.
Specific heat at constant pressure.

Data from Carmichael (1984).

of heat loss vertically above, promoting isotherm suppression. Another reason why there is a steep metamorphic gradient at the terrane boundary is that if the folds were still forming after the rocks quenched then it is possible that the frozen-in metamorphic zones were attenuated by flattening and shear (especially on the overturned limbs of folds like the Spanish Lake anticline; see Map 4 in the southeast of the thesis area), just as a passive layer would be thinned in progressive deformation.

In contrast, because heat flow into the Seller Creek syncline (and into other thermal synforms along the folded terrane boundary) was dissipated more slowly, there was some thermal recovery and prograde metamorphism in the infolded Quesnel terrane. This is well shown by the garnet and staurolite-kyanite isograds which transect the boundary (see Campbell 1971, Campbell 1978, as well as this thesis), but always in the cores of the synclines, not around the anticlines, lending support to this explanation, and also to the preceding explanation for diachronous isograds. Even in these synclines, however, thermal recovery did not advance very far, partly because of the regional uplift, erosion and cooling, but mainly because of the sustained (lateral) heat loss to the Quesnel terrane.

Isograds are approximately concordant with the terrane contact because isotherms were approximately concordant with it before D₂ (cf. Campbell 1971, p. 183). There is, thus, no paradox in a folded isograd which locally post-dates the fold, when the spatial variations of the relationship between fold formation and isotherm recovery are understood, as has been outlined here.

(26) P-T-t paths

Proceeding from the hypothesis in the preceding sub-section, schematic P-T-t paths are proposed for representative parts of the thesis area which graphically portray the change in P-T through time (Fig. 6.15) with respect to the main tectonic events. Fig. 6.16 shows, in particular, the apparent divergence in P-T-t between a rock in the Seller Creek syncline (SCS) which is presently at the erosion surface, and a rock in the core of the Spanish Lake anticline (SLA) which is presently at the surface. The form of the divergence of the paths is meant to reflect the metamorphic textural differences between these two areas which result from the different timing relationships between D_2 strain and M_2 peak metamorphic mineral growth (Fig. 3.12). This was explained in the preceding sub-section.

The paths are described with reference to time frames which subdivide the deformation sequence; tentative dating of these time frames is given (based on sub-sections 9 and 19). For simplicity, it has been assumed that no differential burial or differential uplift (rate) affected the thesis area after t_2 (this is probably not strictly valid but it would have no significant effect on the P-T-t paths); similarly, it was argued previously that erosion rate did not vary throughout the thesis area, over the time interval of interest. Thus, rocks in SLA and SCS maintain approximately the same difference in depth or pressure (ca. 1.6 km) at a given time from t_2 onwards. However, they pass through a different temperature at the same time, and the same temperature at different times, until that point on the cooling path when they rejoin, and rise to the surface on a steady-state geotherm.

Notes. The basic form of the P-T-t path for SCS in Fig. 6.16

Fig. 6.16

Differentiated pressure-temperature-time (P-T-t) paths. Upper diagram shows path for a rock in the Seller Creek syncline (SCS = solid line) and for a rock in the Spanish Lake anticline (SLA = dashed line). For reference: general facies boundaries and ad-ky-si fields from Fig. 4.3, and generalized almandine garnet 'reaction' in upper greenschist facies.

Lower diagram is enlargement showing key P-T-t points (SCS = solid line, SLA = double line). Note the figure-of-eight form of SLA, resulting from cooling while deformation and burial was still in progress. See text for explanation.

emulates the kind of loop proposed by England & Richardson (1977) and Thompson & England (1984) for the situation in which burial or crustal thickening by folding and (or) thrusting (usually relatively rapid) is accompanied or followed by the recovery of isotherms which have been tectonically disturbed. Recovery is expected to continue for a time after the beginning of erosion which induces a simultaneous decompression. Metamorphic 'quenching' at $T_{\max}$ occurs when isotherms begin to relax or are suppressed (migrate up the local temperature gradient) which typically occurs at less than $P_{\max}$. The path for SLA is different: its unusual figure-of-eight form results from cooling while deformation and burial were still in progress.

t_1 to t_2 (190-185 Ma to ca. 180 Ma). D_1 obduction begins at t_1 and ends at t_2 . There is a pressure increase in the lower plate (Snowshoe Group) due to its being overthrust by the Quesnel terrane. The lower plate remains 'flat' for a while after t_1 before it acquires some (south-westward) dip, at which point the paths bifurcate because some vertical relief now separates SLA (deeper) and SCS (shallower). Specifically, at t_2 SLA is 1.6 km deeper than SCS and they are 10 km apart horizontally; the shear zone dips ca. 9° . Thus, $10 \text{ km} (\tan 9^\circ) = 1.6 \text{ km}$.

Temperature increases in the Quesnel Lake Shear Zone between t_1 and t_2 up to the middle or upper greenschist facies (see sub-section 14 and Chapter 3.2). This is M_1 . Garnet may start to grow now, along with a coarse quartz-mica fabric, in rocks that become the Spanish Lake anticline in D_2 . Thermal recovery continues somewhat after t_2 , partially annealing S_1 mylonitized fabrics. t_3 marks the onset of D_2 , and follows t_2 probably within 5 Ma.

t_3 (180-175 Ma). Major F_2 strain and folding begins at t_3 , causing more crustal thickening. An increase in load pressure due to burial results from the formation of the Mt. Borland anticlinorium (Fig. 6.9) structurally above the thesis area. The folded isotherms begin to recover during the folding, prograding the rocks. This is M_2 .

t_4 (175 to 170 Ma). Erosion starts sometime in this range. It does not necessarily start at P_{max} (the crests of the paths): P_{max} is decided by the competition between erosion decompressing the rocks, and continuing D_2 crustal shortening, thickening and burying the rocks. Rate of erosion must begin to exceed burial rate at P_{max} .

[Erosion suppresses isotherms in high crustal levels (not preserved), quenching metamorphic rocks there.]

t_4 to t_5 (175-170 Ma). D_2 strain continues. Erosion continues to induce a vertical heat flow, but heat is primarily being lost laterally to the Quesnel terrane that has been 'rammed' against and infolded with the hotter Omineca Belt. As a result, thermal recovery first stops at t_5 in the Spanish Lake anticline (see sub-section 25). These rocks quench and begin to cool, while strain continues. It is probably this continuing strain that catalyses retrograde reactions in this area: garnet and staurolite and biotite are partially converted to chlorite and muscovite, especially in L_2 pull-aparts and pressure shadows.

Thermal recovery continues in the Seller Creek syncline because of the retarded heat loss there.

It was postulated in sub-section 24 that because F_3 structures have a similar geometry to F_2 structures, they formed under the same kinematic regime; their lack of recrystallization fabric was attributed

to declining temperature (Elsby 1985 suggested the same thing). If this is correct, it implies that F_3 structures, on all scales, started to form in the Spanish Lake anticline and its vicinity at roughly t_5 , presumably before they started to form in the Seller Creek syncline.

Viewed in this way, the change between D_2 and D_3 is diachronous - actually a function of diachronous cooling - and it migrated through the thesis area from southwest to northeast just like the quenching of the garnet isograd did. It was not clear-cut in space or time. A popular conception of 'polyphase deformation' is that the phases are discretely episodic, but the implication here is that this is not necessarily so: it depends on how the phases are defined.

t_6 to t_7 (ca. 170 Ma). A marked waning in the rate of D_{2-3} southwesterly-verging deformation is inferred at this time, based on the post-kinematic, sub-idioblastic properties of M_2 porphyroblasts in the Seller Creek syncline.

t_7 (170-165 Ma). Thermal recovery stops in SCS. Peak mineral assemblages are quenched and the garnet isograd freezes in. Cooling starts. Rock ductility and annealing recrystallization abate. Continued strain under M_3 conditions takes the form of F_3 structures here, as has been the case in SLA since t_5 .

Porphyroblasts are partially retrograded in the Seller Creek syncline (as well as the Spanish Lake anticline). Garnet and biotite seem to be particularly susceptible, the rare chloritoid much less so.

This account of P-T-t provisionally predicts that rocks in the Spanish Lake anticline quenched at an earlier time and hence at a slightly greater depth than those in the Seller Creek syncline. That

is, that there is a pressure gradient, increasing from north to south. This agrees with Getsinger's (1985) opinion that there is a north to south increase in P in the adjacent Three Ladies - Mt. Stevenson area.

Post-t₇ (post- 165-160 Ma). Sometime on these cooling paths, ductile deformation effectively stops, probably in the Late Jurassic. SLA and SCS eventually coalesce as they rise to the surface on a steady-state geotherm.

(27) Correlation of structures and deformation history between other workers and this thesis.

It is not easy to make unequivocal correlations of deformation phases because the distinguishing criteria of other workers and certain details of their observations are not completely familiar to the author. Also, the major (D_2) structures are somewhat conical, limiting the distance up and down plunge in which they can be extrapolated confidently. However, the structural correlations described below (see Table 6-3) work reasonably well.

Getsinger (1985) described Middle Jurassic, tight to isoclinal syn-metamorphic F_2 folds which the author correlates with his F_2 . However, Getsinger identified these folds as northeasterly-verging. The author tentatively infers that many of Getsinger's F_2 minor folds are on the overturned limb of the major regional F_2 southwesterly-verging fold, the Mt. Borland anticlinorium, such that they would be northeasterly-verging; this would be expected if the author's major downward-facing F_2 structure is extrapolated southeastwards across Grain Creek into Getsinger's Ph.D area. This is a plausible explanation because of the proximity of the studies, but making interpretations farther afield is more speculative.

TABLE 6-3: CORRELATION OF DEFORMATION SEQUENCES DOCUMENTED BY OTHER WORKERS IN THE QUESNEL LAKE REGION WITH RESPECT TO THE DEFORMATION SEQUENCE IN THIS THESIS

AUTHORITY	D1	D2	D3	POST-D3
THIS THESIS	NE-verging strain, formation of foliation, mylonitic at QLSZ. Pre-met. peak.	Tight, recumbent SW-verging folds. NW-trending & plunging. Dominant structures. Syn- or pre-met. peak. Major foliation. Makes map pattern.	Moderate to open, upright to overturned, SW- (some NE-) verging folds, crenulations. Post-met. No significant foliation.	N or NE trending faults, fractures, minor warps.
GETSINGER 1985	Possibly some F1A isoclinal, intrafolial folds	F1A, F1B tight to isoclinal, NE-verging, syn-metamorphic folds, S0-l. L1 quartz-rodding. (Author: downward-facing?) Possibly some F2 SW-verging folds. Syn-met peak. L2 ca-// L1. NW plunge.	F2: SW-verging, tight to normal folds. F3: crenulations & kink folds. Ca. coaxial with F1, F2. SW & NE-verging. NW-plunging. Late metamorphic. Steep.	F4: Upright, open, NE-trending broad buckle folds & warps. Minor crenulations.
PIGAGE 1979		D1: Isoclinal, west-verging, N-NW plunging. Synmetamorphic. Met. peak outlasted D1.	D2: Large upright folds, SW-verging, coaxial with F1. Synmetamorphic but lower than D1. Retrogression.	NE to N-trending folds, faults, fractures.
FLETCHER 1972		D1: SW-verging, syn-met. May include D2, Q1 & D2 not well distinguished.	Possibly D2. Open folds.	D3: Kinks, open folds, retrogression.
MURPHY 1985	D1: Tight, isoclinal E-verging folds.	D2: SW-verging folds. Dominant structures. Syn-met. Peak slightly post-D2.	D3: Upright folds, crenulation regional warps. NE-verging, NW-trending.	F4: Upright crenulations, regional warps. NE-trending.
STRUJK 1980	Pundata thrust, easterly-directed	D2: Open to isoclinal, W-verging (some E-verging). NE-dipping cleavage. W-dipping thrusts.	Some D2 structures. ?Some D3 structures. ?D4 structures.	?D4 structures
ENGI 1984	D1: Rare isoclinal, NW-trending & plunging. S0-l foliation.	D2: SW-verging recumbent folds, NW-trending & plunging. L2 // F2 axes. Major structures. Met. peak syn- to post-kinematic.	Poss. some D2 folds. D3: Upright, broad to open, N-trending folds. Post met. peak.	Poss. some D3 folds. D4: NE-trending kink folds & crenulations.

TABLE 6-3 (concluded):

AUTHORITY	D1	D2	D3	POST-D3
CAMPBELL 1971	-	"Early-stage" Isoclinal, minor folds. Synmetamorphic. "Late-stage" minor folds, regional folds. Overturned to SW. Coaxial. Synmetamorphic.	Late to post-metamorphic minor folds. NE-dipping axial planes.	Minor cross-folds.
ROSS et al. 1985	D1: Rootless, mesoscopic isoclinal folds. Axial planar foliation. S1 in Quesnel Lake Gneiss.	D2: NE-verging. Axial planar, synmetamorphic foliation. Peaks slightly post-D2. (Author: downward-facing?)	D3: W-verging. E-dipping foliation. Lower met. grade than D2. D4: NE-verging, non-penetrative cren. cleavage	
MONTGOMERY 1985	Possibly some F1 rootless isoclinal folds.	Possibly some F1 rootless isoclinal folds. F2: tight to isoclinal folds. L2 // F2. Syn-metamorphic.	F3: SW-verging upright folds. Ca. coaxial with F2. NW-plunging. Syn- to post-metamorphic.	F4-F6: Broad brittle warps & fractures. Minor crenulations.
FILLIPONE 1985	Possibly some F1 isoclinal folds & assoc. mylonitic S1, L1.	Possibly some F1. F2: E-verging, tight to isoclinal. L2 // F2. Syn-met. Dominant foliation. (Author: downward-facing?)	F3: SW-verging, upright or NE-dipping. Crenulations & crenulation cleavage.	F4, F5: crenulations, folds, faults, shear zones, open warps. NE- & NW-trending.
ELSBY 1985	Possibly some E-verging rootless isoclinal folds.	Possibly some E-verging isoclinal folds. F2: E-verging recumbent folds, S2, L2. NW-trending, syn-metamorphic. (Author: downward-facing?)	F3: SW-verging, upright or NE-dipping folds & cren. cleavage. NW-trending. Post-met. peak.	E-verging buckle folds, crenulations, open to tight. NW plunge. Fractures.
CARYE 1986	Possibly some D1 isoclinal, intrafolial rootless folds. S0-1.	Possibly some D1 isoclinal folds. S0-1. D2: open to tight, upright to recumbent, syn-metamorphic folds. S2 usually // S0-1.	Probably some D2 folds. D3: open to moderate, upright folds, cren. cleavage. Late metamorphic.	Gentle to open buckle folds, faults & fractures. NE-trending.
MONTGOMERY 1978	F1: E-verging rootless isoclinal folds	F2: W-verging folds	Some F2 folds	F3: NE-trending folds, crenulations.

Note: Deformation phases under Fillipone (1985) and Elsbey (1985) refer to their "basement" deformation phases.

For example, Ross et al.'s (1985) northeasterly-verging "D₂" structures in the Crooked Lake area are interpreted by the author to be downward-facing folds compatible with the author's D₂. This is the same argument as applied to Getsinger's structure, above, but in this case the northeasterly-verging folds are presumed to be on the downward-facing limb of the continuation of the Spanish Lake anticline. There is a complication, however, in that Ross et al. (1985) relate their D₂ northeasterly-verging folds to the emplacement of the Quesnel terrane. The crux of this dilemma is whether or not their northeasterly-verging "D₂" folds occur in upward-facing structures (Quesnel terrane structurally above Omineca Belt) as well as downward-facing structures, but unfortunately they don't differentiate their structural domains in a way that demonstrates this.

Similarly, Struik's (1982a,b) cross-sections from northwest of the thesis area show northeasterly-verging recumbent folds in the west and southwesterly-verging recumbent folds in the east. He simply joined them in the middle (a rather unlikely structure), but it is suggested instead that they are separated by the axial surface of the northwestern continuation of the Mt. Borland anticlinorium. The main effect of this impertinent interpretation is to invert the stratigraphy in the west of Struik's cross-section - something he would probably object to.

Southeast of the north arm of Quesnel Lake, the cross-section of Fletcher (1972) is dominated by southwesterly-verging structures (Fig. 6.17). These are Fletcher's D₁, possibly including his D₂, and the author correlates them with the D₂ folds of this thesis. The overturned and right-way-up limbs of the continuation of the Spanish Lake anticline show up well. The eastward continuation of the Eastern Quesnel Lake

Fig. 6.17

Cross-section from Fletcher (1972). Inset shows location map and line of section. Note southwesterly-verging folds, and overturned Antler Formation and isograds.

Gneiss is shown isoclinally-folded (probably Fletcher's D_1) in the core of this fold; this is an interpretation that could not be corroborated by the author in the thesis area. The remaining, northeastward part of Fletcher's section presumably passes through the continuations of the Seller Creek syncline - possibly the synformal structure at Roaring River in Fig. 6.17 - and the Mt. Borland anticlinorium, although overturned structures attributable to that structure are not readily apparent in Fletcher's section.

Two important differences between this thesis and the structural interpretations of Montgomery, Elsby and Fillipone are that these workers recognized a deformation (foliation) in the Omineca Belt which is absent in the Quesnel terrane, and that F_3 structures are responsible for the map pattern (although Fillipone, 1985 p. 36, noted that F_3 folds coincide with F_2 folds in the Boss Mountain area).

POST-MIDDLE JURASSIC

(28) Brittle faults

Brittle faulting post-dated all ductile deformation, so it is inferred to be post-Middle Jurassic; more precise timing of the faulting is difficult. The faults in the thesis area, such as the small graben structure west of Mt. Brew, are typically rather small, steep and oriented north to northeast. The faulting implies west-east to northwest-southeast crustal flexure or extension. Faults of this type are common in the superstructure of the Cariboo Mountains, which outcrops north of the Quesnel Lake area (see Campbell et al. 1973, Struik 1982b).

EARLY CRETACEOUS

(29) Plutonism

Large intrusions of granite, granodiorite and quartz monzonite were emplaced in the region in the Early to mid-Cretaceous, although their ages are not tightly constrained (none were mapped in the thesis area). The Naver pluton (105 Ma, K-Ar on biotite; Tipper et al. 1979, Struik 1985a), the Raft batholith (96 ± 5 Ma and 80 ± 6 Ma, K-Ar on biotite; see Campbell & Tipper 1971), and the Baldy batholith (105 ± 9 Ma and 140 ± 9 Ma, K-Ar on biotite; see Campbell & Tipper 1971) are shown in Fig. 2.3; the Bootjack stock nepheline syenite in the Morehead Lake area (111 ± 4 Ma, Bailey 1978, Fig. 5.2) is another example. Apart from emplacement deformation at their contacts, these plutons are completely post-kinematic; they cut the contacts between the Antler and Fennell Formations and the Omineca Belt, and presumably truncate the regional fold structures. Thus, they provide a minimum age for the main deformations: Early Cretaceous. Furthermore, andalusite (in Black Phyllite) and sillimanite (in Snowshoe Group equivalents) in contact metamorphic aureoles around the Raft batholith (Campbell & Tipper 1971), which was probably at a comparable crustal level to the thesis area, indicate maximum pressures to have been between 375 and 400 MPa (= ca. 15 km depth) in the Early to mid-Cretaceous, which shows there had been significant erosion since D_2 and D_3 .

(30) Was there a Cretaceous 'thermal event', or had the region remained hot since the Middle Jurassic?

The spate of plutonism in the late Early to mid-Cretaceous probably means that this was a discrete event, during which the geothermal gradient was raised and isotopic systems were reset, such as in biotite

in the Quesnel Lake Gneiss (see Chapter 2.2). Support for this conclusion is provided by paleomagnetism. The secondary (Y) magnetization identified in Takla Group rocks (see Chapter 5) was interpreted to be a viscous partial thermoremanent magnetization, meaning that it was acquired gradually ('viscous') over a sustained period of time under 'partial unblocking temperatures' of 100-300°C (E. Irving, pers. comm.). The QL paleopole was interpreted to be mid-Cretaceous because of its coincidence with other paleopoles more confidently dated as that age, and because of the absence of reverse magnetizations. If these elevated temperatures had persisted since the regional metamorphism in the Middle Jurassic, the magnetic overprinting would probably have been more complete and certainly more complex to resolve because of the wide change in the geomagnetic field over that interval of time (about 50 Ma) due to plate motions. The fact that the QL magnetization is reasonably well defined suggests that although it was acquired relatively slowly, it was in response to a discrete, relatively short-lived Early to mid-Cretaceous 'geothermal event'.

POST-MID-CRETACEOUS

(31) Role of crustal extension in the Quesnel Lake region

In southern British Columbia and northern Washington there was important large-scale normal-sense faulting in the Eocene. The faults generally dip outwards from the axial culmination of the region. The faulting affects areas which were compressionaly thickened in the Late Cretaceous to early Tertiary, such as the Monashee complex basement duplex (Brown et al. 1986), and also affects areas which were thermally

and magmatically active at that time, such as the Okanagan and Valhalla gneiss complexes. As a result, it is uncertain how much these faults owe their formation to the arching, uplift and gravitational instability of the compressionally thickened and thermally weakened culminations like the Monashee complex (Brown & Journeay 1987) versus bulk crustal extension and ductile necking (Price et al. 1981) aided by hot viscous flow and magmatism (Parrish et al. 1985). The latter is the popular interpretation of normal faults flanking "metamorphic core complexes" in the United States. The processes are not necessarily exclusive, and Coney & Harms (1984) have tried to reconcile them.

In evaluating the contributions of these processes, much discussion focusses on whether the thermally and magmatically active areas had a closely-preceding history of significant crustal shortening, and conversely, on whether the exhumed basement duplex of the Monashee complex was thermally active in the Eocene.

The issue is raised here because of the question of extension in the Quesnel Lake region. The author sees little evidence for it. Apart from comparatively small and laterally spaced, northerly to northeasterly-trending faults, there are no known normal-sense low-angle shear zones in the region (except the Little River Fault: see below). There is no particular reason to infer them either because known shear zones and metamorphic gradients are explicable in terms of Jurassic compressional deformation.

Mention should be made of the Little River Fault (Klepacki 1981) which trends approximately east-west, northwest of the North Arm of Quesnel Lake and has been studied by Getsinger (1985). The fault appears to be a gently northeast-dipping, brittle normal fault with an

easterly or southeasterly displacement of its hanging wall. Klepecki (1981) estimated a throw on the fault of 10 km, whereas Getsinger (1985) gave a maximum throw of 6 km. Although Getsinger (1985) documented a significant peak metamorphic discordance between footwall and hanging-wall (200°C and 200 MPa), she also noted similar late kinematic retrograde minerals in both plates; she concluded that displacement probably began during F_3 folding (presumably Jurassic). However, she also invoked (p. 85) some Tertiary displacement because the fault post-dates the cooling of a 86 Ma pegmatite.

[6] 3. REGIONAL TECTONIC SYNTHESIS

In this final section, the author discusses the regional-scale implications of the results and interpretations arising out of this thesis.

OTHER TECTONIC MODELS OF THE INTERMONTANE-OMINECA BOUNDARY

How does the tectonic history presented in this thesis compare with that proposed for other parts of the Intermontane-Omineca boundary in the Canadian Cordillera, such as in the Yukon, northern British Columbia, and the Kootenay Arc? Is there an analogous tectonostratigraphic record in the conterminous United States?

(32) Yukon

The tectonic model of Tempelman-Kluit (1979) for the Yukon is quite similar to the version given by the author in this thesis, in terms of the tectonic elements involved as well as their structural evolution.

Tempelman-Kluit (1979) postulated the rifting of a continental-crust based block, 'Stikinia', from the North American continental margin in the Early Mississippian, with the production of a very large (1100-5500 km) ocean basin, the 'Anvil ocean'. The Stikinia block could be analogous to the presumed basement of the Quesnel terrane; the Anvil ocean could be analogous to the Antler-Fennell-Kaslo back-arc basin, although it is alleged to be much wider. Between the Upper Triassic and the Middle Jurassic the Anvil ocean was subducted to the west, generating the Lewes River volcanic arc assemblage in Stikinia (probably analogous to the Quesnel Belt; but see below). An accretionary prism comprising three distinct allochthons (mylonitized siliciclastic metasediments, sheared ophiolitic rocks, and sheared plutonic rocks) was formed, represented by the Teslin suture. The allochthons were finally obducted onto the continental margin in the Late Jurassic to Early Cretaceous. Siliceous mylonitized rocks from the allochthons appear to be identical to those in the Quesnel Lake Shear Zone (for example, compare Tempelman-Kluit, 1979, Figure 4B with Plate 3B in this thesis).

To a first approximation, the emplacement of the Stikinia arc assemblage on the allegedly easterly-verging Teslin suture is analogous to the emplacement of the Quesnel terrane onto the Omineca Belt along the Quesnel Lake Shear Zone. The stack of sheared siliciclastic rocks and 'ophiolitic' klippen in and northeast of the Teslin suture would correspond to the Snowshoe Group, and the Antler Formation and Black Riders complex, respectively (R.B. Campbell, pers. comm. 1980), although no significant imbrication of these units has been recognized by the author. One major difference is in the mode of formation of the Lewes River arc: Tempelman-Kluit attributed it to the same (westward)

subduction that eventually emplaced the arc itself, while the author (following Monger 1984) attributes the Quesnel Belt arc to a different and slightly earlier subduction zone lying to the west.

However, Mortensen & Jilson (1985) and Mortensen (pers. comm. 1986) have pointed out certain over-simplifications in the Tempelman-Kluit (1979) model such as, first, that the major deformation and metamorphism in Stikinia and in the allochthonous rocks occurred between the Early Permian and the Late Triassic (by which time the metamorphic rocks had been uplifted and eroded), prior to collision with the continental margin. Thus, the mylonites are probably upper Paleozoic, not Middle Jurassic. Second, kinematic evidence for the vergence of the later, brittle thrusting that carried the allochthons is not documented (i.e. the vergence is an interpretation based on circumstantial evidence). Third, the allochthonous rocks are more imbricated than the straightforward tripartite division mapped by Tempelman-Kluit, while the internal stratigraphy of each thrust sheet is more coherent than he implied. Fourth, there is no good evidence for the polarity of the Lewes River arc.

Although more rigorous documentation of the tectonic evolution of the Yukon is needed, essentially the same tectonic elements of the Intermontane-Omineca boundary are present there as in the southern Canadian Cordillera, namely a basin of upper Paleozoic oceanic crust separating a continental margin from a Triassic-Jurassic island arc. The basin was subducted, probably westwards beneath the arc (although this has yet to be demonstrated) in the ?Middle Jurassic, leading to the obduction of the arc over the continental margin.

■ This summary neglects the role of major dextral-strike-slip faulting that affected the Yukon in the Late Cretaceous to early Tertiary.

However, the tectonic assemblages involved do occur on one side (northeast) of the Tintina Trench strike-slip fault. Certainly, large-scale reconstructions are complicated by this faulting, but to a first approximation the model is probably a workable one.

(33) Kootenay Arc

The Kootenay Arc in southeastern British Columbia is essentially a west-facing 'monocline' consisting of the deformed and metamorphosed boundary zone between upper Paleozoic and Mesozoic Intermontane terranes (Quesnel and Slide Mountain terranes) rooting to the west, and upper Proterozoic and Paleozoic rocks to the east representing the North American continental margin sedimentary prism (Archibald et al. 1983). Thus, the basic situation is the same as that in the Quesnel Lake area farther north. Archibald et al. (1983) refer to the Quesnel and Slide Mountain terranes (under different names) in the Kootenay Arc as "allochthonous". Earlier, Okulitch & Peatfield (1977) suggested that the "eugeoclinal" rocks were obducted by eastward or westward underthrusting of the continental margin.

What are the structural relationships at this boundary, and how do they compare with the thesis area? Detailed work by Klepacki (1983) and Klepacki & Wheeler (1985) has shown that the Kaslo Group is duplicated by a low-angle fault, the Whitewater Fault: the upper plate is floored by serpentinite, overlain by pillowed volcanics, and these are overlain unconformably by rocks of the Upper Triassic Slocan Group (correlated with the Black Phyllite). This package is obviously similar to that in the upper plate of the Quesnel Lake Shear Zone, but the Whitewater fault is Early Permian (it actually pre-dates the Slocan Group). A more

likely analogue of the Quesnel Lake Shear Zone (D. Klepacki, pers. comm. 1984) is the ?Middle Jurassic ?easterly-directed Stubbs Fault which places McHardy Assemblage, Kaslo Group and Slocan Group on top of the remainder of the Milford Group and underlying Lardeau Group. The Stubbs Fault is approximately synchronous with major, Middle Jurassic upright folding (Dryden anticline: Klepacki 1983, Klepacki & Wheeler 1985), which is post-dated by two phases of minor, southwesterly-verging folding. From these observations, it appears that in the Kootenay Arc the fold and fault kinematics and timing of the structural history is very similar to that in the thesis area.

Archibald et al. (1983, Fig. 12) made a schematic, cross-sectional compilation of folds and faults in the Kootenay Arc (documented from numerous workers from 1951 to 1980); both easterly and westerly directed structures occur, and they may be superimposed. Presumably, the variations between their northern and southern sections are a function of depth of erosion and the dominance of early versus later structures. They devised a tectonic model which incorporated their own detailed metamorphic (barometric) and geochronological information as well as structural data, proposing that in the Middle Jurassic the North American margin was driven westwards beneath "Quesnellia" (= Quesnel terrane). The "miogeocline" was buried rapidly to depths of 20 km, deformed (presumably by easterly-verging structures - this is implied in their sketch diagram but not specified), metamorphosed, and intruded by syn- to late-kinematic granitic plutons. This thickening and heating of the miogeocline "restored its latent buoyancy", and it was uplifted rapidly in the Late Jurassic to Early Cretaceous while Quesnellia was still converging with North America. As a result of this uplift

component focussed in the east, the miogeocline was thrust westwards over Quesnellia. This reversal in vergence is represented in the superposition of structures.

This aspect of their model is basically the same as that proposed in this thesis for the structural vergence in the Quesnel Lake region (Section 2). That is, first the Omineca Belt was subducted beneath the northeasterly-obducting Quesnel terrane (D_1), and afterwards its displacement vector relative to the Quesnel terrane was deflected upwards to the west due to rebound, with the production of major southwesterly-verging folds (D_2 and D_3). The strength of the author's modelling is that the superposition of structures and their vergence is quite clear in the Quesnel Lake region; it complements the metamorphic and geochronologic arguments of Archibald et al. (1983) well. One difference, however, is that Archibald et al. (1983) date the D_1 - D_2 vergence reversal as Late Jurassic, about 15 Ma later than the 180-175 Ma age postulated by the author.

Ellis & Watkinson (1984) suggested a very similar evolutionary scenario for the Kootenay Arc, in which the North American continent was subducted beneath a "convex-east over-riding plate", followed by its

"re-emergence in the form of an asymmetrical crustal pop-up, and the successive inboard failure of the margin leading to the formation of the Purcell anticlinorium and the Canadian Rockies".

Price (1986) used these interpretations of the structural history in the Kootenay Arc to develop his ideas on tectonic wedging (see Section 2). He proposed that "Quesnellia" was first obducted, but continued easterly-verging shearing on its basal fault zone caused the lower part of Quesnellia to indent itself into the Omineca Belt to the east, while

the upper part was folded (westerly) back on itself to balance the basal thrusting. As explained in sub-section 22, this mechanism may have produced the D₂ folds in the Quesnel Lake region also, but it is not the author's favoured explanation.

(34) Northern British Columbia

The Sylvester allochthon in northern British Columbia consists of oceanic rocks including basalt, cherts and serpentinite, and is separated by a fault from miogeoclinal rocks of the Omineca Belt. The Sylvester allochthon has been grouped with the Eastern Assemblage (Monger 1977), the Eastern terrane (Monger et al. 1982), and presently with the Slide Mountain terrane (Monger 1984). Harms (1986) disagreed with the consensus view (see Section 1) that this terrane represents marginal or back-arc basin oceanic crust: she found highly variable rock types which represent pre-obduction telescoping of a very large, major ocean basin.

The deformation history of the tectonic boundary also differs markedly from the thesis area and the Kootenay Arc. Harms (1985b) mapped a southwesterly-verging duplex in the autochthonous or para-autochthonous North American rocks immediately beneath the Sylvester allochthon. The folding in the duplex does not involve the allochthon, requiring that they be detached. The roof thrust of the duplex is also the basal thrust that simultaneously emplaced the allochthon from the west. Harms (1985b) inferred from this that the duplex must have wedged itself between the allochthon and the autochthon by delamination; the event is bracketed between the end of the Triassic and the mid-Cretaceous (Harms 1985a).

Another important difference is that following the emplacement of the allochthon and the formation of the duplex, there was no significant

folding of the contact, unlike in the thesis area and the Kootenay Arc because of continued convergence there. Perhaps the Sylvester allochthon was too thin to induce much of a dynamic response in the lower plate, or perhaps convergence and shortening ceased entirely at this structural level.

(35) Tectonostratigraphic assemblages in the conterminous United States

Cordilleran evolution in the conterminous United States is broadly similar to that in Canada in that it involved the eastward emplacement of peri-cratonic, arc, and oceanic assemblages onto the paleo-continental margin, followed by the major development of an Andean-type magmatic margin. However, the extrapolation and correlation of terranes and tectonic events between Canada and the U.S. south of about latitude 48°N is made difficult by two things. First, there is a comparative paucity of exposure of pre-Tertiary rocks in southern Washington, Oregon and Idaho due to the cover of Cenozoic volcanic and plutonic rocks. Second, the timing of orogenic events in the U.S. seems to be different from similar events in Canada, while conversely, approximately coeval events in the two regions may be dissimilar in character; this suggests diachroneity as well as diversity of tectonic processes along the length of the evolving continental margin.

The major features of tectonic events in the U.S. are summarized below in chronological order, based on Davis et al. (1978), Coney et al. (1980) and Suppe (1985), and their possible relationships with those in Canada are briefly discussed.

Antler orogeny. The earliest recognizable crustal shortening event in the Cordillera is the emplacement of the Roberts Mountain allochthon.

of west-central Nevada in the Late Devonian to Mississippian, in which distal continental margin or marginal oceanic rocks were thrust eastward over the miogeocline. The allochthon is covered by the "Antler overlap sequence" consisting of sandstone, limestone and chert-pebble conglomerate, which sounds similar to the coeval Guyet Formation and Milford Group in British Columbia. However, the latter do not overlap any known analogue of the Roberts Mountain allochthon. Indeed, the Antler orogeny was coeval with back-arc extension in Canada and the formation of the oceanic crust of the Antler and Pennell Formations and Kaslo Group (of the Slide Mountain terrane; Section 1); they would appear to be unrelated tectonic events. Furthermore, no Devonian-Mississippian structures have been found in the Cariboo Mountains, leading Struik (1981b) to reject the suggestion of previous workers (e.g. Monger et al. 1972, Campbell et al. 1973) that the Antler orogeny is represented there.

Sonoma orogeny. The next major event in the U.S. was the collision and emplacement of an extensive upper Paleozoic terrane or microplate consisting partly of island arc volcanics and sediments; it is known as Sonomia. Sonomia accreted against the continental margin, and an intervening accretionary prism called the Golconda allochthon was thrust eastwards onto the coeval, and possibly laterally equivalent Antler overlap sequence (cf. the Kaslo-Milford primary relationship postulated by Klepacki 1983) in western Nevada, in the Late Permian to Early Triassic (Speed 1979). The Golconda allochthon (or terrane) may be correlative with the Harper Ranch Group of the Quesnel terrane. The thrusting is approximately contemporaneous with the Early Permian thrusting within the Kaslo Group (Klepacki & Wheeler 1985), but that thrusting was apparently intra-oceanic. Permo-Triassic deformation in

southern British Columbia is also indicated by important angular unconformities documented by Read & Okulitch (1977), who correlated the implied tectonism with the Sonoma orogeny. However, as the significance of these stratigraphic hiatus apparently diminishes to the north in the Quesnel terrane (Read & Okulitch 1977), it is inferred that the Sonoma orogeny is not recorded in the Quesnel Lake region. Indeed, the more or less constant thickness of the Antler Formation between the Raft batholith and Prince George implies the absence of an angular unconformity (versus a more probable disconformity) between the Permian and the overlying Late Triassic Black Phyllite.

Furthermore, if the correlations and extrapolations with the Sonoma orogeny in the northern U.S. are warranted, they suggest that the Harper Ranch arc, Kaslo back-arc basin, and Milford continental margin were partially telescoped in the Permo-Triassic, prior to the formation of the overlying Quesnel Belt. The reconstruction of the Quesnel Belt offered in this thesis (Sections 1, 2) requires that the telescoping cannot have been very large, since the reconstruction calls for the existence of a reasonably broad oceanic basin between the Quesnel Belt arc and the continental margin.

Because of the Cenozoic cover in Washington and Oregon, it is not known if there was an arc outboard of the new continental margin in the U.S. in the Late Triassic-Early Jurassic analogous to the Quesnel Belt farther north, although there are Upper Triassic to Middle Jurassic volcanic and sedimentary rocks and blueschists in east-central Oregon and Idaho that might be correlative, and which might indicate an arc or inter-arc setting there (Davis et al. 1978). (This is also the approximate location of the Triassic "Seven Devils arc", but this is

another terrane which was not emplaced until the mid- to Late Cretaceous: Lund et al. 1985.)

Triassic-Jurassic rocks very similar to all the main components of the Intermontane superterrane form the pre-Nevadan (see below) basement to the eastern Klamath and northern Sierra terranes in northern California (Saleeby 1983). He suggested that this composite terrane was accreted in the western U.S. in the Middle to Late Jurassic, analogous to (and perhaps connected with) the "Stikine terrane" (= Intermontane terrane) in Canada (although the latter was probably emplaced somewhat earlier). The complex distribution of basement terranes and overlap arcs in California suggested to Saleeby (1983): "rifting and interarc-basin-forming events, interspersed with amalgamation and accretion events" and transform faulting.

In the Late Jurassic (Nevadan orogeny) to Late Cretaceous (Sevier orogeny), the continental margin gradually evolved from formerly a longitudinally diverse, marginal oceanic, arc-trench type of setting (cf. formation, amalgamation and accretion of the Intermontane superterrane in Canada) to a mainly subaerial, Andean-type magmatic mountain belt, beneath an easterly-dipping (oblique?) subduction zone. The batholithic belt of the Sierra Nevada truncates the aforementioned Paleozoic and Mesozoic orogenic trends, signifying a major plate readjustment in the late Mesozoic. Oceanward of the Sierra Nevada is the Great Valley, consisting of a forearc sequence possibly built on Jurassic oceanic crust (Saleeby 1983), and the Franciscan (subduction-) complex of the Coast Ranges.

By this approximately mid- to Late Cretaceous time, all the

accretion of major allochthonous terranes and superterranes that presently comprise the collage of the Cordillera had been completed (including the Insular superterrane in Canada and the northeastern U.S., and the Seven Devils arc terrane in Oregon). Crustal shortening continued in the form of intraplate, usually easterly-verging thrusting, such as in the Sevier and Laramide overthrust belts in the Late Cretaceous and early Tertiary, respectively. These thrusting events are contemporaneous with that in the Foreland belt in Canada.

POST-ACCRETION FAULTING AND RELATIVE LATITUDINAL MOVEMENTS:
THE OUTSTANDING QUESTION IN CORDILLERAN TECTONICS

(36) Review of terrane and superterrane accretion

Part of the conclusion of this thesis and of previous studies is that in the late Early to early Middle Jurassic the western margin of continental North America was subducted approximately westwards beneath the eastern margin of an already amalgamated, composite 'oceanic' superterrane, the Intermontane superterrane, resulting in its obduction onto the Omineca Belt.

A second composite superterrane, the Insular superterrane, was attached to the Intermontane superterrane in the Early to mid-Cretaceous, probably by the eastward subduction of oceanic crust; the 'suture' between them is marked approximately by the batholithic Coast Plutonic Belt.

Thus, the geological evidence as compiled by Monger & Price (1979), Monger et al. (1982) and Monger (1984) is that the main exotic elements of the Canadian Cordillera were amalgamated and attached to the Omineca Belt by the mid-Cretaceous.

(37) Complications and the debate they produce

The picture is made more complicated by two things. One is post-accretion dextral strike-slip faulting in the northern Canadian Cordillera. Approximately 900 km of dextral strike-slip displacement has been proposed for the Tintina Trench - Northern Rocky Mountain Trench (TT-NRMT) fault system (Fig. 1.1), while cumulative displacements on these and other faults in the north-central Canadian Cordillera probably amount to more than 1000 km (Gabrielse 1985). Most of the movement on the fault system occurred in the Late Cretaceous to early Tertiary (Gabrielse 1985), so it post-dates the amalgamation of all major terranes and their accretion to the continental margin. Strike-slip offset apparently diminishes along the fault system southeastwards: no significant offset has been proven south of latitude 56°N where faults become mainly shallow-dipping thrusts, constituting the southern Fofeland Belt. Geological relationships from east to west across the Southern Rocky Mountain Trench (SRMT, the physiographic extension of the NRMT south of 54°N ; Figs. 1.2, 2.3) have been interpreted to preclude significant dextral offset there (Price & Carmichael 1986, Murphy 1984), although other workers have questioned some of the criteria (Chamberlain & Lambert 1985), or have suggested that strike slip has been masked by younger faulting (Gabrielse 1985). It is still an open question as to what happens to TT-NRMT strike slip in the south.

The other 'complication' is the paleomagnetic evidence (see Chapter 5, especially Section 7): a group of aberrant paleopoles from a major part of the Canadian Cordillera indicate that in the mid-Cretaceous it was on a different plate from the North American plate, and situated about 2000 to 2500 km farther south and obliquely oriented by about 60-

70° anticlockwise. This aberrant super-duper-terrane comprises the Insular and Intermontane superterrane and at least part of the Omineca Belt to which they were firmly attached. It has been called "Cordilleria" by Chamberlain & Lambert (1985) and "Baja British Columbia" by Irving (1985 and Irving & Wynne, in prep., although this term was introduced by D.B. Stone in 1977); the term Baja B.C. will be used hereafter. The inferred (relict) plate boundary between Baja B.C. and the North American plate must lie somewhere between the westernmost Omineca Belt (the most easterly locality with a discordant paleopole: QL, this thesis) and the eastern limit of Cordilleran deformation. It should, of course, be definable on stratigraphic and structural bases as well as by paleomagnetism.

At first sight, it might appear that these two complications can be reconciled in that the dextral strike-slip faulting represents an intra-continental transform fault system along which Baja B.C. has moved relative to North America. At present, however, most workers see a major problem in this hypothesis, namely that the TT-NRMT fault system records only ca. 50% at most of the displacement of Baja B.C. implied by paleomagnetism, and anyway, the continuation of its strike slip south of 56° (where the discordant paleopoles come from) is far from proven - some would contend it is disproven because there is no tectonic discontinuity in the appropriate place that signifies a displacement in the order of 2000 km. It is even more difficult to find evidence for transcurrent movement or a plate boundary in the northern United States.

As a result, the issue comes down to a debate which may be reviewed by reference to two recently published tectonic models. Although neither of these models necessarily epitomizes one side of the debate,

they serve to bring out the salient points of view.

One strategy is to play down the paleomagnetism, given the many assumptions involved in its processing and interpretation, and emphasize instead the apparent lateral geological continuity in the southern Canadian Cordillera. With this in mind, Price (1984) and Price & Carmichael (1986) have proposed a model for the accommodation of TT-NRMT strike slip by transpressive thrust faulting in the south, although they considered a strike slip of only 450 km - the minimum estimate available (see Gabrielse 1985). Noting that the TT-NRMT fault trajectory forms a small circle until latitude 56° where it diverts into a great circle trajectory, they proposed that strike slip in the north was transformed into a form of transpressional 'subduction' ("A-subduction zone": Bally 1981, Price et al. 1985) in the south in which the North American cratonic basement was 'rotationally underthrust' up to 200 km southwestwards beneath its shelf and miogeoclinal cover, forming the folds and thrusts of the southern Foreland Belt. They further invoked a simultaneous distributed shear and related crustal extension southwest of the SRMT, which transferred dextral strike slip into the en echelon Fraser River - Straight Creek fault zone in southwestern British Columbia along which there is an additional ca. 100 km of movement (Monger 1985). The pole of relative rotation (Euler pole) describing the strike slip is 158.7°W , 35.4°N , which is 1600 km north of Hawaii. The timing matches: like the TT-NRMT faulting, the thrusting in the Foreland Belt is approximately Late Cretaceous through early Tertiary.

The timing may be approximately consistent, but the magnitudes of strike slip versus thrust displacement in the north and south, respectively, may not balance, especially if Gabrielse's (1985) estimate

of strike-slip (900 to 1000+ km) is the more accurate one. Nevertheless, the Price & Carmichael (1986) model is clever and appealing; most notably it is a neat, self-contained model that acts as a foil to more radical tectonic modelling which is premised on paleomagnetism.

Thus, the other side of the debate emphasizes the paleomagnetic discordance of Baja B.C., and maintains that despite appearances strike-slip persists undiminished south of 56° where it is probably distributed over a broad array of faults which may have superimposed dip-slip displacements, or faults which are buried beneath younger thrust sheets, such that there are no prominent geological breaks on the surface.

The tectonic model of Chamberlain & Lambert (1985) advocates large displacements for Baja B.C., although it proposes that the eastern boundary of Baja B.C. was at least partly a major subduction zone, not simply a continental transform fault. In this somewhat radical model, Baja B.C., or "Cordilleria" in their terms, was a microcontinent on the Kula plate (see later) which was already freely moving northwards relative to North America in the mid-Cretaceous from a position corresponding to California, when the aberrant magnetizations were acquired. This motion was effected by the Kula plate being underthrust southwestwards by a Late Jurassic back-arc basin on the western edge of the North American plate. They imply that it was by this time (mid-Cretaceous) that the bulk of the paleolatitude anomaly and rotation had been accommodated. Collision with North America began at about 95 Ma, in the northern Cordillera, forming the Mackenzie Mountain thrust belt. Thereafter, Cordilleria was constrained to slide northwestwards against North America along the incipient TT-NRMT (transform) faults, while the southern part of Cordilleria also rotated anticlockwise, closing the

remaining "wedge-shaped basinal area" between the southern Omineca Belt and the craton, "pushing a miogeoclinal wedge of thrust sheets in front of it" (Foreland Belt) by southwestward oblique subduction.

The later phase of this model that relates TT-NRMT strike-slip faulting to contemporaneous oblique subduction farther south is basically the same as that of Price & Carmichael (1986). The difference between the models is that prior to these Late Cretaceous-early Tertiary tectonics, the alleged Cordillera was quite separate from North America and had travelled northwards in the order of 1000 km via subduction as a quasi-independent microcontinent, thus accounting for its paleomagnetic discordance.

A full critique of this model is not really warranted here, but a few important aspects need to be highlighted:

- * There should be the hallmarks of the alleged suture zone, perhaps a Cretaceous subduction complex, between Cordillera and the collapsed North American marginal basin; Chamberlain & Lambert ascribed the absence of appropriate features (apart from granites) to their being overridden during collision or being covered by later thrust sheets.

- * There is no adequate explanation offered for why Omineca Belt stratigraphy displays far more similarities to Foreland Belt stratigraphy than differences, considering that they are supposed to have quite separate origins. The evidence they offer to undermine the significance of the Malton gneiss complex, which straddles the SRMT and has been used by some workers as an argument against strike slip, is not particularly convincing, namely age, compositional, and isotopic differences between the dismembered gneiss bodies.

- * Chamberlain & Lambert's timing may be wrong or need revision, in

that they contend that most of the latitudinal movement (ca. 1000 km) began at about 120 Ma and was completed before about 100 Ma. In a discussion of their model, Irving (1985) objected that major movement could not even have begun until about 85 Ma because that is when the Kula plate formed (Woods & Davies 1982; see later), and also because the aberrant magnetizations are about 100 Ma old or slightly younger.

These criticisms of the Chamberlain & Lambert (1985) model would probably be used by opponents of large displacements to dispute that Baja B.C. is far-travelled. Nevertheless, the model is audacious and useful.

Summary. The debate boils down to the feasibility of tectonic displacements for Baja B.C. in the order of 500 to 1000 km, which are reasonably acceptable geologically and mechanically, versus those of 2000 to 2500 km which presently are not, but which explain its paleopoles. The enigma is confounding. A lot hinges on the authenticity of the paleomagnetism, but although its interpretation may be vulnerable to criticism, paleomagnetism has potentially a powerful ally in Mesozoic and Cenozoic plate reconstructions. This will now be reviewed.

(38) Plate geometries and reconstructions

As Riddiough (1982) emphasized, any discussion of tectonic models and their feasibility must be compatible with regional plate interactions, as far as they are understood for the last 100 Ma.

A ridge-ridge-ridge triple junction is inferred to have been the focus of the Kula plate (KP), the Farallon plate (FP) and the Pacific plate (PP) since the Late Cretaceous (Fig. 6.18; see Atwater 1970, Riddiough 1982, Engebretson et al. 1985). Reconstructions are

Fig. 6.18

Reconstructions of relative plate motions in the eastern Pacific basin, mid-Cretaceous (110 Ma) to mid-Tertiary (37 Ma), based on Engebretson et al. (1985; see their Figures 3 and 8). Fixed hotspot reference frame. Double heavy lines = oceanic ridges (dashed where inferred), single heavy lines = transforms, arrows = plate motions (arrow length = 10 Ma of motion). Large open arrow in each time window = approximate convergence vector of oceanic plate (labelled K or F) relative to fixed NAP.

NAP = North American plate, KP = Kula plate, FP = Farallon plate, PP = Pacific plate, I = Izanagi plate, H = Hawaiian hotspot, Y = Yellowstone hotspot.

Note the uncertainty of KP-FP-NAP triple junction at 80 and 65 Ma, and the major dextral relative motion of KP-NAP.

difficult before then, during the Cretaceous Normal Superchron (roughly 120 to 85 Ma) because of the dearth of ocean floor magnetic anomalies, but the picture becomes clearer afterwards. Most models envisage the westwardly or southwestwardly-moving North American plate (NAP) overriding KP and FP (e.g. Riddihough 1982) for most of the last 85 Ma, so their relative interaction has generally been convergent and rapid, especially between FP and NAP (about 12 to 17 cm a⁻¹); the KP-NAP interaction was much more oblique (dextral). This geometry requires that the KP-FP-NAP triple junction migrated, perhaps erratically, along the NAP margin, with KP usually moving dextrally at a low angle of convergence with respect to NAP, and FP also usually moving somewhat dextrally, but at a high angle of convergence (Fig. 6.18). Any part of NAP which might have become welded to KP, especially, would be translated northwards with it by oblique subduction and dextral transform faulting. [The west-east FP-NAP trajectory earlier in the Mesozoic was probably the mechanism by which Tethyan faunas in the Intermontane superterrane were transported thousands of kilometres across the proto-Pacific ocean: Cox & Engebretson 1984.] This process has been advocated by several workers, based on paleomagnetism not only in British Columbia, but as far south as California and Mexico, and as far north as Alaska. Beck (1986) put it succinctly: FP-NAP interaction could be held responsible for the accretion of allochthonous terranes by subduction, and for the creation of batholithic complexes in the Jurassic and Cretaceous, and KP-NAP interaction could be held responsible for their post-accretion displacements and re-arrangements.

[It should be emphasized that the actual plate interactions were probably far more complex than this simple scenario, so tectonic

activity may have varied in space and time between high angle convergence and crustal shortening, and low angle convergence and transform faulting, on both KP-NAP and FP-NAP fronts.]

The plate that Baja B.C. travelled on (regardless of its latitude of origin) is generally assumed to have been the Kula plate, because geometrically its interaction with NAP has been dominantly dextrally transformal. If this is correct, then the paleogeographic source of Baja B.C. is constrained by the location of the KP-FP-NAP triple junction at about 85 Ma (the time of the initiation of the Kula plate: Woods & Davies 1982). Atwater (1970) tentatively put the KP-FP-NAP triple junction in northeastern Mexico at 80 Ma, allowing for very large (2000+ km) displacement of Baja B.C., consistent with the paleomagnetism. More recent reconstructions (Riddihough 1982, Engebretson et al. 1985), however, indicate the possibility that in the Late Cretaceous and early Tertiary (ca. 85-50 Ma), the KP-FP ridge intersected NAP between about the Canada-U.S. border and northern California-Oregon, allowing for only up to about 1000 km of displacement, compatible with the larger estimate of TT-NRMT offset but not with the paleomagnetism. Fig. 6.18 illustrates the range of uncertainty. Note, however, that although this 'northern option' for the KP-FP ridge argues against a far-south origin for Baja B.C., it doesn't help Price & Carmichael (1986) either!

Whatever the position of the KP-FP-NAP triple junction, there is some consensus on the part of plate reconstructions that KP-NAP interaction was predominantly transformal in the interval ca. 85-50 Ma, and was situated such that it could drive several hundred kilometres of dextral displacement for which there is geological evidence. Plate

velocities (1 to 10 cm a⁻¹; Beck 1986) were sufficiently high to do the job in a few tens of millions of years or less. Perhaps, as a function of the shape of the Late Crétaceous continental margin and the regional resolved stress field, KP-NAP interaction was also responsible for the shortening in the Foreland Belt (Laramide orogeny) and the Eocene crustal extension documented in southern British Columbia, which could then be viewed as side effects of a complex plate boundary system. In a general way, this suggestion is implicit in the models of Price & Carmichael (1986) and Chamberlain & Lambert (1985). The question still comes down to the validity of the paleomagnetism versus the validity of southern Omineca Belt-Foreland Belt primary lateral continuity.

Poles of relative plate motion. Yet another twist to the problem is that the Euler pole of relative plate motion between Baja B.C. and NAP, and the Hawaiian Euler pole based on TT-NRMT spherical geometry (Price & Carmichael 1986) are radically different.

Irving & Wynne (publication in prep. 1986) have calculated that the Euler pole of Baja B.C.-NAP relative motion is situated near Chicago (EPA, Fig. 6.19). Thus, a rotation of about 60° about this axis accomplishes the translation and rotation of Baja B.C. on a small circle northwards along the margin of NAP at the same time as rotating the mean paleopole on a larger small circle from its expected or original position north of the Bering Strait to its observed position in the northeast Atlantic (Fig. 6.20).

This is not particularly edifying unless this Euler-pole can be shown to be also representative of KP-NAP relative motion. This has been investigated by Engebretson (1982) who determined a KP-NAP Euler pole in the region of Central America (EPB, Fig. 6.20). What this

Fig. 6.19

Possible coastwise migration route of Baja B.C. assuming an original position in northern Mexico (from Irving & Wynne, manuscript in preparation 1986). Corresponds to trajectory A in Fig. 6.20. 85 Ma to 50 Ma is the most likely interval of displacement, which amounts to 2000 to 2500 km.

Fig. 6.20

Map showing possible paths of translation and rotation of Baja B.C., from Irving & Wynne (manuscript in preparation 1986). A 'coastwise' trajectory for Baja B.C. originating at A requires an Euler pole of rotation with respect to North America situated near Chicago (EPA). An offshore origin for Baja B.C. at B yields an Euler pole in Central America (EPB; from Engebretson 1982). Because paleolongitude cannot be determined paleomagnetically, either EPA or EPB is possible, or any intervening Euler pole and trajectory; resolution of these options may depend on the geological evidence for (EPB) or against (EPA) accreted oceanic crust between Baja B.C. and North America. Presently, a coastwise trajectory would be more acceptable on this basis.

Both EPA and EPB rotations accomplish the rotation of the expected (cratonic) paleopole from the Bering Strait to the observed northeast Atlantic position.

implies is that Baja B.C. could indeed have moved northwards on the Kula plate, but at least part of its journey was 'offshore' of NAP, i.e. that relative movement was not merely an intra-continental transform fault but involved some subduction as well (similar to path B in Fig. 6.20).

The Hawaiian Euler pole (Fig. 6.22) is obviously completely different from both the Chicago and Central America Euler poles, and actually indicates an anticlockwise rotation of Baja B.C. because the TT-NRMT is concave westwards. It cannot account for the northeast Atlantic paleopoles, and appears to add to the anomaly of the observed paleopoles instead of partially offsetting it! This is another quandary that tends to polarize the debate. However, it may be that because the Hawaiian Euler pole determined by Price & Carmichael (1986) pertains to only a small component (450 km) of strike slip, on only one fault (TT-NRMT), it should not be regarded as diagnostic of relative motion on what might have been a much larger and much more complex system of faults marking the plate boundary (E. Irving, pers. comm. 1986).

(39) Possible corrections of paleomagnetic results

At this point, then, it is pertinent (and might provide some relief) to mention the ways in which the magnitude of the paleomagnetic anomalies might be reduced or even removed.

Age of magnetizations. In the mid-Cretaceous, the North American plate was at its most northerly position in its post-Jurassic history of motion. Thus, any paleopole anomaly and implicit relative movement based on magnetizations of that age will be at a maximum. Revised, younger or older ages for the magnetizations would automatically reduce the anomaly (though not enough to cancel it). The ages of overprints

such as the γ magnetization (QL, see Chapter 5) are difficult to date precisely and could be younger or older than mid-Cretaceous (and hence probably less discordant), but some primary magnetizations from mid-Cretaceous intrusions (e.g. SP and CS, see Chapter 5 and Irving et al. 1985) are probably as well-dated as magnetizations can be. Overall, the interpreted mid-Cretaceous ages for the anomalous paleopoles in Fig. 5.15 are certainly the most reasonable at present. In any case, no matter how substantial any age revision might be, it would not be enough to eliminate the need for some latitudinal movement.

Reference field. Every paleomagnetic interpretation depends on a comparison with an accurate reference palepole. Some parts of the North American APW path (e.g. the Triassic-Jurassic boundary) are uncertain and are undergoing review (May & Butler 1986), but the mid-Cretaceous results are quite consistent (see Irving & Irving 1982). This is probably not a source of significant error.

The tilt problem. An alternative to the interpretation of latitudinal movement and rotation involves the role of post-magnetization tilt. The paleofield of North America, or the expected mid-Cretaceous magnetization, is steeply inclined downward to the northwest, versus observed magnetizations from Baja B.C. which are northeast and downward. If there was an in situ tilt of the mid-Cretaceous paleohorizontal of about 30° to the southwest it would rotate the expected paleofield direction into the observed direction (Fig. 6.21), and thus cancel the need for relative plate movements.

So what is the control on the all-important mid-Cretaceous paleohorizontal? In the Quesnel Lake locality and in the other studies it has generally been assumed to be the same as present-day horizontal,

Fig. 6.21

Equal-angle stereogram of mid-Cretaceous paleofield directions of North America (ENA, 74, 322), the hypothetical original position of Baja B.C. (REP, 69, 330.5), and the mean observed direction from Baja B.C. (OP, 57, 027). Tilt required to reconcile OP with ENA = 31° to the southwest (horizontal axis of rotation = 146°); tilt required to reconcile OP with REP = 28° to the west-southwest (horizontal axis = 157°).

that is that there has been no significant tilting since the mid-Cretaceous; thus no tilt corrections have been made. This may seem like a bold assumption: not only can it rarely be supported by presently horizontal Cretaceous strata, but in fact in some parts of British Columbia Upper Cretaceous and Eocene strata are quite highly deformed, and local angular unconformities do indeed indicate a southerly tilt (see Irving et al. 1985, p. 592). More results are needed from localities where the paleohorizontal can be constrained but it seems inevitable that some tilting has occurred, probably largely due to small-scale faulting and (or) differential uplift.

But this is not the point: if these widely distributed localities had undergone tilts of random directions and magnitudes (the most likely effect of local deformations), this would merely have scattered paleopoles around the expected North American mean. The observed grouping could only arise if the tilts among the localities were approximately uniform and unidirectional. Opponents of tilt as the source of the paleomagnetic anomaly say that such consistent tilting over such a large area of British Columbia and northern Washington is very unlikely; more likely, they say, the anomalous paleopoles are primarily due to plate motions, and some random, generally gentle differential tilting between localities has produced a second order dispersion of the paleopoles in the northeast Atlantic region. So here lies the justification for making no tilt corrections: random tilts, within a given paleomagnetic locality and Baja B.C. as a whole, would be averaged out.

However, the author believes that a systematic southwestward tilting of the Cordilleran crust should not be dismissed too readily: it is

possible and must be investigated. What is the minimum tilting that would be required? An interesting question is: suppose there was a geologically accountable, ca. 1000 km of transform faulting on the TT-NRMT system, combined with mechanically related, transpressional strain farther south - how many degrees of uniform tilting would then be required to move this 'revised expected paleofield' for Baja B.C. to the observed direction? The author made a simple calculation. An original paleolatitude for Baja B.C. was taken at the Oregon-California border (Fig. 6.22), compatible with the northern option for the KP-FP-NAP triple junction (Fig. 6.18). It was imagined to move directly north along the continental margin: this trajectory is a compromise between the anticlockwise rotation and translation required by the Hawaiian Euler pole, and the clockwise rotation and translation of the Chicago Euler pole. This moves the paleopole into the East Siberian sea, (roughly 162°E , 73°N). The resulting paleofield on the stereogram (Fig. 6.21) would be approximately 69° , 330.5° .

The question is how much tilting would move this direction into the observed direction? It turns out to be almost the same as before: 28° ; there is only a slight difference in the direction of tilting - it would be more westward than southwestward (Fig. 6.21). While this solution is valid only for the proposed paleolatitude and trajectory, it serves to suggest that only a substantial tilt (ca. 30°) or a large clockwise rotation will explain the observed paleopoles for Baja B.C., even after 1000 km of relative latitudinal movement is accepted.

Is thirty degrees of uniform westward or southwestward tilting a feasible process, and is there any geological support for it? This is a subject for future research and it is only possible to speculate at

Fig. 6.22

Equal-area polar projection showing present-day position of Baja B.C. (dashed) and a hypothetical mid-Cretaceous original position (dotted) at about 1000 km farther south along the continental margin. ENA = expected North American paleopole, REP = revised expected paleopole for Baja B.C., i.e. paleopole expected after hypothetical movement from ENA, OP = mean observed paleopole for Baja B.C., CEP = Chicago Euler pole, HEP = Hawaiian Euler pole. Baja B.C. is moved 1000 km northwards (arrow) - a compromise path between the anticlockwise rotation required by HEP and the clockwise rotation required by CEP. ENA moves to REP. Diagram shows how even this 1000 km of northward movement of Baja B.C. does not get its paleopole much nearer OP than is ENA (cf. Fig. 6.21).

present. Flexure of the lithosphere is thought to occur in a number of situations, such as at a plate boundary under compression, or when a plate overrides a thermal anomaly in the mantle. A regional flexure in the lower crust would induce faulting in the brittle upper crust which would tend to have regionally consistent geometry and kinematics, and hence a regionally consistent tilt of fault blocks. The western NAP boundary was under compression when the Insular superterrane was accreted, completing the assembly of Baja B.C., followed by the formation of the Coast Plutonic Belt magmatic arc. Subsequently, two things could have caused the western edge of the North American plate to flex: first, a major dynamic readjustment from compressional to transformal conditions at about 85 Ma when the Kula plate formed, and second, a decay of the thermal anomaly that accompanied arc magmatism. Another kind of thermal anomaly was the KP-FP ridge: if Baja B.C. overlay the ridge when the magnetizations formed, its passage away from the ridge might have caused it to subside (in a southwestward direction?), analogous to the subsidence of a passive continental margin during sea-floor spreading. [Incidentally, the massive uplift of high pressure metamorphic rocks documented in the Coast Plutonic Belt is probably not relevant to regional tilt because the greatest uplift occurred before the Late Cretaceous, at the latest (Crawford & Hollister 1982).]

These are mainly hypothetical effects, of course, and as well as demonstrating them on independent grounds, it would be necessary to show how they could generate consistent fault kinematics, and fault block rotation of sufficient magnitude. This would be difficult, but nevertheless the quest for geological evidence for tilt as an alternative to the plate tectonic interpretation is surely as reasonable and worthwhile as the quest for a plate boundary that although is easy

to infer, is presently very hard to see.

In fact, any documentable mechanism (age revision, reference pole adjustment, tilt correction) for lowering the daunting paleomagnetic anomalies for Baja B.C. would be welcome! Alas, none of these corrections is particularly warranted at the moment.

(40) Summary and conclusion

Proponents of a northern Mexico source for Baja B.C. should endeavour to show that the KP-FP-NAP triple junction was that far south in the mid-Cretaceous. They must also explain how the continental transform fault system or plate boundary that was generated in transporting Baja B.C. northwards is so well disguised, particularly south of latitude 56°N.

Conversely, opponents of this scenario should explain how TT-NRMT strike slip is accommodated farther south, and also exactly why the paleomagnetic data is misleading.

The solution to the problem is not yet available. More work is needed on displacement vectors and kinematics on faults, measurements of fault offsets, plate reconstructions and Euler poles, primary magnetizations in Cretaceous bedded rocks, and particularly on critical evidence of terrane continuity - not just east of Baja B.C. but all across the Cordillera.

In conclusion, the author sees two main alternatives, (a) and (b).

(a) In the mid-Cretaceous, Baja B.C. was situated 500 to 1000 km farther south relative to the North American craton. In the Late Cretaceous to early Tertiary it became attached to the Kula plate and moved northwards to its present relative position, as the KP-FP-NAP

triple junction migrated along the margin of NAP from a position roughly corresponding to Oregon. The Quesnel terrane, probably an offshore island arc and never far from the continental margin, contains Triassic-Jurassic faunas equivalent to those found in the miogeocline at the latitude of Nevada, suggesting an upper limit of Baja B.C. discordance of about 1000 km, regardless of where the other components of Baja B.C. originated. As a corollary, suspect terranes in the Omineca Belt are merely distal parts of the North American miogeocline, and are not exotic.

The Baja B.C.-North American plate boundary was a complex transform fault system, marked today partly by dextral offset on the TT-NRMT and probably by related transpressional (and transtensional?) deformation farther south.

More radical, Triassic-Jurassic paleomagnetic and paleontologic discordances between components of Baja B.C. and North America were partly accommodated by relative motions before the final amalgamation of Baja B.C. in the mid-Cretaceous. The difference in magnitude between the mid-Cretaceous paleomagnetic anomalies and the geologically substantiated displacements could be explained by undetected westward or southwestward tilting of Baja B.C., or corrections to the age of magnetizations or to the reference paleopole, or all of these. Presently undetected, post-100-80 Ma systematic faulting in the western Cordilleran lithosphere is a feasible source of error in the prevailing interpretation of the paleopoles that the present horizontal is statistically identical to the mid-Cretaceous paleohorizontal.

(b) In the mid-Cretaceous Baja B.C. was situated 2000 to 2500 km farther south relative to the North American craton, approximately in the position of northern Mexico. This was also the location of the KP-

FP-NAP triple junction at about 85 Ma. Baja B.C. became attached to the Kula plate and slid northwards along the NAP boundary on some kind of transform fault system or via highly oblique transpression, possibly also involving subduction of intervening oceanic crust. Only a part of this fault system or suture is preserved, in the form of the TT-NRMT and SRMT, and only part of the displacement is discernible from geological offsets (only on the TT-NRMT). The structural signature and mechanisms of the remaining tectonic displacements (1000 to 1500 km) have still to be defined. They may have been obscured by continental arc magmatism and transient compressional and extensional tectonic regimes, effects of constantly changing KP-NAP and FP-NAP interactions over the last 100-80 Ma.

(41) Epilogue

The author tentatively favours alternative (a) because it is more compatible with stratigraphic, paleontologic and structural evidence. It is also, in principle, explicable in terms of a plausible tectonic model along the lines of Price & Carmichael (1986) which was designed with this evidence foremost in mind. Either this physical evidence must be shown to be invalid or misconstrued, or the constraints on the paleomagnetic data base need to be improved, before more radical plate tectonic reconstructions for Baja B.C. can be accepted.

CHAPTER SEVEN

SUMMARY OF INTERPRETATIONS

Few references are cited here, but it is emphasized that this summary includes the interpretations of other workers (as cited elsewhere in the text) in addition to those of the author.

1. TECTONIC ASSEMBLAGES

INTERMONTANE BELT

This corresponds to the Intermontane superterrane and in the thesis area it consists of an assemblage of oceanic crust unconformably overlain by island arc volcanics and sedimentary rocks.

Oceanic crust

The oceanic crust at the base of the assemblage is the Antler Formation, and was formed in the Mississippian or Pennsylvanian to Early Permian in a back-arc basin. The basin formed by the rifting and spreading away to the west of the edge of the (paleo-)continental margin of North America, which included an overlying contemporaneous (Upper

Devonian to Permian) arc. The subduction zone dipped eastwards, towards North America.

Coeval rocks of Antler-affinity occur farther southeast in British Columbia, suggesting that the back-arc basin was at least 500 km long; in the Kootenay Arc there is actual evidence of the basin-continent interface. The width of the basin is not known but was at least 150 km.

Subduction ceased before the Triassic, based on the apparent termination of arc volcanism. There may have been plate re-adjustments then, related to the Sonoma orogeny recorded in the United States.

Island arc

Subduction resumed in the Late Triassic, again dipping towards the continent. The trench was situated west of the old upper Paleozoic arc; a new arc was built approximately on the old one. This Late Triassic to Early Jurassic island arc, represented regionally by the Nicola Group and Takla Group and collectively called the Quesnel Belt, probably began with calc-alkaline volcanism. In the Antler Formation oceanic basin there was no back-arc spreading; it is termed a marginal basin here to emphasize this difference from its upper Paleozoic character. Oceanic circulation in this basin was limited, restricted by the arc to the west. Hemipelagic carbonaceous mud of the Black Phyllite was deposited in the basin disconformably or unconformably on the Antler Formation oceanic crust; most of it was probably derived from the west. The Black Phyllite is here included as part of the Quesnel Belt.

As the volcanoclastic apron of the arc prograded into the basin, the Black Phyllite was overlapped gradationally by a coarsening-upward sequence of tuffaceous argillites, volcanic wackes and debris flows.

Centres of Volcanism also spread eastwards into the basin with time: the youngest Takla Group rocks in the thesis area are augite porphyry basalt flows.

Much of the Quesnel Belt, even the Late Triassic volcanics, is potassium-rich alkalic basalt rather than calc-alkaline basalt and andesite, which is unusual in large amounts for an island arc. Nevertheless, such potassic alkalic volcanism is certainly known from some modern island arcs such as in Fiji and New Ireland; if the basalts are regarded as shoshonites, there are many more analogues. The chemical feature that principally identifies the Quesnel Belt as subduction-generated is the consistently low TiO_2 contents, along with other characteristic trace element values. This conclusion is significant because it refutes previous interpretations that the Quesnel Belt was formed by rifting by virtue of its chemical similarities with intraplate oceanic island basalts.

The reason for the alkalic volcanism is not known but, again based on comparison with modern island arcs, it may be related to block faulting and uplift due to arc stabilization, or to plate re-adjustments. A subduction complex formed in the trench and forearc region to the west of the Quesnel Belt arc; it is represented by the Cache Creek Group of the Cache Creek terrane. Subduction ceased in the Late Triassic or earliest Jurassic with the collision of the Stikine terrane, carried in on the subducting plate. This collision may be implicated in the conversion to alkalic volcanism described above.

The Quesnel terrane, comprising the upper Paleozoic rifted arc and the overlying Quesnel Belt, was about 25 to 30 km thick in the Early Jurassic.

OMINECA BELT

The Omineca Belt in the thesis area is represented by the Snowshoe Group, which is interpreted as 'distal turbidites' formed in the outer part of a submarine fan complex. It was originally about 10 km thick and was deposited on attenuated continental or transitional crust basement between 5 and 15 km thick. The basement was attenuated, and the Snowshoe Group deposited, in response to Late Proterozoic or late Paleozoic rifting, subsidence and passive continental margin sedimentation, or a combination of both phases.

Granite to granodiorite plutons, some porphyritic, were intruded into the Snowshoe Group in the Late Devonian or early Mississippian, now forming the Quesnel Lake Gneiss. They are not alkalic but may be connected with the upper Paleozoic rifting.

2. TECTONIC MODEL

LATE EARLY JURASSIC

The tectonic development of the thesis area began in the late Early Jurassic with the collapse and obduction of the Antler marginal basin onto the Snowshoe Group of the continental margin. This is the first phase of deformation (D_1) and began at about 190 Ma based on the demise of marine sedimentation in the Quesnel terrane in the Toarcian, which is taken to mark uplift due to underplating. This is also approximately the time of the opening of the north-central Atlantic ocean.

The obduction detachment is marked by a mylonitic shear zone at the contact between the Antler Formation and the Snowshoe Group and Western

Quesnel Lake Gneiss, the Quesnel Lake Shear Zone. Stretching lineations in parts of the shear zone, where their orientation was not significantly affected by later deformations, indicate displacement was at a high angle to the present strike of the orogen, namely northeast-southwest. The sense of shear is upper plate to the northeast, determined from systematically rotated megacrysts, asymmetric syn-kinematic C and S type fabrics, and shear band foliation in the Western Quesnel Lake Gneiss. The detachment rooted to the southwest, probably at the brittle-ductile transition in the upper mantle beneath the interface between the Quesnel terrane arc and the Antler marginal basin; this is alleged (theoretically, by others) to be a particularly weak zone and a likely site for obduction-subduction initiation. The detachment propagated northeastwards and upwards and probably 'daylighted' in the basin near its northeastern flank. The Antler Formation in the Black Stuart synclinorium is interpreted to be near the northeastern limit of the obducted allochthon; thus, the Pundata thrust at its base is the continuation of the Quesnel Lake Shear Zone (or Eureka thrust). The thinner, northeastern toe of the allochthon here was thickened and strengthened by layer-parallel imbricate thrusts, indicated by repetition of conodont zones in Antler Formation cherts.

An easterly paleoslope on the upper plate is indicated by recycled chert in the thesis area, transported from the Cache Creek terrane far to the west.

Because the northeastern flank of the Quesnel terrane arc came to rest on the Snowshoe Group, almost the entire width of the intervening marginal basin must have been obducted, estimated to be at least 150 km. Given this value and estimated plate velocities, obduction took 7 to 10 Ma and was completed by ca. 180 Ma.

D₁ deformation produced a foliation (S₁) and lineation (L₁), best developed in the lower plate. The formerly-attenuated basement contracted somewhat by reverse movement on re-activated faults; a shortening and thickening of about 15% has been modelled for D₁ internal strain. In the Quesnel Lake Shear Zone, temperatures reached a maximum of about 400° to 450°C during D₁ (middle to upper greenschist facies), testified by the nature of mylonitic dynamic recrystallization of quartz and feldspar. These temperatures were realized quite quickly due primarily to the thermal blanketing effect of the overthrust Quesnel terrane, but there was not enough heat transfer to produce an inverted thermal aureole in the lower plate. The depth of the shear zone in the thesis area was about 17 km, corresponding to a pressure of ca. 475 MPa.

MIDDLE JURASSIC

After D₁ obduction, the coupled Quesnel terrane (which henceforth includes the Antler Formation) and Omineca Belt were folded by southwesterly-verging folds of the second deformation phase, D₂. D₂ began within about 5 Ma of the end of D₁ obduction, namely at or sometime before 175 Ma; this is based on the dating of regional syn-kinematic intrusions and on isotopic resetting due to D₂ metamorphism.

Two sub-recumbent F₂ folds of the Quesnel Lake Shear Zone dominate the structure of the thesis area: the Spanish Lake anticline (SLA) and the Sellar Creek syncline (SCS) to its north. Northeast of SCS the rocks continue to lie on the overturned limb of a major regional F₂ fold, the Mt. Borland anticlinorium. Northeast of this structure is the Black Stuart synclinorium, in which the continuation of the Quesnel Lake

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Shear Zone - the Pundata thrust - appears again, with the (type-)Antler Formation in its upper plate. F_2 structures plunge to the northwest typically about 10° .

In the thesis area a foliation (S_2) formed parallel to the axial surfaces of F_2 folds; a stretching lineation (L_2) formed syn-kinematically, parallel to F_2 fold axes and to the S_0 - L_2 intersection. The D_2 metamorphic grade was generally middle to upper greenschist facies; garnet forms the only well-defined first appearance isograd. Staurolite is restricted to one locality. The grade drops southwestwards in the Quesnel terrane to the prehnite-pumpellyite facies. The thesis area reached a maximum depth of about 20 km (about 550 MPa) during D_2 due to thickening and burial; these pressures are inferred from metamorphic analyses from other theses in the region.

Peak metamorphic porphyroblasts are syn-kinematic in SLA and are overprinted by syn-kinematic retrogression, well shown by retrograde L_2 pull-aparts and pressure shadows and by unannealed quartz strain. In contrast, peak metamorphic minerals in SCS slightly post-date S_2 although they were also retrograded later. Assuming that deformation proceeded in SLA and SCS synchronously, the textural variation between these areas indicates diachronous quenching, being earlier in SLA than SCS. This is explained by more rapid heat loss to the enveloping Quesnel terrane in SLA during D_2 than in SCS which was enclosed by warmer and more heat retentive rocks. The garnet isograd is thus diachronous; it conforms closely to the folded Quesnel Lake Shear Zone around SLA because it was frozen in subparallel to it at an early stage in its folding, whereas the isograd cuts through the 'synformal' shear zone in SCS because isotherm recovery outlasted the folding there. Consequential P-T-t paths illustrate this with a conventional loop for

SCS and an unusual figure-of-eight form for SLA resulting from cooling while deformation and burial were still in progress.

The L_2 axial extension is tentatively interpreted to be due to a pure shear quasi-constrictional component of the overall D_2 strain regime (constrictional axis horizontal, northwest-southeast); the other component is transverse or oblique (transpressive) simple shear. The simple shear was responsible for the southwesterly overturning of F_2 folds and S_2 foliation but did not contribute much of a southwestward stretch because this was buffered or counteracted by the pure shear component. The only extension in the resultant strain path that remained additive in progressive deformation was that parallel to the fold axes, and so L_2 emerged as the finite maximum. The magnitude of L_2 extension was not measured or calculated.

An oblique shear component is supported by the en echelon pattern of F_2 fold axes at the terrane boundary, but it was not strong enough to be held responsible for L_2 (this opinion contrasts with regional tectonic models of others which do explain the lineation by a strike-slip component).

A corollary of this hypothesis is that the Omineca Belt continental margin was extended longitudinally as well as thickened vertically during D_2 crustal shortening. This may be because it was irregularly shaped and (or) was not confined at its northwest or southeast ends such that convergence with the Intermontane superterrane induced a 'lateral (longitudinal) extrusion', crudely analogous to that known to affect the Himalayas.

Major northwesterly-trending, southwesterly-verging folds are documented by numerous other workers in the Quesnel Lake region and farther east in the Omineca Belt, confirming the regional importance of

this deformation. It is usually recorded as the second preserved phase of deformation, and is always associated with the peak of metamorphism. There is broad but not, however, full agreement between this thesis and other workers on the deformation history of the terrane boundary in the Crooked Lake area.

The cause of the D_2 southwestward vergence, which affected the entire width of the Omineca Belt in the Middle Jurassic, is a key aspect of the tectonic history. It is suggested that when the continental margin was further compressed following the cessation of D_1 obduction, the basement contracted by reverse movement on listric faults that originally had attenuated it during rifting. The sedimentary cover was forced to contract by folding. Although there was no major shear in the basement - as in basement subduction or duplexing - a southwestward rotation (overturning) of the folds was produced by an upward rebound of the Omineca Belt which automatically ensued from the basement shortening and thickening: that is, the Quesnel terrane's displacement vector was deflected downwards from its overthrusting D_1 trajectory to a northeastward 'undertucking' in D_2 . The mode of basement shortening, moreover, uniform along its width rather than being concentrated in one or two major shear zones as has been suggested by Murphy (1985) and Price (1986), induced a component of layer-parallel pure shear shortening in the cover with a vertical flattening plane and a strike-parallel extension (see above).

This model complements the strain model for S_2 and L_2 well. It differs from other models in that it explains D_2 geometry and kinematics without recourse to major basement thrusting. This is desirable because it is hard to justify synthetic basement thrusting (Murphy 1985) in

light of the fact that the Early Jurassic to early Tertiary orogenic history (excluding D_2) is so dominated by westerly-rooting, easterly-verging thrusting and duplexing. Antithetic basement thrusting would be more in character, but requires forming D_2 structures by tectonic wedging (Price 1986); this is plausible, but is not intuitively favoured by the author at present in the form envisioned by Price.

An implication of the preferred model is that the basement was too thin to be duplicated en masse until it re-achieved something like its pre-rifting thickness and strength (cf. Jackson 1980); perhaps it did this in the Late Jurassic, when the Monashee decollement began to form - a development that indeed closely coincides with the demise of D_2 and D_3 southwesterly-verging deformation.

Third phase deformation (D_3) consists of moderate to steeply dipping northeasterly-verging but mainly southwesterly-verging folds and crenulations which have very limited axial surface foliation. They formed under a very similar or identical kinematic regime as D_2 structures with little or no hiatus, but during declining temperatures. Considering the P-T-t interpretation above, the onset of D_3 emulated peak metamorphic quenching in migrating from SLA to SCS, beginning around 170 to 165 Ma; it ended around 160 Ma (Late Jurassic).

Minor northeast-southwest trending brittle faulting affected the area in the Late Jurassic or later.

Post-tectonic Early Cretaceous plutonism in the Quesnel Lake region provides a minimum age for the deformations described above. Elevated temperatures reset isotopic systems at around 100 Ma, and caused a partial viscous thermoremanent secondary magnetization in Takla Group volcanoclastics southwest of Quesnel Lake. The paleopole of this magnetization (QL) is discordant with respect to the expected North

American paleopole.

At present, there is no evidence for significant crustal extension of Tertiary age in the region.

3. REGIONAL TECTONIC SYNTHESIS

OTHER PARTS OF THE INTERMONTANE-OMINECA BELT BOUNDARY IN THE CANADIAN CORDILLERA

The tectonic model in this thesis, namely the obduction of an island arc and marginal basin onto the North American continental margin, appears to be applicable all along the Cordillera, except possibly in northern British Columbia. In this, the model corroborates the early ideas of Monger et al. (1972), Monger (1977) and others. There are, however, disagreements in timing, and in mode of formation and dimensions of the pre-obduction tectonic elements.

Furthermore, in the Kootenay Arc at least, obduction is overprinted by westerly-verging folding correlative with that documented in this thesis, suggesting that the dynamic responses to obduction were also consistent along the length of the Omineca Belt collision zone. That this response involved some upward rebound of the continental margin, reversing the kinematic regime (see Section 2), is corroborated by detailed metamorphic and geochronologic analysis in the Kootenay Arc (Archibald et al. 1983).

No well defined correlations can be made between this thesis and tectonic events in the conterminous United States, although the basic pattern of accretion of arc and marginal basin assemblages there is similar.

POST-ACCRETION FAULTING AND RELATIVE LATITUDINAL MOVEMENTS

Accretion of the Intermontane and Insular superterranes to the North American (paleo-)continental margin was completed by the mid-Cretaceous. Paleopoles of that age from British Columbia and northern Washington from the Wrangell terrane in the extreme west to the Intermontane-Omineca collision zone, including the QL paleopole of this thesis, are well grouped, confirming the internal continuity of this area which is called Baja B.C. However, these paleopoles radically disagree with the reference paleopole for cratonic North America, indicating a relative latitudinal displacement of Baja B.C. of 2000 to 2500 km from the south, including a 60-70° clockwise rotation.

Some plate reconstructions indicate that the Kula plate (KP) had a sufficiently rapid, oblique dextral trajectory with respect to North America (NAP) that it could have transported Baja B.C. 2000+ km from a paleolatitude in northern Mexico to its present position in the Late Cretaceous to early Tertiary (85 to 50 Ma). This is provided that KP originated that far south: other plate reconstructions suggest that it originated no farther south than Oregon, meaning that only about 1000 to 1500 km of the apparent displacement of Baja B.C. is attributable to KP-NAP interaction.

The most serious objection to the paleomagnetic interpretation, however, is that there is no fault or fault system between Baja B.C. and the NAP miogeocline which adequately signifies the nature and magnitude of their discordance. The Tintina Trench - Northern Rocky Mountain Trench (TT-NRMT) strike slip system in the northern Cordillera has the right age, location and sense, but at most only half the magnitude, and

anyway it appears to be transformed into dip-slip thrusting farther south, in the Foreland Belt east of Baja B.C. Even allowing for obscuring the hallmarks of strike slip by later thrusting, crustal extension and magmatism, the absence of a suitable tectonic feature(s) combined with apparent east-west stratigraphic continuity is presently perceived as a compelling argument against accepting the full amount of the paleomagnetic discordance. On the other hand, a relative displacement of only 1000 km, compatible with the estimated TT-NRMT offset and with a more northerly origin of the Kula plate (see above), may possibly be accommodated by oblique convergence (transpression) on known faults between Baja B.C. and the miogeocline without seriously violating stratigraphic data, the remaining paleomagnetic anomaly being explained by error in interpretation.

Sources of error in the paleomagnetic interpretation include the age of magnetizations and the accuracy of the reference paleopole, although both of these have reasonable assignments at present, and the role of tilting which is not well constrained. The anomaly could be reduced, or indeed entirely removed without latitudinal translation, by an undetected regional southwestward tilting of Baja B.C. of about 30° , or by a smaller tilt combined with some mechanism of clockwise rotation. A regional flexure or extension of the western North American lithosphere in the Late Cretaceous to early Tertiary, perhaps due to the major plate re-adjustments at this time, is perhaps one way to generate regionally consistent faulting and tilting of the upper crust.

THE END

PLATE 1

In this and other plates, captions may refer to features which are explained in later parts of text.

A: Typical flaggy interbedded pelites and psammites of the Snowshoe Group. Loc. 641, near Goose Peak.

B: Snowshoe Group, subunit HPSg quartz-granule conglomerate. Loc. 880, looking NW. Pen is plunging down L₂, on S₀-1-2, marked by stretched quartz granules.

C: Interbedded Snowshoe Group quartz psammites and pelites, showing F₂ minor folds. Loc. 499, near Badger Peak, looking NW.

D: Snowshoe Group psammites and pelites, showing F₃ minor folds. Loc. 158, near Mt. Brew. White blebs are dismembered trondhjemite veinlets, flattened in S₂ and stretched out in L₂, visible on surfaces at base of picture. Pencil in lower centre.

E: Snowshoe Group paragneiss from Loc. 153, cut normal to S₀-1-2, and parallel to L₂. Sample not oriented but relative sense of shear discernible from flattening foliation and rotated garnet porphyroblasts.

F: Comparatively mildly deformed Western Quesnel Lake Gneiss at Loc. 16. Note large microcline megacrysts. Pencil in centre.

G: Surface parallel to plane of shear in Western Quesnel Lake Gneiss at Loc. 16, showing L₁ stretching lineation defined by feldspar megacrysts. Hammer on left.

H: Antler Formation greenstone (subunit, PPAb) at Loc. 372. Colour banding and lenticular structure may represent highly sheared and transposed pillows and selvages.

I: Antler Formation greenstone (subunit PPAb) at Loc. 215, south of Mt. Brew. Looking SE, showing SW-verging F₃ folds. Hammer at bottom.

PLATE 1

PLATE 2

- A: Antler Formation gabbro (subunit PPag) at Loc. 204, near Mt. Brew. Several cumulate layers visible, grading from mafic base (actinolite/hornblende) to plagioclase-rich top.
- B: Antler Formation serpentinite (sedimentary) breccia (subunit PPAs) from Loc. 28. Dark serpentinite fragments suspended in pale carbonate matrix.
- C: Typical Black Phyllite pelite (uTrBPP) at Loc. 7. Pale image in centre is glare from graphitic sheen on cleavage surfaces.
- D: Volcanic conglomerate (subunit uTrBPc) within Black Phyllite at Loc. 276. Note zonation of large clast on left, suggesting it is a pyroclastic deposit.
- E: Bedded argillites and siltstones of subunit TrJTa at Loc. 354. Representative of stratigraphic transition between Black Phyllite pelites and Takla Group volcanic wackes.
- F: Bedded, graded volcanoclastic siltstones of subunit TrJTs at Loc. 333. Small dark flecks may be crystal fragments; round features in centre may be silicified mud balls. Pencil on right.
- G: Augite porphyry basalt flow of subunit TrJTb at Loc. 324. Dark phenocrysts are pyroxene; pale minerals are amygdules of prehnite and albite.
- H: Basaltic flow breccia of subunit TrJTb at Loc. 117. Hammerhead at top.
- I: Chert-pebble conglomerate of subunit ImJc at Loc. 77. Dark pebbles are chert; others are mainly volcanic. Penny in upper right for scale.

PLATE 2

PLATE 3

- A: Highly deformed Western Quesnel Lake Gneiss from Quesnel Lake Shear Zone at Loc. 700. Note size reduction of feldspar porphyroclasts compared with their comparatively undeformed state in Plate 1F.
- B: Highly deformed Snowshoe Group psammite from Quesnel Lake Shear Zone at Loc. 52. Parallel to stretching lineation (XZ plane). Note very fine foliation and highly transposed compositional layering. In thin section this rock resembles Plate 4H.
- C: Antler Formation greenstone in Quesnel Lake Shear Zone at Loc. 696. Note sheared compositional layering, parallel to foliation. Pen on right.
- D: Antler Formation amphibolitic greenstone from Loc. 954. Note dissimilar compositions of layers, indicating intense transposition of protoliths.
- E: Typical XZ face of Western Quesnel Lake Gneiss at Loc. 16 (cf. Fig. 3.28) on which kinematic study was made. Note bias of megacryst long axes dipping to west (left). Scale bar is 15 cm. long.
- F: Oriented sample of augen gneiss from Loc. 16, showing XZ face (looking south). Mylonitic foliation approximately parallel to edge of scale card. Note tails of feldspar porphyroclasts attenuating into shear plane, and shear band in lower left.
- G: XZ face at Loc. 16 showing heterogeneous shear strain. Upper half shows horizontal protomylonitic foliation parallel to plane of shear. Lower half is less deformed in which XY foliation is oblique to shear plane, dipping west (to left). Cf. C and S fabrics. Scale bar is 15 cm long.
- H: Oriented XZ section of protomylonite from Loc. 16, showing well-developed shear band foliation, dipping to right (east). Numerous thin grey lenses are completely dynamically recrystallized quartz porphyroclasts. Most of the feldspar has been converted to fine white mica (dark foliation).
- I: Oriented XZ section of Western Quesnel Lake Gneiss from Loc. 715 (overturned limb of Seller Creek syncline). S₁ mylonitic foliation parallel to edge of scale card. Rotated feldspar porphyroclasts and recrystallized tails (cf. Plate 4G), and domains of SW-dipping flattening foliation (also S₁) indicate upper surface to northeast sense of shear.

PLATE 3

PLATE 4

A: Western Quesnel Lake Gneiss in Quesnel Lake Shear Zone at Loc. 914, looking N. Overturned limb of Seller Creek syncline. Three principal planes of strain visible (labelled).

B: Photomicrograph of microcline megacryst in Western Quesnel Lake Gneiss at Loc. 16 showing brittle fractures infilled with polygonized quartz. Note alteration to fine white mica (mu) along upper edge. Scale bar = 1 mm.

C, D and E: Oriented XZ photomicrographs of Western Quesnel Lake Gneiss from Loc. 16; looking N, east to right, subhorizontal mylonitic plane of shear parallel to long edge. Upper surface to the east sense of shear. Scale bars = 1 mm.

C: Microcline porphyroclast with partially recrystallized 'tail' in upper right.

D: Quartz porphyroclast with dynamically recrystallized tails at its lower left and upper right. Centre of crystal shows transition from undulose extinction to subgrains to 'new grains' with high angle boundaries.

E: Mostly dynamically recrystallized quartz porphyroclast, with remnants of deformation bands, dipping west.

F: Western Quesnel Lake Gneiss from Loc. 700 (not oriented). Note more rounded shapes of white feldspar porphyroclasts compared with Loc. 16 (Plate 3F) and their attenuation in the mylonitic foliation; grey, quartz ribbons have aspect ratios of 8:1 or more.

G: Photomicrograph of microcline porphyroclast from Western Quesnel Lake Gneiss at Loc. 715. Oriented XZ section, looking NW. Asymmetric recrystallized tails in upper right and lower left, and distorted twinning indicate upper surface to the northeast sense of shear. Note recrystallized quartz ribbons at top, upper centre, and bottom. Scale bar = 1 mm.

H: Photomicrograph of impure quartz mylonite from Quesnel Lake Shear Zone near Loc. 1042. XZ section, not field-oriented. Note partially recrystallized quartz ribbons. Mylonitic foliation tilted for optimum resolution of ribbon grains. Scale bar = 1 mm.

I: View looking NW from Loc. 992 on ridge north of Mt. Brew, towards Badger Peak. Superimposed lines indicate trace of Quesnel Lake Shear Zone, folded about D₂ folds: Seller Creek syncline in middle distance, Wolverine anticline above it, and Badger Peak syncline. Compare cross-section I-I'. Viewer is standing above hinge zone of Spanish Lake anticline.

PLATE 4

PLATE 5

- A: P2 minor folds in Snowshoe Group grits at Loc. 23, looking SE. S2 is axial planar to folds while S3 is steeper. The two are clearly distinguishable below hand at left.
- B: Quasi-symmetrical (M-shape) P2 minor folds in Snowshoe Group at Loc. 1032 in hinge zone of Spanish Lake anticline. Looking NW.
- C: Photomicrograph of Snowshoe Group pelite at Loc. 695 in N of area near Seller Creek syncline. S2 foliation is horizontal and clearly crenulates an earlier S0-1 fabric; both defined by muscovite and chlorite. Garnet (top) is subidioblastic and overgrows S2 with S1 = Se, and no pressure shadow, indicating late or post-kinematic peak metamorphic recrystallization. Scale bar = 1 mm.
- D: Photomicrograph of Snowshoe Group pelite from Loc. 205 south of Mt. Brew, parallel to L2, looking NE. S0-1-2 horizontal and another syn-kinematic mica fabric dipping left (NW) indicates upper surface to the southeast sense of shear associated with L2. Scale bar = 1 mm.
- E: Photomicrograph of Black Phyllite pelite from Loc. 707 in hinge of Seller Creek syncline. Shows P3 crenulations, with no mineral growth in axial surfaces ('S3'). Note idioblastic garnet with S1 = Se and no pressure shadow (cf. Plate 5C). Garnet is mostly retrograded, except for core. Scale bar = 1 mm.
- F: P2 minor fold in Snowshoe Group pelite from Loc. 153 in Spanish Lake anticline; S2 deflected around garnet porphyroblasts (arrowed) indicating syn-kinematic growth (see Plate 5G).
- G: Photomicrograph of Snowshoe Group paragneiss (shown in Plate 1E) from Loc. 153 in Spanish Lake anticline; section parallel to L2. Note irregular shape of garnet porphyroblasts, S1 not parallel to Se, and highly serrated unannealed quartz grain boundaries, all indicating syn-kinematic peak metamorphism. Scale bar = 1 mm.
- H: Photomicrograph of Snowshoe Group pelite from Loc. 1010 in Spanish Lake anticline, parallel to L2, looking SW. Garnet porphyroblast has asymmetric chlorite pressure shadows dipping right (NW), indicating upper surface to the southeast sense of shear associated with L2. Faint spaced foliations dipping left may be shear bands. Scale bar = 1 mm.
- I: Photomicrograph of Snowshoe Group pelitic paragneiss from Loc. 153 in Spanish Lake anticline, parallel to L2. Black patches in upper centre are dismembered garnet with coarse chlorite growing in pull-aparts. Note unannealed quartz with highly serrated grain boundaries. Indicates post-peak metamorphic retrogression and strain. Scale bar = 1 mm.

PLATE 5

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APPENDIX A

DETAILED DESCRIPTION OF THE SNOWSHOE GROUP

Subunit HPSg. This subunit is characterized by the predominance of competent 'gritty' siliciclastic lithologies over pelites, and includes quartz psammites, (feldspathic-) quartzites and quartzofeldspathic pebbly conglomerates. These rocks are well bedded, with units ranging in thickness up to about 50 cm, although some are over 1 m thick. They are interbedded with subordinate fine grained, finely bedded, well foliated grey phyllitic pelites and thin psammites, which are locally calcareous. Such micaceous beds and partings render the exposures flaggy.

The rocks in this subunit are various shades of pale to dark grey, pale green, pink or buff, and off-white. 'Gritty' psammites are fine to medium grained, with a bimodal grain size consisting of a variable proportion of 1-2 mm well rounded detrital granules of blue or vitreous dark quartz and subordinate sericitized feldspar, in a fine or medium grained quartzofeldspathic matrix. Clinozoisite is present in slightly calcareous psammites. The grits and psammites are usually well foliated, defined by muscovite and chlorite (locally biotite) in disseminated flakes amounting to 10 or 15%, or as thicker partings between layers of psammite. Quartz granules are strained and many have been completely polygonized. Mica foliae are deflected around the granules producing a subtle augen structure. At least some white mica in the matrix was derived from the alteration of feldspar. Other psammites lacking granules are equigranular and apparently well sorted. Limonitic spots and general weathering stain are widespread.

Quartzofeldspathic conglomerates, pebbly siliceous grits, and quartzites contain detrital granules and pebbles up to about 10 mm across (usually about 5 mm) of blue or clear quartz and variable amounts of white, sericitized feldspar. No rock fragments were found. The quartz is subrounded; the feldspar is subrounded to subangular. Primary grain size, where it can be recognized, appears to be generally well sorted, and graded bedding was found in a few localities, providing the local younging direction. Primary textures are everywhere obscured by recrystallization. Clasts in the conglomerates are commonly but not always flattened in the foliation, some with an augen shape, especially where deformation is comparatively strong. In some localities the pebbles and granules are completely polygonized by deformation and are drawn out into 'rods' in the order of 1-2 cm long, forming a prominent stretching lineation (see Chapter 3.3). Roughly 10% muscovite and chlorite (and biotite locally) are present in the matrix with fine grained quartz and feldspar.

Fissile and 'splintery' Snowshoe Group rocks immediately below the Antler Formation in the Mt. Brew area are at least in part siliciclastic rocks of subunit HPSg which have been converted to phyllonites at this location by extremely strong deformation and dynamic recrystallization.

Subunit HPSq. Subunit HPSq consists of quartzites, including those in thin (≥ 30 cm) isolated horizons within mixed pelitic and psammitic rocks of subunit HPSp, and significant beds of quartzite in the siliceous rocks of subunit HPSg.

Quartzites are mainly pale greenish- or bluish-grey to white, and fine to medium grained. Some are well foliated with disseminated mica, whereas other localities are comparatively massive. The quartz is

recrystallized; primary grain size and texture are not discernible. A strong stretching lineation is common in quartzites (see Chapter 3.3).

Subunit HPSP. This subunit is the most widespread and typical in the Snowshoe Group. It comprises well- and thinly bedded phyllitic pelite and psammite. Bedding is defined by fine compositional variations on the scale of a few millimetres, and by beds of psammite a few centimetres thick, giving the rocks a flaggy parting.

Pelites are fine- to very fine grained, very well foliated (cleaved) and usually crenulated. In higher grade areas, coarse grained flecks of mica and porphyroblasts of garnet or chloritoid occur. Pelites are pale- to dark greenish-grey to brownish-grey in colour, depending on whether chlorite or muscovite is dominant. Some pelites are very dark grey, possibly due to very fine graphite. In addition to chlorite and muscovite the pelites generally contain fine- to very fine grained (depending on recrystallization) quartz, feldspar, magnetite, ilmenite (or leucoxene) disseminated in the rock or in finely spaced layers. The pelites are commonly soft, easily weathered and fissile, and have rusty, lustrous cleavage surfaces. Rusty limonite spots and pits are common on both fresh and weathered surfaces. Transposed or blebby quartz segregations are ubiquitous, usually parallel to the layering and the foliation.

Some pelitic rocks are highly flattened and sheared due to deformation, in which the alternating layers of pelite and psammite or equivalents have been strongly thinned and transposed into finely cleaved rocks which could be termed 'phyllonites'.

The interbedded psammities are fine- to medium grained, consisting of equigranular quartz and feldspar, some of which may be 'gritty'.

Psammites contain roughly 10% muscovite and chlorite (locally biotite).

An uncommon variety of the mixed pelite and psammite unit consists of calcareous pelitic phyllite (HPSc). This rock, possibly a marl originally, is soft and friable, fine- to medium-grained, and pale- to medium grey, weathering a streaky rusty orange-brown. It consists mainly of carbonate, with thin discontinuous foliae of chlorite, interbedded with psammite and pelite. Primary layering is strongly transposed in these incompetent rocks, which are usually only a few metres thick.

Subunit HPSk. Subunit HPSk is compositionally similar to HPSp and HPSg, being composed of moderately thickly to thinly interbedded fine grained phyllitic pelite and fine grained, well sorted psammite. The latter may contain some coarser quartz granules but it is never conglomeratic. This unit is distinguished from HPSp because it tends to contain white to pinkish-white quartzite, and also because the pelites and psammites have a distinctive olive to khaki colour (this may be of use to future workers in the area). Otherwise, there may be no significance to the distinction of this subunit: it is usually associated with rocks of subunits HPSp and HPSg. The quartzite was found in only a few localities and could not be traced as a marker.

The phyllite is well cleaved and crenulated, the more competent rocks less so. As is common in the Snowshoe Group, fine iron oxide or limonitic spots are disseminated in the rocks, especially on weathered surfaces.

Subunit HPSm. Subunit HPSm represents rare, thin bands of marble within subunits HPSg or HPSp. These marbles are quite similar to the carbonates of subunit HPSc, but are thicker and more homogeneous (at

least 3 m thick). The marble is creamy white, weathering a buff-brown. It is composed dominantly of medium- to coarse grained calcite, is weakly foliated to massive, and is thinly bedded. Bedding is defined by very thin foliae of ?phlogopitic mica and minor quartz, and by colour variations.

Subunit HPSH. This subunit consists of black psammites and pelites. It is a distinctive rock type which was traced in a roughly 200 m thick band for about 6 km along the comparatively well exposed ridge east of Browntop Mountain. These or very similar rocks were found in other parts of the thesis area, but poor exposure prevented definition of their outcrop. This subunit is the same as that mapped as the Harvey succession by Struik (1983a), and by Campbell (1978) as the ?Midas Formation, although it has no relation to the type-Midas Formation of the Cariboo Group.

Subunit HPSH consists mainly of dark blue-grey to black, fine to medium grained graphitic psammite and pelite. It is composed of quartz, variable amounts of muscovite, and minor feldspar. It is slightly calcareous, evident from minor calcite and rare tremolite. It can be well foliated and crenulated, but the foliation and layering is usually irregular because these rocks are more heterogeneous than typical Snowshoe Group psammites and pelites. Bedding is defined by very thin (a few millimetres) white or grey discontinuous siliceous laminae which are strongly transposed or isoclinally folded parallel to the foliations. Quartz segregations are common. Some psammites are competent and slightly gritty with dark quartz granules, but most are fairly fine grained and flaggy.

Phyllitic pelites in this subunit are also dark grey or black, well cleaved, with very fine compositional layering.

Subunit HPSH locally also contains impure limestone or calcareous siltstone, roughly 10 m thick. This rock is a streaky, dull medium to dark charcoal- or bluish-grey, weathering a grey-buff colour. It consists of fine grained calcite and disseminated graphite and mica. The fine grain size could be tectonic in origin (mylonitic, R.L. Brown, pers. comm.), but this could not be demonstrated by progressive grain size reduction. Coarser calcite segregations occur. Somewhat flaggy bedding is defined by very thin foliae of silicate minerals and by thicker (a few centimetres) colour (presumed compositional-) banding.

Subunit HPSl. This is a streaky, pale- to dark charcoal grey, fine grained impure limestone. It was found in one locality only (829). The colour variation in the limestone is probably related to the proportion of graphite and (or) argillaceous material; the rock is interbedded with dark fine grained pelites. This impure limestone is virtually identical to that found within subunit HPSH, but it lacks the association of the black psammites that dominate subunit HPSH.

Subunit HPSd. This is a very sporadic rock type consisting usually of very pale creamy-grey, hard, fine- to medium grained 'silicate-carbonate'. It consists predominantly of carbonate which has replaced roughly 60% or more of the original rock; the remainder of the rock is sericitized sodic plagioclase, muscovite, fuchsite mica and pyrite. The rock may be moderately foliated, but it is usually massive. It has a rind up to a few centimetres deep of limonite, suggesting that the carbonate is ankeritic.

This subunit locally occurs with other subunits as pods or veins,

but its characteristic occurrence is at the contact between the Snowshoe Group and the Western Quesnel Lake Gneiss. Identical rocks occur as dykes within the gneiss (see Chapter 2.2). Subunit HPSd is most probably altered and carbonatized diabase, judging by its resemblance to rocks known by the author to be of this origin.

Subunit HPSi. This subunit includes a variety of apparently late intrusive rocks which occur sporadically in the Snowshoe Group as dykes or small plugs. Some are similar to subunit HPSd. Other types include a massive dark green-grey, coarse grained diabase-gabbro; a medium grained green-grey ?lamprophyre; and a medium grained, pale green ?meta-diabase.

There is one occurrence of white pegmatite (labelled HPSip on Map 1).

Subunit HPSa. This subunit is a garnet amphibolite, which was found in only one locality (151). It is medium grained, dark green speckled with garnet and fine, creamy-grey plagioclase. It is composed dominantly of amphibole with a hornblende habit, but with the extinction angle of actinolite. Garnet is poikiloblastic and skeletal, and apparently not retrograded. The rock is moderately foliated and lineated, and quite uniform, or with a faint compositional banding. It is interpreted to be a pre-metamorphic mafic intrusion, perhaps a diabase dyke or sill.

It is probably not equivalent to the para-amphibolites mapped by Getsinger (1985) in the area to the north.

APPENDIX B

ON THE SIGNIFICANCE OF THE MEGACRYST ORIENTATION IN THE WESTERN QUESNEL LAKE GNEISS AT LOCALITY 16

The angular distribution of the megacrysts in this gneiss (see Chapter 3.2) was produced during D_1 by simple shear or by a combination of simple shear and pure shear (transpression, in fact; see Harland, 1971). These two alternatives are discussed in this appendix. There is no evidence of significant shortening in the Y direction, from the appearance of the gneiss in the YZ joint faces, so it will be assumed for simplicity that the deformation approximated plane strain, in both cases. The rotational behaviour of rigid inclusions in a viscous medium in simple shear and pure shear was comprehensively investigated by Ghosh & Ramberg (1976), and their method of analysis is applied to the data from the Western Quesnel Lake Gneiss. It is conceded at the outset that the following discussion is only an application of quantitative theory to a natural system in which numerous heterogeneities and perturbations preclude a straightforward analogy with theoretical processes; it also neglects the possible contribution of D_2 flattening. Nevertheless, the exercise confirms the essential conclusion of the kinematic analysis in Chapter 3.2, raises interesting questions, and prompts provisional answers. To quote Ghosh & Ramberg (1976, p. 2):

The theoretical results are rigorous only for isolated rigid inclusions with ellipsoidal shape. Even so, the theory combined with the experimental results are qualitatively applicable to geologic structures involving rotations of inclusions and passive markers.

Ghosh & Ramberg (1976) define s_r as the ratio between pure shear strain rate ($\dot{\epsilon}$) and simple shear strain rate ($\dot{\gamma}$).

SIMPLE SHEAR STRAIN ($s_r = 0$)

If the megacryst distribution in the Western Quesnel Lake Gneiss (Fig. B.1) was produced entirely by simple shear ($s_r = 0$), all megacrysts rotated in a sense synthetic to the shear couple. For megacrysts of a given $R > 1$ (where R = aspect ratio in XZ plane), those that were originally at 90° to the shear plane ($\phi_0 = 0^\circ$; Z axis) rotated faster than those originally at less than 90° , and the higher the R value, the greater this difference in the rotation rate. This is illustrated in Fig. B.2a. Rotation rates decrease as ϕ increases, with those for megacrysts with a high R decreasing at a faster rate than those with a low R . Equant inclusions rotate continuously at constant rate ($\gamma/2$). Theoretically, all inclusions with $R < \infty$ would rotate continuously, even rotating through the shear plane.

The accumulation of megacrysts with high R close to the shear plane can be explained, thus, as a result of early rapid rotation followed by slower rotation and metastabilization within a few degrees of $\phi = 90^\circ$. The potential to keep them rotating is only marginally greater than their high resistance to further rotation - a function of their high R . A comparatively small amount of shear strain would accomplish a lot of rotation for these megacrysts.

Megacrysts with lower R would be spread over the whole range $\phi = 0^\circ$ to 180° , with a less prominent concentration around $\phi = 90^\circ$. This is due to their quasi-constant (and comparatively slow) rotation rates. The potential to keep them rotating does not vary much, and is much greater than their resistance to further rotation, which is limited by their low R . Their finite distribution depends on R , on their original

Fig. B.1

R versus ϕ diagram (axial ratio versus finite angle of rotation) for megacryst distribution in Western Quesnel Lake Gneiss at locality 16. $\phi = 0^\circ$ is approximately vertical, up, and $\phi = 90^\circ$ is subhorizontal and trending east (looking north). Parabolic curves are ϕ_5 and ϕ_6 for all values of R calculated from $s_r = 0.288$, with ϕ_2 at vertex (see text for explanation).

Fig. B.2

(a) Graph of rotation rate of inclusions (w.r.t. shear strain rate) on ordinate versus orientation ϕ , on abscissa, for the case of ideal simple shear ($s_r = 0$). From Ghosh & Ramberg (1976, Fig. 2). (b) Graph of rotation rate of inclusions versus orientation for the case of equal components of pure shear and simple shear ($s_r = 1$). Negative rotation rate = backward rotation, bracketed between ϕ_5 and ϕ_6 for a given $R > R_{crit}$ (see text). From Ghosh & Ramberg (1976, Fig. 3).

orientation, and on the amount of bulk shear strain.

Of the megacrysts in the range $\phi = 90^\circ$ to 180° in Fig. B.1, there is a relatively large number between $\phi = 90^\circ$ and 105° , with a tendency for those of higher R to be closer to the shear plane. They are in this range either because they originated there and did not rotate (synthetically) out of it because rotation rates there are so slow, especially for those with high R (see Fig. B.2a), or they originated in the $\phi = 0^\circ$ to 90° range and rotated through the shear plane. The latter is quite likely for megacrysts with low to moderate R , but very unlikely for those with high R .

COMBINED SIMPLE SHEAR AND PURE SHEAR ($s_r > 0$)

The strain in the Western Quesnel Lake Gneiss may have been produced by simultaneous components of simple shear and pure shear (Fig. B.3a). This case is more complicated than that of ideal simple shear, but it is probably more realistic. The megacryst distribution is interpreted somewhat differently than in the case of $s_r = 0$, providing some interesting insights into the bulk deformation. To begin with, the main characteristics of $s_r > 0$ are as follows.

Unlike ideal simple shear, not all inclusions rotate continuously in combined simple shear and pure shear. Below a critical aspect ratio, R_{crit} , given by

$$R_{crit} = \frac{1 + \sqrt{1 + 4s_r^2}}{2s_r} \quad (\text{eq. B-1})$$

inclusions rotate in a synthetic sense, continuously. However, inclusions above this critical ratio will eventually arrive at a position (ϕ_5) somewhere between $\phi = 90^\circ$ and 135° , depending on R and s_r .

Fig. B.3

(a) Components of the strain where $s_r > 0$: simultaneous simple shear and pure shear. (b) Schematic diagram of zones of inclusion orientation, delineated by varying characteristics of rotation rate and sense, for the case of $s_r > 0$ (see text for explanation). Zone of backward rotation (ϕ_5 to ϕ_6) dotted.

where the rotation vanishes (Fig. B.3b; Fig. B.2b is an example from Ghosh & Ramberg for the case of $s_r = 1$). (Recall that for ideal simple shear, R_{crit} is ∞ , theoretically.)

While most inclusions rotate synthetically ('forwards') with the sense of simple shear, others with $R > R_{crit}$ which originated in a certain angular field, rotate antithetically ('backwards'), eventually coming to rest at their ϕ_5 . Fig. B.3b is based on Ghosh & Ramberg (1976, Figs. 4 and 17), and shows that the fields of forward and backward rotation are delineated by ϕ_5 and ϕ_6 .

ϕ_0 = original position (anywhere between 0 to 180°)

ϕ_1 = position of maximum forward rotation = $\frac{1}{2} \arctan(2s_r)$ (eq. B-2)

ϕ_2 = position of minimum (maximum) forward (backward) rotation = $\phi_1 + 90^\circ$

ϕ_3 and ϕ_4 = positions where all inclusions rotate forward at the same rate ($\dot{\gamma}/2$), which is the constant rate of rotation of an equant inclusion

$$\phi_4 = \frac{1}{2} \arctan\left(\frac{-1}{2s_r}\right) = 90^\circ - \phi_3 \quad (\text{eq. B-3})$$

ϕ_5 = position of vanishing rotation

$$= \arctan\left[-s_r(R^2 - 1) - \sqrt{s_r^2(R^2 - 1)^2 - R^2}\right] \quad (\text{eq. B-4})$$

ϕ_6 = position of no rotation

$$= \arctan\left[-s_r(R^2 - 1) + \sqrt{s_r^2(R^2 - 1)^2 - R^2}\right] \quad (\text{eq. B-5})$$

It is important to note that $\phi_5 = 90^\circ$ (that is, parallel to the shear plane) only for infinitely long inclusions or passive marker lines. All other inclusions of $R > R_{crit}$ rotate synthetically or antithetically until they reach their ϕ_5 between 90° to 135° . ϕ_6 is a special position, where the rotation rate is always zero, and forms the other boundary between the fields of forward and backward rotation. No

inclusion can rotate into this position: it must originate there (ϕ_0).

Finally, the positions of maximum rotation rate (forward, ϕ_1 , or backward, ϕ_2), the positions of stabilization (ϕ_5 and ϕ_6), the relative timing of stabilization, and the angular size of the fields of forward versus backward rotation, are all dependent on R and s_r .

For a given R , ϕ_5 increases with decreasing s_r . For a given s_r , ϕ_5 increases with decreasing R .

Now we can begin to interpret the main features of the megacryst distribution in the Western Quesnel Lake Gneiss, under the condition $s_r > 0$.

The majority of the megacrysts occur in the range $\phi = 0^\circ$ to 90° , the frequency increasing towards $\phi = 90^\circ$, indicating a clockwise (looking north) simple shear component. Theoretically, none of these megacrysts stabilized (cf. Fig. B.2b, where no lines cross the $\phi/\dot{\gamma} = 0$ line between $\phi = 0^\circ$ and 90°), although those between about $\phi = 70^\circ$ and 90° are metastable as their rotation rates were so low. This interpretation is essentially the same as for $s_r = 0$.

The range between $\phi = 90^\circ$ and 180° contains most or all of the following categories (compare Fig. B.3):

(a) Megacrysts which have always rotated forwards (some through the shear plane) without stabilizing because their $R < R_{crit}$.

(b) Megacrysts of $R > R_{crit}$ that were in the process of rotating forwards (some through the shear plane) towards their ϕ_5 , but did not reach it, or

(c) megacrysts that did reach it, and stabilized.

(d) Megacrysts ($R > R_{crit}$) that were originally in the field of backward rotation, and which were in the process of rotating backwards towards their ϕ_5 , but did not reach it, or

(e) megacrysts that did reach it, and stabilized.

(f) Megacrysts ($R > R_{crit}$) that were stranded in the ϕ_6 position.

Which category a megacryst belongs to depends on s_r and where it originated (ϕ_0), neither of which are known.

This is getting complicated. The point is: can we use any of the properties of the megacryst distribution to determine s_r ? For example, can we determine R_{crit} , which would yield s_r from equation B-1, or can we determine any other ϕ value (ϕ_1 to ϕ_6), which would yield s_r from equations B-2 to B-5?

Compare Fig. B.1, and Fig. B.2b which is for the case of $s_r = 1$ (from Ghosh & Ramberg 1976, Fig. 3). In the latter, all inclusions of $\infty > R > R_{crit}$ originating between their ϕ_5 and ϕ_6 would rotate backwards towards their ϕ_5 position. It is proposed here that, after a reasonable amount of shear strain, the majority of these inclusions would lie somewhere between their ϕ_2 position (121.7° , up to which point inclusions had been accelerating backwards) and ϕ_5 , which lies somewhere between ϕ_2 and $\phi = 90^\circ$, depending on their R . Those that didn't reach ϕ_2 probably did not because their R and hence rotation rate were low, but above R_{crit} , of course.

Now, apply this reasoning to the data in Fig. B.1. There is a noticeable feature at about $\phi = 105^\circ$. Beyond 105° , megacrysts have a fairly low R and are quite evenly spread up to $\phi = 180^\circ$. (This group of megacrysts may represent mainly those below R_{crit} .) At $\phi = 105^\circ$ and from there back towards the shear plane, R increases as the frequency of megacrysts increases. These megacrysts are inferred to be in the zone

between ϕ_2 and the shear plane, following the above reasoning. That is, they are (at least in part) megacrysts that originated in the zone of back-rotation between their ϕ_5 and ϕ_6 and have back-rotated at least as far as ϕ_2 , and most have proceeded further on their way towards their ϕ_5 . (Others may have rotated forwards out of the $\phi = 0$ to 90° range, towards their ϕ_5 .)

So, if $105^\circ = \phi_2$, and $\phi_2 = \phi_1 + 90^\circ$, then

$$\phi_1 = 105^\circ - 90^\circ = 15^\circ = \frac{1}{2} \arctan(2s_r) \quad (\text{from eq. B-2})$$

which yields an s_r of 0.288. Using this value, ϕ_5 and ϕ_6 can be calculated from equations B-4 and B-5, as functions of R ; these are plotted in Fig. B.1. R_{crit} can now be calculated from equation B-1, as 3.73. The results are shown diagrammatically in Fig. B.4a, in which ϕ_1 to ϕ_6 have been calculated on the basis of $s_r = 0.288$. Fig. B.4b shows the same analysis in terms of the R - ϕ diagram (cf. Fig. B.1).

It is implied in Fig. B.2b that inclusions significantly below R_{crit} would have a fairly regular, moderate rotation rate, do not ever stabilize, and should be quite evenly distributed between $\phi = 0^\circ$ and 180° . This is reflected in this study: R_{crit} is 3.73, and megacrysts below about $R = 2.5$ are indeed quite evenly distributed.

These calculations are based on the assumption that ϕ_2 is at $\phi = 105^\circ$. If it was more than 105° , s_r would be higher and R_{crit} would be lower. This is unlikely because then the size of the field of backward rotation would be larger, and so there would probably be more megacrysts of moderate R at $\phi > 105^\circ$ than there are. It is more likely that ϕ_2 is less than 105° and that some inclusions didn't quite make it to ϕ_2 , being around 105° . For this reason, it is suggested that the calculated

Fig. B.4

(a) Calculated ϕ values for the observed megacryst distribution, assuming $s_r > 0$ and $\phi_2 = 105^\circ$. $s_r = 0.288$. Values of ϕ_5 and ϕ_6 are shown for an R of 5, for example; they would vary with R , as shown by the parabolic curves in Fig. B.4b, but ϕ_2 is fixed at 105° . ϕ_5 and ϕ_6 converge on ϕ_2 as R decreases, and vice versa. (b) Calculated ϕ values plotted on the $R - \phi$ graph. Zone of backward rotation (towards ϕ_5) lies inside parabola. Compare Fig. B.1. Note that ϕ_1 (shown), ϕ_3 and ϕ_4 (not shown) are independent of R , under the given s_r .

value of s_r is close to the maximum.

Implications

The implication of this analysis of the megacryst distribution, assuming it is the result of combined simple shear and pure shear, is that the lower plate of the Quesnel Lake Shear Zone was subjected to a pure shear (plane strain?) flattening (vertical shortening axis) as well as an upper plate to the east simple shear, as the Quesnel terrane was emplaced during D_1 . The pure shear extension axis would be parallel to the stretching lineation ($\theta = 90^\circ$); this extension could be implicated in the formation of the shear bands in the gneiss because those features have been explained as an "extensional crenulation cleavage" (Platt 1984; see Chapter 3.2).

CONCLUSION

Two interpretations (for $s_r = 0$ and for $s_r > 0$) of the megacryst distribution in the Western Quesnel Lake Gneiss have been presented. In both, the sense of the simple shear component is confirmed as upper plate to the east. In the case of $s_r > 0$, a maximum value of 0.288 is proposed for s_r ; that is, a minimum of about 3.5 times as much simple shear as pure shear.

There is probably no inherent reason to prefer one interpretation over the other, but it is unlikely that s_r was absolutely zero. In conclusion, it is proposed that the strain in the Western Quesnel Lake Gneiss was produced by simple shear, or by simultaneous components of simple shear and pure shear, with a minimum of 3.5 times as much simple shear as pure shear. Some pure shear is indeed implied by the presence of shear band foliation in the gneiss.

6

Metro

MAP I BEDROCK GEO

GEOLOGY

Bedding & D₁ fabrics

Bedding & D₁ fabrics

QUATERNARY VOLC	
QV	b = olivine b
LOWER-MIDDLE	
lM	g = volcanic c = chert pe
UPPER TRIASSIC	
TrJTb	augite p sh
TrJTc	volcanic con
TrJTg	volcanic wic
TrJTt	tuffaceous p
UPPER TRIASSIC	
uTrBP	p = granitic f = quartz
uTrBPc	volcanic sp
PENNSYLVANIAN	
PPA	b = granitic g = gneiss
DEVONIAN	
DMWg	Western Gneiss g = granitic
DMEg	Eastern Gneiss g = granitic
PROTEROZOIC	
	p = pelitic c = conglomeratic

+

qv . . . b = olivine basalt, c = debris flow

g = volcaniclastic wackes, breccia
c = chert-pebble conglomerate, l =

TrJtb augite porphyry basalt flow & flow

TrJTe volcanic conglomerate (debris flow)

TrJtg = volcanic wacke; s = siltstone, l =

TrJTa tuffaceous phyllite, argillite & siltstone
f = volcanic wacke, v = pillowed
s = tuffaceous siltstone, l = interbedded

uTrBP p = graphitic pelite; f = volcanic
l = quartz-feldspar porphyry, d =

uTrBPc volcanic (pyroclastic?) conglomerate

PENNSYLVANIAN-PERMIAN ANTLER FORMATION

PPA : b = greenstone, g = metagabbro,
 s = serpentinite, t = steatite

DEVONIAN-MISSISSIPPIAN QUESNEL LAKE C

DMWQ Western Quesnel Lake Gneiss:
g = granitic orthogneiss

DMEQ Eastern Quesnel Lake Gneiss:
g = granodioritic orthogneiss

PROTEROZOIC-PALEOZOIC SNOWSHOE GROUP

p = pelites, g = psammites & minor
conglomerate, c = carbonate, h = s

LEGEND

low
cia & shale,
= limestone

LA GROUP
low breccia
(low)

t = intrusion
s siltstone;
wed mafic volcanics
intrusive

miclastic schist,
d = meta-diabase
erate

ATION

KE GNEISS

ROUP

ignor pebbly
= subunit HPSb,

Geological contact (definite,
approximate, inferred)

Bedding (or compositional layering):
S₀ or S₀₋₁

S₁ foliation in Quesnel Lake Gneiss

L₁ stretching lineation

Outcrop exposure

Younging indicator

Fault

Cross-section line

Notes

Selection of S₀₋₁, S₁ and L₁ measurements shown. See Map 2 for structural domains and stereograms. Locations of cross-sections indicated: A-A' in northwest to M-M' in southeast. Outcrop exposure symbols (x) omitted for clarity in some ridges and cirques where there is much exposure. Contact between PPA and HPS or DMWQ is a thrust, the Quesnel Lake Shear Zone.

0 1 2 3 4 5 km

UPPER-TRIASSIC BLACK PHYLLITE

uTrBP p = graphitic pelite; f = volcanic
i = quartz-feldspar porphyry, d = m

uTrBPc volcanic (pyroclastic?) conglomerate

PENNSYLVANIAN-PERMIAN ANTLER FORMATION

PPA b = greenstone, g = metagabbro,
s = serpentinite, t = steatite

DEVONIAN-MISSISSIPPIAN QUESNEL LAKE GNEISS

DMWQ Western Quesnel Lake Gneiss:
g = granitic orthogneiss

DMEQ Eastern Quesnel Lake Gneiss:
g = granodioritic orthogneiss

PROTEROZOIC-PALEOZOIC SNOWSHOE GROUP

p = pelites, g = psammites & minor p
conglomerate, c = carbonate, h = gabbro
k = psammites & pelites, q = quartzite
marble, d = meta-diorite, a = amphibolite
l = limestone, i = intrusive (dyke),
ip = pegmatite

elastic schist,
= meta-diabase.

ate

ION

GNEISS

UP
or pebbly
subunit HPSH,
rtzite, m =
phibolite,
(ke),

Outcrop exposure

Younging indication

Fault

Cross-section line

Notes

Selection of S_0 -1, S_1 and L_1 measurements shown. See Map 2 for structural domains and stereograms. Locations of cross-sections indicated: A-A' in northwest to M-M' in southeast. Outcrop exposure symbols (x) omitted for clarity in some ridges and cirques where there is much exposure. Contact between PPA and HPS or DMWQ is a thrust, the Quesnel Lake Shear Zone.

0 1 2 3 4 5 km

E

Tr

OMEQ

A E

$\begin{array}{c} + \\ 121^{\circ}00' \end{array} 52^{\circ}30'$

MAP 2: Structural domains & S₀₋₁ stereograms

MAP 3 D₂ & D₃ fab

fabrics, minor structures & axial surfaces

cs, minor structures & axial surface t

traces

aces

LEGEND

Geological contact (definite, approximate, inferred)

S₂ foliation where different from S₀₋₁ at same locality

S_{0-1x2} intersection lineation or F₂ minor fold axis.

Arrowheads indicate F₂-vergence, as follows:

southwesterly-verging

northeasterly-verging

non-verging (hinge zone)

non-verging but biased towards southwesterly- or
northeasterly-verging

F₂ minor folds, viewed downplunge: "S, Z, M"

Axial surface trace, with number of deformation phase:

F₂ overturned anticline

F₂ overturned syncline

F₃ antiform

'S₃ foliation'

F₃ fold axis

Fault

Seller
Creek
syncline

uTr8P

2

2
31

2

TrJTg

Spanish
Lake
anticline

TrJTb

2

ImJ

TrJTg

TrJTb

F₂ minor folds, viewed downplunge: "S, Z, M"

Axial surface trace, with number of deformation phase:

F₂ overturned anticline

F₂ overturned syncline

F₃ antiform

'S₃ foliation'

F₃ fold axis

Fault

HPs

Spanish
Lake
anticline

TrJTB

Spanish
Lake
anticline

DMEQ

L A K E

+ 52°30'
121°00'

7 of/de 7

Micred

MAP 4

METAMOR

ISOMORPHISM & Locality Map

ISM & Locality Map

LEGEND

Prehnite-pumpellyite - greenschist facies boundary
(symbol on higher grade side)

Tentative biotite first-appearance isograd
(symbol on higher grade side)

Almandine garnet first-appearance isograd
(symbol on higher grade side)

Mineral assemblages in pelitic rocks (simplified)

P.1	=	ch - mu	
P.2	=	cz - ch - mu	
P.3	=	bi - ch - mu	
P.4	=	bi - cz - ep - ch - mu	(all assemblages
P.5	=	ga - cz - ch - mu	contain plagioclase
P.6	=	ga - ct - ch - mu	and quartz)
P.7	=	ga - bi - cz - ch - mu	
P.8	=	st - ga - bi - mu	

Mineral assemblages in mafic rocks (simplified)

M.1	=	pu - pr - ch	
M.2	=	ac - ch	(all assemblages
M.3	=	cz - ch	contain plagioclase)
M.4	=	ep - bi - ch	
M.5	=	ac - cz - ep - bi - ch	

Mineral assemblages in ultramafic rocks (simplified)

U.1	=	se - mg
U.2	=	se - tc - tr - mg
U.3	=	se - tc - mg
U.4	=	tc - mg

Abbreviations

actinolite	ep	epidote	pu	pumpellyite
biotite	ga	garnet	se	serpentine
chlorite	mg	magnesite	st	staurolite
chloritoid	mu	muscovite	tc	talc
clinozoisite	pr	prehnite	tr	tremolite

Notes

A few localities are omitted from map in high density areas, but their positions may be estimated from the pattern of neighbouring numbered localities. See Map 1 for other map symbols and for rock

their
number
units.

M
M
M

U
U
U

ac
bl
ch
cl

A
their
number
units.

A few localities are omitted from map in high density areas, but their positions may be estimated from the pattern of neighbouring numbered localities. See Map 1 for other map symbols and for rock units. See Chapter 4 for more detailed information.

M.1 = cz - ch
M.3 = cz - ch
M.4 = ep - bi - ch
M.5 = ac - cz - ep - bi - ch

all assemblages contain plagioclase

Mineral assemblages in ultramafic rocks (simplified)

U.1 = se - mg
U.2 = se - tc - tr - mg
U.3 = se - tc - mg
U.4 = tc - mg

Abbreviations

actinolite	ep	epidote	pu	pumpellyite
biotite	ga	garnet	se	serpentine
chlorite	mg	magnesite	st	staurolite
chloritoid	mu	muscovite	tc	talc
clinozoisite	pr	prehnite	tr	tremolite

Notes

A few localities are omitted from map in high density areas, but their positions may be estimated from the pattern of neighbouring numbered localities. See Map 1 for other map symbols and for rock units. See Chapter 4 for more detailed information.

0 1 2 3 4 5 km

121°30'

121°

21° 30'

121°15'

844
ga

241

812

+ 52°30'
121°00'

CROSS-SECTIONS

Sheet

CROSS-SECTIONS

Sheet

Spanish Lake
anticline

et 1: A-A' to G-G''

Legend on

S

Black Bear
unf. form

Seller Creek
syncline

uTrBP

PPA

Wolverine
anticline

Seller Creek
syncline

BA

Sheet 1: A-A' to G-G''

Legend

d on Sheet 2

BPS

W4

HPs

Budger Peak
syn

QLSZ

2000

CROSS-SECTION

Spanish Lake
anticline

Seller Creek
syncline

TrJ¹

TrJ²

TrJ³

UTr¹

PPa

2000

6000

4000

feet

CROSS-S

SECTIONS

Sheet 2: H-H RC

HP

Spanish Lake
anticline

0 100
4000
feet

CROSS-SECTIONS

Sheet 2: H-

H-H'' to M-M'

metres

2000

1000

B

TrJ₁

uTr₁

HP

PPA

metres

2000

1000

C

ImJ

TrJ₁

TrJ₁

uTr₁

PPA

PPA

DM

sea eve G

Tr J's

Tr J:

Tr J's

Tr J's

6000
4000
2000

B' sea level

metres

sea level F

6000
4000
2000

C''

? uTr BP

HPs

DM wdg

SLR EVE G

TrJ⁺ TrJ⁺ TrJ⁺ TrJ⁺

2000
metres
1000

sea level F

DM wdg

? uTrBP

HPs

uTrps

PPA

Waverline
anticline

Sevier green
syncline

BA

S₂A

TrJr

uTrps

PPA

feet

C

TrJ⁺

uTrBP

PPA

Wolverine
anticline

Seller Creek
syncline

BA

metres

2000

1000

sea level

TrJ⁺

uTrBP

PPA

feet

S-A

2000

G''

HPs

Budger Peak
Syncline

9.52

6000

feet

4000

2000

F''

HPs

9.52

BFS

6000

TrJ

TrJ

uTr

PP

HP

SLA

4000

TrJ

uTr

PPa

800

600

4000

2000

HP

S-3

PA

6000

4000

2000

feet

J

S-4

HPs

metres

4000 fms

PP:

TrJ₂

uTree

PP:

feet
4000

DM

DM

HP

feet
6000
4000

2
1
0

DM

HP

feet
6000
4000

2000

M

metres

sea level

E

C''

?uTrBP

HP3

MWD3

metres

sea level

E

D'' sea level

QLS2

metres

sea level

2

metres

sea level

uTrBP

uTrBP

SLS

S. 4

6000

feet

4000

2000

J^{II}

BA

TrJ T₀

uTrBP

PPA

TrJ

HP₂

HA

H

HP

All 94
 legend for
 relations
 cross-section
 southeast

HP

All sec
legend for
relationship
cross-section
southeast

LEGEND

Structural symbols

Compositional layering in metasediments or metavolcanics:

S₀₋₁ or S₀₋₁₋₂

S₂ foliation where different from S₀₋₁

S₁ foliation in Quesnel Lake Gneiss

Younging indicator

Axial surface of major fold

(all F₂ except F₃ Blackbear antiform)

3 F₂ minor fold shape (viewed to northwest)

Fault

Abbreviations

BA = Blackbear antiform
BPS = Badger Peak syncline
PA = Pits anticline
QLSZ = Quesnel Lake Shear
SCS = Seller Creek syncline
SLA = Spanish Lake anticline
WA = Wolverine anticline

Explanatory notes

Sections vertical. Vertical scale = horizontal. Locations of cross-sections are for rock units shown on Map 1. For each cross-section, local S₀₋₁ versus S₂ relationships and minor folds are shown just above topographic surface. Structure between sections extrapolated with the aid of a northwestward plunge of 10 ± 5°, except in east (sections K, L, M) where about 20° was used (see F₂, L₂ stereograms).

END

30.06.88

FIN