1

MICROCOPY RESOLUTION TEST CHART
NATIONAL BUREAU OF STANDARDS
STANDARD REFERENCE MATERIAL 1010a
(ANSI and ISO TEST CHART No. 2)

National Library
of Canada

Bibliothèque nationale
du Canada

Canadian Theses Service

Services des thèses canadiennes

Ottawa, Canada
K1A 0N4

CANADIAN THESES

THÈSES CANADIENNES

NOTICE

The quality of this microfiche is heavily dependent upon the quality of the original thesis submitted for microfilming. Every effort has been made to ensure the highest quality of reproduction possible.

If pages are missing, contact the university which granted the degree.

Some pages may have indistinct print especially if the original pages were typed with a poor typewriter ribbon or if the university sent us an inferior photocopy.

Previously copyrighted materials (journal articles, published tests, etc.) are not filmed.

Reproduction in full or in part of this film is governed by the Canadian Copyright Act, R.S.C. 1970, c. C-30.

**THIS DISSERTATION
HAS BEEN MICROFILMED
EXACTLY AS RECEIVED.**

AVIS

La qualité de cette microfiche dépend grandement de la qualité de la thèse soumise au microfilmage. Nous avons tout fait pour assurer une qualité supérieure de reproduction.

S'il manque des pages, veuillez communiquer avec l'université qui a conféré le grade.

La qualité d'impression de certaines pages peut laisser à désirer, surtout si les pages originales ont été dactylographiées à l'aide d'un ruban usé ou si l'université nous a fait parvenir une photocopie de qualité inférieure.

Les documents qui font déjà l'objet d'un droit d'auteur (articles de revue, examens publiés, etc.) ne sont pas microfilmés.

La reproduction, même partielle, de ce microfilm est soumise à la Loi canadienne sur le droit d'auteur, SRC 1970, c. C-30.

**LA THÈSE A ÉTÉ
MICROFILMÉE TELLE QUE
NOUS L'AVONS REÇUE**

STRATIGRAPHY AND STRUCTURE OF THE EAST-CENTRAL
CARIBOO MOUNTAINS, BRITISH COLUMBIA
AND
IMPLICATIONS FOR THE GEOLOGICAL EVOLUTION OF THE SOUTHEASTERN
CANADIAN CORDILLERA

by

Donald Currie Murphy, B.Sc., M.Sc.

A thesis submitted to the Faculty of Graduate Studies
in partial fulfillment of the requirements of the degree
of Doctor of Philosophy

Carleton University
Ottawa, Ontario
September, 1985

© copyright

1985, Donald C. Murphy

Permission has been granted to the National Library of Canada to microfilm this thesis and to lend or sell copies of the film.

The author (copyright owner) has reserved other publication rights, and neither the thesis nor extensive extracts from it may be printed or otherwise reproduced without his/her written permission.

L'autorisation a été accordée à la Bibliothèque nationale du Canada de microfilmer cette thèse et de prêter ou de vendre des exemplaires du film.

L'auteur (titulaire du droit d'auteur) se réserve les autres droits de publication; ni la thèse ni de longs extraits de celle-ci ne doivent être imprimés ou autrement reproduits sans son autorisation écrite.

ISBN 0-315-29867-7

The undersigned hereby recommend to the Faculty
of Graduate Studies and Research acceptance of this
thesis, submitted by Donald Currie Murphy, B.Sc., M.Sc.,
in partial fulfillment of the requirements for
the degree of Doctor of Philosophy.

.....
Chairman, Department of Geology

.....
Supervisor

.....
External Examiner

ABSTRACT

The east-central Cariboo Mountains are underlain by a 10 km-thick structural sequence of deformed and metamorphosed sandstone and grit, pelite, and rare thin limestone of the Proterozoic Kaza Group. The sandstones and grits are interpreted as mid-fan turbidites. The only departure from mid-fan sedimentation is represented by a laterally persistent, siltstone and shale marker unit at the top of the middle subdivision of the Kaza Group. The siltstone unit is interpreted as a clastic shelf deposit; the overlying shale is interpreted as a slope deposit. This marker unit and its sedimentological interpretation is used as a basis for regional correlations.

The structural geometry of the Kaza Group results from four phases of deformation. The most prominent folds (syn-metamorphic, second phase) change from upright and open at shallow structural levels to near-recumbent and tight-to-isoclinal at deeper structural levels. This transition is attributed to a downward increase in the shear component of the bulk strain; strain modelling indicates that the transition can be produced by 6 to 10 km of northeastward displacement of deep structural levels with respect to shallow structural levels.

The amount and distribution of structural thickening in the 10 km-thick structural sequence of the Kaza Group in this region suggests that 45 to 50 km of convergence is accommodated at deep structural levels. A pre-metamorphic peak fold and thrust belt located in the

Queensland Highlands is inferred to be the shallow level manifestation of this convergence.

The southwestward vergence of second phase structures qualifies them as backfolds and backthrusts. Backfolding and backthrusting occurred following the eastward emplacement of an oceanic terrane onto the Cordilleran continental margin but before the eastwardly directed overthrusting in the Rocky Mountain Belt. Rheological profiles of palinspastically restored cross sections of the Cordilleran continental margin at the times of reversal of regional vergence are constructed to determine the controls on the orientation of displacement trajectories and also to evaluate two tectonic models which have been proposed to account for the kinematic evolution of the southeastern Canadian Cordillera.

ACKNOWLEDGEMENTS

Several people deserve recognition for their special contributions to this thesis and also to various other aspects of the five years involved in its research and writing: my supervisor and friend, Dick Brown, who suggested, supported, and patiently supervised all aspects of this thesis, choreographed a headstrong group of graduate students in such a way as to create a stimulating and productive intellectual climate, and on occasion, brought good weather to the Cariboos and gracefully allowed me to whip him in squash; Murray and Carol Journeay and Randy Parrish, with whom many dreams, meals, movies, musical passions, good and bad field days, and ideas were shared and whose incredible generosity permitted me to live in comfort and companionship during the writing of this beast; fellow travellers through the Carleton hallways, who made the trip more enjoyable and stimulating: Larry Aspler, Don Brown, Sharon Carr, Olga Kukal, Larry Lane, Rob Rainbird, Rob Scammell, and Terri Smith; Chris Rees and Gerry Ross, for field companionship and a healthy degree of skepticism that, for the most part, kept my interpretations grounded in reality; Giorgio Ranalli, for patiently encouraging a mathematical illiterate to attempt to combine geology and geodynamics; Dick Campbell of the Geological Survey of Canada for his enthusiastic interest, discussions, patience with ideas from someone who has set foot on probably less than one one-thousandth of the parts of the Cordillera that Dick knows from first hand experience, and, importantly, for providing funds for the project; Hugh Greenwood of the University British Columbia, for generously permitting my extended visit to U.B.C.; Al Donaldson, for discussion on

the interpretations of sedimentary structures even though I didn't know the difference between a symmetrical and asymmetrical ripple mark when I started this thing; squash partners, Henry Dillon-Leitch, Frank Duimovich, Gerry Ross, Rob Scammell, Roger Skirrow, and Brad Wilson, whose games gave me something to look forward to at 8:00 AM; Ernie Buehler of Cariboo Lodge, for letting me use the lodge as a radio base for my first two field seasons; Maryann Snowden, for being the voice at the other end of the radio and being a great expediter; Garry Forman and Todd McCready of Yellowhead Helicopters, for safe flying and only forgetting us once; Rob Berman and Martin Engi of U.B.C., for use of their computer account to support my word processing at U.B.C.; Jack Hogg of Carleton University and Ed Montgomery of U.B.C., for knowing how and where to do things when no one else had a clue; Sally Hill, who looked after my in absentia registration and all those rules for me; Dad and Mom, for making sure that I was educated enough to follow my interests to the extent that I have and for their support in patiently waiting for their 30 year old son to get out of school; Ray Price of the Geological Survey of Canada, who had no active involvement in this work, but whose ideas and geological intuition permeated most aspects of the thesis; and Bert Struik of the Geological Survey of Canada, for interesting discussions and a review of an early draft of Chapter 3.

I appreciate most of all the love and patience of my companion, Joan Turner, who, in spite of long separations with too few letters, my tendencies for chaos, and passion for geology, still wants me to explore life with her.

TABLE OF CONTENTS

TITLE PAGE	1
ACCEPTANCE PAGE	11
ABSTRACT	111
ACKNOWLEDGEMENTS	v
TABLE OF CONTENTS	vii
LIST OF FIGURES	xi
LIST OF TABLES	xiv
LIST OF PLATES	xv
LIST OF FIGURES IN POCKET	xvi
 CHAPTER 1. INTRODUCTION	 1
(1-1) Introductory Statement	2
(1-2) Setting of the Study Area	3
(1-2a.) Location, physiographic sketch, and access	3
(1-2b.) Geological setting	5
(1-3) Previous Work	10
 CHAPTER 2. STRATIGRAPHY AND REGIONAL CORRELATION OF WINDERMERE (LATE PROTEROZOIC) "GRIT" UNITS, CARIBOO MOUNTAINS AND ROCKY MOUNTAIN MAIN RANGES, BRITISH COLUMBIA	 12
(2-1) Abstract	13
(2-2) Introduction	14
(2-3) Geological Setting of Windermere "grit" units, Cariboo and Rocky Mountains	18
(2-4) Previous work	21
(2-5) The Kaza Group in the central Cariboo Mountains	22

(2-5a.) Structure, metamorphism, and contact relationships	22
(2-5b.) Depositional model for the interpretation of the sedimentary facies	29
(2-6) Stratigraphy and depositional environments of the Kaza Group	30
(2-6a.) Lower Kaza Group	30
(2-6a-1.) Stratigraphy	30
(2-6a-2.) Depositional environments	31
(2-6b.) Middle Kaza Group	34
(2-6b-1.) Stratigraphy	34
(2-6b-2.) Depositional environments	43
(2-6c.) Upper Kaza Group	47
(2-6c-1.) Stratigraphy	47
(2-6c-2.) Depositional environments	48
(2-7) Regional correlation of the Kaza Group	50
(2-7a.) Distribution of sedimentary environments in the Kaza Group of the central Cariboo Mountains	50
(2-7b.) Kaza Group of the southern Cariboo Mountains	50
(2-7c.) Middle Miette Group of the Rocky Mountain Main Ranges	52
(2-8) Paleogeographic interpretation	55
(2-9) Discussion	60
(2-10) Conclusions	62
Acknowledgements	62
 CHAPTER 3. SUPRASTRUCTURE/INFRASTRUCTURE TRANSITION, EAST-CENTRAL CARIBOO MOUNTAINS, BRITISH COLUMBIA: GEOMETRY, KINEMATICS, AND TECTONIC IMPLICATIONS	 64
(3-1) Abstract	65

(3-2)	Introduction	66
(3-3)	The structural sequence of the Kaza Group	70
(3-3a.)	Suprastructure	80
(3-3b.)	Infrastructure	86
(3-3c.)	Transition zone	91
(3-4)	Kinematic analysis	93
(3-4a.)	Transition from upright to recumbent folding	93
(3-4b.)	Amount and distribution of structural thickening	97
(3-4c.)	Magnitude of crustal displacements	100
(3-5)	Tectonic implications	102
(3-6)	Conclusions	112
	Acknowledgements	113
CHAPTER 4. LITHOSPHERIC STRENGTH AND THE KINEMATIC EVOLUTION OF THE SOUTHEASTERN CANADIAN CORDILLERA		114
(4-1)	Abstract	115
(4-2)	Introduction	116
(4-3)	Kinematic history of the southeastern Canadian Cordillera (early to late Jurassic)	118
(4-4)	The Jurassic continental margin	127
(4-4a.)	Early Jurassic (pre-Sinemurian): geometry	127
(4-4b.)	Early Jurassic (pre-Sinemurian): thermal structure	133
(4-4c.)	Early Jurassic (post-Sinemurian): geometry	135
(4-4d.)	Early Jurassic (post-Sinemurian): thermal structure	136
(4-4e.)	Middle Jurassic: geometry	137
(4-4f.)	Middle Jurassic: thermal structure	138

(4-5) Mechanical properties of the Jurassic continental margin	141
(4-5a.) Analytical method	141
(4-5b.) Calculations of lithospheric strength	141
(4-5c.) Results	146
(4-5c.-1) Early Jurassic (pre-Sinemurian)	146
(4-5c.-2) Early Jurassic (post-Sinemurian)	154
(4-5c.-3) Middle Jurassic	154
(4-6) Comparison of model rheological stratigraphy with observed patterns/of deformation	154
(4-6a.) Early Jurassic (pre-Sinemurian)	155
(4-6b.) Early Jurassic (post-Sinemurian)	156
(4-6c.) Middle Jurassic	158
(4-7) Implications of the rheological structure of the lithosphere for the kinematic evolution of the southeastern Canadian Cordillera	159
(4-8) Discussion	165
Acknowledgements	167
CHAPTER 5. SUMMARY OF CONCLUSIONS	168
REFERENCES	174

LIST OF FIGURES

FIGURE	DESCRIPTION	PAGE
1-1	Physiographic subdivisions of southern British Columbia	4
1-2	Tectonic subdivisions of the Canadian Cordillera	7
2-1	Regional geological map of the Cariboo Mountains and adjacent Rocky Mountains and Quesnel Highlands	17
2-2	Cross section of Cariboo Mountains and adjacent Rocky Mountains and Quesnel Highlands	20
2-3	Distribution of rock units in the east-central Cariboo Mountains and locations of areas mapped detail	23
2-4	Longitudinal (NW-SE) cross section of east-Cariboo Mountains	24
2-5	Composite vertical cross section across areas A to F shown in Fig. 2-3	27
2-6	Sketch illustrating how Kaza Group rocks at deep structural levels would restore further to the west than Kaza Group rocks at shallow structural levels	28
2-7	Stratigraphy of the upper middle, and lower upper Kaza Group	pocket
2-8	Variation of depositional environments along line of section shown in Fig. 2-4	51
2-9	Correlation of sections of the Kaza and middle Miette Groups, Cariboo and Rocky Mountains	53
2-10	Preliminary reconstruction of depositional environments during the deposition of the marker unit at the top of the middle Kaza Group	57
3-1	Regional geological map of the Cariboo Mountains and parts of the adjacent Rocky	

	Mountains and Quesnel Highlands	68
3-2	Distribution of rock units in the east-central Cariboo Mountains and locations of areas mapped in detail	71
3-3	Longitudinal (NW-SE) cross section of the east-central Cariboo Mountains	72
3-4	Perspective diagram illustrating the ship between Figs. 3-2, 3-3, and 3-5	73
3-5	Composite vertical cross section across areas A through F shown in Fig. 3-2	75
3-6	Stratigraphy and regional correlation of the Kaza and middle Miette Groups	77
3-7	Sequence of deformation and metamorphism in the Cariboo Mountains	79
3-8	Equal-area stereoplots of D_2 planar and linear fabric elements	82
3-9	Kinematic analysis of transition from upright folds and foliations to near-recumbent folds and gently dipping foliations	96
3-10	Diagrammatic model illustrating the relationship between infrastructural ductile thickening and thrust faulting in the Quesnel Highlands	99
3-11	Areas in the southeastern Canadian Cordillera in which regional-scale southwestwardly verging structures have been described	103
3-12	One tectonic model for the formation of southwestwardly verging structures in the southeastern Canadian Cordillera	106
3-13	Alternate tectonic model for the formation of southwestwardly verging structures in the southeastern Canadian Cordillera	108
4-1	Regional geological map of the Cariboo Mountains and adjacent Rocky Mountains and Quesnel Highlands	120
4-2	Cross section A - A' illustrating structural relationships from cratonic North America to within Terrane I	123
4-3	Scaled reconstructions of the Jurassic	

	continental margin (Model 1)	130
4-4	Scaled reconstructions of the Jurassic continental margin (Model 2)	131
4-5	Rheological evolution of the Jurassic continental margin: Early Jurassic (pre-Sinemurian), Model 1	148
4-6	Rheological evolution of the Jurassic continental margin: Early Jurassic (pre-Sinemurian), Model 2	149
4-7	Rheological evolution of the Jurassic continental margin: Early Jurassic (post-Sinemurian), Model 1	150
4-8	Rheological evolution of the Jurassic continental margin: Early Jurassic (post-Sinemurian), Model 2	151
4-9	Rheological evolution of the Jurassic continental margin: Middle Jurassic, Model 1	152
4-10	Rheological evolution of the Jurassic continental margin: Middle Jurassic, Model 2	153
4-11	Comparison of strengths of abandoned displacement trajectories with strengths of new displacement trajectories at the time of changeover.	2

LIST OF TABLES

TABLE	DESCRIPTION	PAGE
1	Summary of geothermobarometric data used to constrain the mid-Jurassic geotherms	139
2	Creep parameters for rock types used in analysis	145

LIST OF PLATES

Plate 2-1	33
Plate 2-2	39
Plate 2-3	42
Plate 3-1	84
Plate 3-2	90

LIST OF FIGURES IN POCKET

- Figure 2-7 Stratigraphy of the upper middle and lower upper Kaza Group
- Figure 4-3 Scaled reconstructions of the Jurassic continental margin (Model 1)
- Figure 4-4 Scaled reconstructions of the Jurassic continental margin (Model 2)
- Figure 4-5 Rheological evolution of the Jurassic continental margin
to
Figure 4-10
- Fig. P-1: Structural geometry of the area between Castle Creek and Mica Mountain (North Half): Map and serial vertical cross sections
- Fig. P-2: Structural geometry of the area between Castle Creek and Mica Mountain (South Half): Map and serial vertical cross sections
- Figs. P-3 through P-9: Orientation of fabric elements, Castle Creek to Mica Mountain: Map and equal-area stereoplots, Areas A through F (Figs. 2-3, 3-2)

CHAPTER 1

INTRODUCTION

2

(1-1) INTRODUCTORY STATEMENT

This thesis is a study of various aspects of the stratigraphy, structural geology, and geodynamic evolution of the Cariboo Mountains and adjacent Quesnel Highlands and Rocky Mountain Main Ranges of eastern British Columbia (Fig. 1-1). It is composed of three self-contained papers, one on the stratigraphy, sedimentology, and regional correlation of the Kaza Group, the "grit" unit of the Windermere Supergroup which outcrops extensively in this region; one on the geometry, kinematics, and tectonic implications of second phase (D_2) deformation in the Cariboo Mountains and Quesnel Highlands; and the third on the rheological structure of the lithosphere underlying this region during the Jurassic and its influence on Cordilleran orogenesis.

Most of the conclusions of Chapters 2 and 3 are based on data collected and analyzed during 6 months of field mapping in the east-central Cariboo Mountains (Fig. 1-1; summers of 1981, 1982, and 1983) and several months of petrography and library research at Carleton University and the University of British Columbia between 1981 and 1985. Some of the working hypotheses developed for the east-central Cariboo Mountains had specific implications for the geology of neighboring regions; for this reason, much of the non-field research was directed to understanding the regional geology well enough to assess the regional validity of these hypotheses. In many cases, it was found that certain working hypotheses were able to account for geological observations in the surrounding area; as such, I was

sufficiently encouraged to broaden the analysis to a regional scale.

Chapter 4 evolved out of several months of balancing cross sections across the Cordillera at the latitude of the Cariboo Mountains and adjacent Quesnel Highlands and Rocky Mountain Main Ranges for a regional synthesis paper co-authored with colleagues at Carleton (Brown, et al. in press), and discussions with Giorgio Ranalli about models of the strength of the lithosphere in different tectonic environments. Although providing numerous insights into geodynamic and orogenic processes, recent geodynamic models are somewhat unsatisfying, suffering from a degree of oversimplification that is not in line with the complexity evident in geological studies. Cross sections that are balanced to deep structural levels and their palinspastic restorations present an opportunity to attempt more "realistic" geodynamic models. Chapter 4 represents an attempt to integrate, with the use of balanced cross sections, field-based geometric and kinematic studies with analytical geodynamic models.

(1-2) SETTING OF THE STUDY AREA

(1-2a.) Location, physiographic sketch, and access

The Cariboo Mountains are the northernmost of an en-echelon series of major mountain ranges that lie southwest of the Southern Rocky Mountain Trench between the "big bend" of the Fraser River and the Canada - U.S.A. border (Fig. 1-1). They are bounded on the west and southwest by the Quesnel Highlands, forested highlands that are transitional between the Interior Plateau and the Cariboo Mountains. The North Thompson River is the southeastern boundary of the Cariboo

Figure 1-1: Physiographic subdivisions of southern British Columbia (modified from Wheeler et al., 1972). Area of Fig. 2-1 shown with stippled pattern.

Mountains, separating them from the Monashee Mountains.

The area examined is bounded on the southeast by the Premier Range and on the northwest by the ridge north of Castle Creek (Fig. 1-1). The relief in the southern part of the area is significant: Premier Range peaks greater than 3000 m are separated by steep walled valleys whose floors are commonly over 2000 m lower in elevation. Glacial and snow cover is significant in the southern part of the area. Maximum elevations decrease to the northwest although relief is still considerable. Several deep, steep walled, forested valleys (Kiwa Creek, Raush River, Castle Creek and Black Martin Creek) divide the northern part of the area into upland "massifs" in which the internal relief is less than the relief at their boundaries.

Above treeline (approximately 1750 m on southern faces), glaciated alpine topography predominates and rock exposure is excellent. Valley walls and bottoms are characterized by poor exposure and are commonly accessible from ridges only with considerable effort and expenditure of time; as such, they were examined only when the opportunity of following a distinctive marker presented itself. Because of the lack of such markers in the Kaza Group, such opportunities were rare.

With the exception of roadwork in the Southern Rocky Mountain Trench and in certain valleys draining into the Trench (Kiwa and Castle Creeks), mapping was done exclusively on foot, out of camps which were placed by helicopter from Valemount.

(1-2b.) Geological setting

The east-central Cariboo Mountains lie within the Omineca Belt, the mountainous NNW - SSE trending belt of metamorphosed, deformed, and

Figure 1-2: a. Tectonic subdivisions of the Canadian Cordillera
(after Monger et al. 1972)

b. Major "suspect terranes" of the Canadian Cordillera
(after Monger 1984)

1. Alexander	8. Monashee
2. Bridge River	9. Olympic
3. Cassiar	10. Pacific Rim
4. Cache Creek	11. Quesnel
5. Chugach	12. Slide Mountain
6. Cascade	13. Stikine
7. Kootenay	14. Wrangell
15. Yukon-Tanana	

BB Bowser Basin

Boundary between Superterranes I and II

c. Thrust faults of the Canadian Cordillera (after Monger 1984)

- westwardly verging
- eastwardly verging

Tr, Triassic; JK, Jurassic or Cretaceous; LJ, late Jurassic; eT, early Tertiary

d. Strike-slip faults of the Canadian Cordillera (after Gabrielse 1985a)

KO, Kutchok; TN, Teslin (Thibert to S); KA, Kechika; F, Finlay (Thudaka to N); TA, Takla; P, Pinchi; Y, Yalakom; Pa, Pasayten; H, Hozameen; SC, Straight Creek; FC, Fraser Canyon; NRMT, Northern Rocky Mountain Trench; QCF, Queen Charlotte Fault

area of Fig. 2-1

Figure 1-2

intruded rocks that lies between the weakly to un-metamorphosed rocks of the Intermontane Belt and the Rocky Mountain Fold and Thrust Belt (Fig. 1-2a). The Omineca Belt comprises deformed and metamorphosed Proterozoic to Mesozoic sedimentary and volcanic rocks that have been correlated with the North American continental margin sequence of the Rocky Mountain Fold and Thrust Belt; Mesozoic and Tertiary plutonic rocks; deformed and metamorphosed sedimentary, volcanic, and plutonic rocks of volcanic arc, marginal basin, or oceanic affinity which are, with local exceptions, paleontologically and paleomagnetically distinct from North American rocks; and a sequence of deformed meta-sediments, meta-volcanics, and meta-plutonic rocks with a history of Paleozoic deformation, metamorphism, and intrusion that is not recorded in either the North American rocks or the oceanic rocks. The latter two tectonostratigraphic successions have been referred to as "suspect" terranes (Coney et al. 1980, Monger et al. 1982, Monger 1984) to emphasize the ambiguity of their relationship with respect to North America; Fig. 1-2b. is a map of the distribution of the major terranes of the Canadian Cordillera and parts of Alaska (Monger 1984).

With the exception of locally important Paleozoic and Tertiary deformation, the structural and metamorphic characteristics of the Canadian Cordillera, in general, and the Omineca Belt, in particular, are the result of early Jurassic to early Tertiary oblique convergence between North America and the suspect terranes (Monger et al. 1982). Over 450 km of convergence between North America and the westernmost suspect terrane is recorded in structures associated with early Jurassic obduction of the easternmost suspect terranes onto the continental margin (60 to 100 km, Brown et al. in press) and subsequent

9

deformation of the continental margin (50 to 75 km of mid-Jurassic shortening in the Omineca Belt; 125 to 200 km of mid-Jurassic to early Tertiary shortening in the Rocky Mountain Fold and Thrust Belt; Price 1981; Brown et al. in press). The paleomagnetic signature of the suspect terranes implies as much as 2000 km of dextral strike-slip displacement between North America and the suspect terranes during the mid-Cretaceous to Oligocene interval (Irving & Yule 1972, Packer & Stone 1974, Hillhouse 1977, Irving 1979, Irving et al. 1980, Van der Voo et al. 1980, Rees et al. 1985). Deformation and metamorphism in the Omineca Belt have been interpreted to be primarily a consequence of the component of convergence (Chapters 3 and 4, Monger et al. 1982, Brown et al. in press) although the possibility that the mid-Jurassic ductile fabrics of the Omineca Belt record a strike-slip component, as has most recently been suggested by Struik (1985a), has not yet received adequate attention.

The eastern physiographic boundary of the Cariboo Mountains is the 750 km-long curvilinear structural trough referred to as the Southern Rocky Mountain Trench (SRMT). Various hypotheses on the origin of the SRMT have been proposed (reviewed in North and Henderson 1954) but thick Quaternary alluvium has precluded a definitive analysis. A post-metamorphic (post mid-Jurassic), steeply dipping, down-to-the-east normal fault along the western margin of the SRMT can account for stratigraphic and metamorphic differences across the segment of the SRMT between McBride and Valemount, but strike-slip faulting at the location of the SRMT before thrusting in the Rocky Mountains cannot be ruled out (Murphy 1984). More recently, Chamberlain and Lambert (1985) have proposed that the SRMT is the physiographic expression of a former

trench and strike-slip fault system which accommodated 1500 to 2000 km of northward displacement between Cordillera, a newly proposed "superterrane" which includes all the supracrustal rocks west of the Purcell and equivalent faults, and North America. According to Chamberlain and Lambert, the supracrustal boundary between Cordillera and North America is presently located at the Purcell and related faults in the Rocky Mountain Fold and Thrust Belt and the deep crustal boundary is located at the SRMT. The distance between the two locations is a measure of the amount of post-strike-slip thrust displacement in the Rocky Mountain Fold and Thrust Belt.

The interesting interpretation of Chamberlain and Lambert (1985), if nothing else, serves to further highlight that supracrustal sedimentary and meta-sedimentary rocks on opposite sides of the SRMT at this latitude are sufficiently similar as to cast doubt on assertions of post-thrusting strike-slip displacements greater than a few kilometres. Some of the conclusions of this thesis are based on the assumptions that there are no major strike-slip faults between the Quésnel Highlands and the Rocky Mountain Foothills (Fig. 1-1) and that there is minimal longitudinal (NW-SE) strain. If future work should identify syn- to post-shortening strike-slip faults or major deviations from plane strain, these conclusions would have to be altered; however, the local stratigraphic, structural, and, to some extent, geodynamic analyses would not be invalidated.

(1-3) PREVIOUS WORK

The exploration of the Cariboo Mountains commenced with railroad

surveys (Palmer 1918) and mountaineering expeditions (Holway 1917, Carpe 1925); the first recorded geological observations in the study area were made by Chamberlin (1925) during a mountaineering expedition in 1924. The area has since been mapped at reconnaissance (1:250,000) scale by Campbell (1968) and Campbell et al. (1973), and locally in more detail by Brown (1968).

A detailed account of previous studies is presented as part of the introduction to appropriate chapters in the thesis.

CHAPTER 2

STRATIGRAPHY, PALEO-ENVIRONMENTS, AND REGIONAL CORRELATION OF
WINDERMERE (LATE PROTEROZOIC) "GRIT" UNITS, CARIBOO
MOUNTAINS AND ROCKY MOUNTAIN MAIN RANGES, BRITISH COLUMBIA

(2-1) ABSTRACT

The east-central Cariboo Mountains are underlain by a 10 km-thick structural sequence of metamorphosed and strained quartzofeldspathic sandstone and grit, calcareous and non-calcareous pelite, and rare thin limestone of the Proterozoic Kaza Group. Three major lithostratigraphic subdivisions of the Kaza Group have been recognized: a basal sandstone and shale unit, a middle sandstone and shale unit, and an upper sandstone unit that is transitional with the overlying shales and limestones of the Isaac Formation of the Cariboo Group. The middle and upper units have been further subdivided; one distinctive siltstone and shale marker unit is present regionally and is used as the basis for regional correlation of Windermere "grit" units of the Cariboo and Rocky Mountains.

On the basis of bedding shape and thickness trends, sandstone to shale ratio, and completeness of Bouma sequence, the turbiditic sandstones and "grits" of the Kaza Group of the east-central Cariboo Mountains are interpreted to have been deposited on the mid-fan portion of a turbidite fan, with both outer mid-fan apron and sandstone lobe deposits represented. The only departure from mid-fan sedimentation appears to be the basal part of the marker unit, which is interpreted to be a clastic shelf deposit, sedimented during regression and consequent outward progradation of more proximal sedimentary environments over the more "basinal" turbidite fan.

The variation in thickness and distribution of sedimentary facies in the "grit" unit of the Cariboo Mountains and adjacent Rocky Mountain

Main Ranges suggest the presence of a basin hinge line east of the rocks now exposed in the Cushing Creek area of the Rocky Mountain Main Ranges. The hinge line of the Cambrian to Devonian part of the Cordilleran miogeocline is presently located near the Cushing Creek area; the difference in the location of the two hinge lines may explain the difference between the restored palinspastic location of the Paleozoic hinge line and the transition in geophysical properties that has been interpreted to represent the transition from craton to transitional or oceanic crust.

2-2) INTRODUCTION

In a recent paper on the subsidence history of the early Paleozoic miogeocline of the southern Canadian Cordillera, Bond & Kominz (1984) presented evidence for a late Proterozoic to early Cambrian (600 m. y.) age of continental breakup and initiation of seafloor spreading at the continental margin of western North America. This conclusion focuses attention on the origin of the underlying Proterozoic Windermere Supergroup, which had been interpreted by previous workers to be either the basal clastic unit of the Cordilleran miogeocline, deposited following mid-Proterozoic (850 m. y.) continental rifting and separation (Stewart 1972, 1976; Gabrielse 1972, Monger et al. 1972, Young et al. 1973, Dickinson 1974, Sears & Price 1979) or a separate unconformity-bounded cycle of extension and possibly rifting (Price & Fermor 1981, Pell 1982, Price 1982, Gabrielse 1985b). In addressing this issue, Bond & Kominz (1984) offered three hypotheses: (1) the Windermere Supergroup was deposited during an earlier complete cycle of

crustal extension, rifting, and formation of a passive continental margin unrelated to the latest Proterozoic cycle of rifting, such as was proposed by Price & Fermor (1981) and Price (1982), (2) the Windermere Supergroup represents the early portion of a prolonged phase of extension that culminated with rifting and continental separation in the latest Proterozoic, and (3) the Windermere Supergroup of the southern Cordillera is younger than the radiometric ages from the base of the sequence (Miller et al. 1973, Armstrong et al. 1982; Evenchick et al. 1984) would indicate, and is related to attenuation before the 600 Ma episode of rifting implied by the subsidence data. To choose among these and other possibilities, knowledge of the stratigraphy, depositional environments, and contact relationships of the Windermere sequence is essential.

In the southern Canadian Cordillera, much of the Windermere Supergroup is deformed and metamorphosed, hampering the collection of primary stratigraphic and sedimentological data. However, throughout the Cariboo Mountains and adjacent Main Ranges of the Rocky Mountains (Fig. 2-1), the Windermere Supergroup is widely exposed and only weakly deformed and metamorphosed. The purpose of this paper is to present stratigraphic and sedimentological data from the voluminous, blue quartz-bearing, sub-arkosic "grit" unit of the east-central Cariboo Mountains, the Kaza Group. These data are herein used to propose new correlations with other sections of the Windermere "grit" unit (Kaza Group of the southern Cariboo Mountains and middle Miette Group of the Rocky Mountains) and to constrain Windermere paleogeography.

The regional correlations and paleogeographic inferences are based

Fig. 2-1 Regional geological map of the Cariboo Mountains and parts of the adjacent Rocky Mountains and Quesnel Highlands (compiled and modified from Wheeler et al. 1972, Pell 1984, and Struik 1985b)

Unornamented box in center of figure outlines area shown in Fig. 2-3.

Fig. 2-1 Regional geological map of the Cariboo Mountains and adjacent Rocky Mountains and Quesnel Highlands

on a 200 m-thick siltstone and fine sandstone - limestone - shale marker that has been traced throughout the east-central Cariboo Mountains and is believed to be present in published stratigraphic sections of the southern Cariboo and Rocky Mountains. This marker unit is, as yet, the most distinctive lithofacies within the entire 10 km structural thickness of the Kaza Group. Because adjacent eastern facies of the Kaza and middle Miette Groups were eroded during the time represented by the unconformity at the base of the overlying Gog Group or are found in deeply buried thrust sheets not yet exposed at the surface, the marker unit is an important clue in the reconstruction of Windermere paleogeography.

(2-3) GEOLOGICAL SETTING OF THE WINDERMERE "GRIT" UNIT, CARIBOO AND ROCKY MOUNTAINS

The continental margin sediment prism of northwestern North America was deformed and thrust eastward onto the North American cratonic margin during early Jurassic to Eocene interaction of North America with oceanic terranes to the west (Monger et al. 1982, Brown & Read 1983, Price 1983, Brown et al. 1984 and in press). Shortening of the continental margin initiated with eastward overthrusting of a composite upper Paleozoic to early Jurassic marginal basin and volcanic arc terrane (Terrane I of Monger et al. 1982) onto the outer portion of the continental margin sequence. Subsequent shortening, associated with regional metamorphism, was accommodated by southwestwardly verging deformation of the central and outer portion of the continental margin prism and the obducted marginal basin sequence. The final stage in the shortening of the continental margin was late- and post-metamorphic

eastward displacement of the deformed outer, and undeformed inner, parts of the continental margin prism on foreland propagating thrust faults of the Rocky Mountain Fold and Thrust Belt.

Windermere Supergroup rocks of the Cariboo and Rocky Mountains occur in the structurally highest thrust sheet of the Rocky Mountain Fold and Thrust Belt (Fig. 2-2). Although imbricated by thrust faults in the Rocky Mountains and Quesnel Highlands and offset along high-angle faults in the vicinity of the southern Rocky Mountain Trench, this sheet is interpreted to extend from the Snake Indian Thrust in the Rocky Mountains westward to the Pleasant Valley Thrust of the Quesnel Highlands (Figs. 2-1, 2-2). Kaza Group rocks of the western and central portions of this thrust sheet are deformed by thrust faults and folds associated with the early (pre- and syn-metamorphic) stages in the shortening of the continental margin. At deeper structural levels in the central part of the sheet, Kaza Group rocks are also deformed by late- to post-metamorphic northeastwardly verging folds and faults associated with early motion on faults feeding into the Rocky Mountain Foreland Belt. In the shallow structural levels of the eastern part of the thrust sheet, middle Miette Group rocks exhibit only the later northeastwardly verging fabrics and faults (Carey 1984, Leonard 1984, McDonough & Simony 1984, Mountjoy et al 1985); earlier, southwestwardly verging structures occur at deeper structural levels in this region, but their relationship to pre- and syn-metamorphic structures in the Cariboo Mountains has yet to be established. Between the Rocky Mountain Trench and the Snake Indian Thrust, this sheet is imbricated by several thrust faults with minor displacement (Campbell et al 1973,

Fig. 2-2 Cross section of Cariboo Mountains and adjacent Rocky Mountains and Quesnel Highlands. Line of section shown in Fig. 2-1. Ornamentation as in Fig. 2-1. Numbered locations discussed in Chapter 4. BSS = Black Stuart Synclinorium; LA = Lanezi Arch; ILS = Isaac Lake Synclinorium. Rocky Mountain portion of the cross section modified from Mountjoy (1980) and Campbell et al. (1982). Quesnel Highlands portion of section modified from Spurr (1980, 1982a, 1983a, and 1983b), Rees (1981), and Rees and Ferri (1983).

Mountjoy 1980, Carey 1984, Carey & Simony 1984).

(2-4) PREVIOUS WORK

Reconnaissance-scale lithostratigraphic information on the Kaza and middle Miette Groups has been published in Mountjoy (1962, 1970, 1971, 1976a and 1976b), Slind & Perkins (1966), Gabrielse (1972), Campbell et al. (1973), and Young et al. (1973). These early studies determined that the "grit" unit consists predominantly of interbedded sub-arkosic sandstone (quartzofeldspathic gneiss) and shale (slate/phyllite/schist) with subordinate pebbly sandstone, conglomerate, and limestone (marble). Sandstone and conglomerate were interpreted as turbidites and grain flows based on commonly observed graded bedding and variably complete Bouma cycles. Interbedded pelitic units, locally containing thin limestones, were interpreted as quiet marine, possibly pelagic, deposits, although few data were collected that discriminate between lacustrine or marine shelf, slope, or basinal settings. These early studies of the "grit" unit were not sufficiently detailed to subdivide the unit or to discriminate turbidite facies, determine their geometry, or show their variations in time and space.

More detailed stratigraphic and sedimentological information on the Windermere "grit" unit is available for the three areas shown in Fig. 2-1. In the southern Cariboo Mountains, Pell & Simony (1984) and Pell (1984), described a complete section of the Kaza Group, subdivided it into upper, middle, and lower units and proposed regional correlations with other Proterozoic rocks of the southern Canadian Cordillera. Murphy & Journeay (1982) and Murphy & Rees (1983) described parts of

the Kaza Group in the central Cariboo Mountains and interpreted the sedimentary facies in terms of a turbidite fan depositional model. Carey & Simony (1984) and Carey (1984) described a section of the middle Miette Group in the Cushing Creek area of the western Rocky Mountain Main Ranges and also applied the turbidite fan model to discriminate specific depositional environments.

(2-5) THE KAZA GROUP IN THE CENTRAL CARIBOO MOUNTAINS

(2-5a.) Structure, metamorphism, and contact relationships

In the central Cariboo Mountains, the Kaza Group has been deformed into a northwestwardly plunging structural sequence approximately 10 km-thick (Brown, et al. 1984 and in press, Murphy in press; Figs. 2-3, 2-4, 2-5). The stratigraphic thickness of the Kaza Group in this area is not known, but it is interpreted to be between 2.87 km (middle Miette Group, Rocky Mountain Main Ranges, (Carey & Simony 1984, Carey 1984)) and 3.8 - 4.0 km (Kaza Group, southern Cariboo Mountains, (Pell & Simony 1984, Pell 1984)). The section of the Kaza Group described in this report is not a vertical stratigraphic section but an oblique section, compiled by mapping down structural plunge along a strike length of 65 km (Fig. 2-3, 2-4). Thus, the sedimentary facies at the base of the sequence near Tete Creek (Fig. 2-1, 2-3, 2-4) would be approximately 50 km southeast and 10 km structurally deeper than those at the top of the sequence near Castle Creek (Fig. 2-1, 2-3, 2-4).

Southwestwardly verging, regional-scale folds which duplicate stratigraphy across their gently dipping axial surfaces are responsible for the large structural thickness of the Kaza Group in the central

Fig. 2-3 Distribution of rock units in the east-central Cariboo Mountains and locations of areas A to F which were mapped in detail (in pocket).

Fig. 2-4 Longitudinal (NW-SE) cross section of east-central Cariboo Mountains, from ridge north of Castle Creek to ridge south of Tete Creek, illustrating how variation in northwestward plunge of syn-metamorphic (D_2) structures exposes deeper structural and stratigraphic levels in the southeastern part of the range. A to F are areas shown in Fig. 2-3. The line of section passes through areas A, B, and D; data from areas C, E, and F, are projected.

Cariboo Mountains (Fig. 2-5). Large-scale folds with gently dipping axial surfaces occur mainly in the lower part of the structural sequence, involving the lower and middle subdivisions of the Kaza Group; the upper part of the Kaza Group is characterized by relatively upright and open folds. The structural geometry and distribution of structural thickening of the Kaza Group requires that rocks at different structural levels be restored to locations at different distances to the southwest (Chapter 3). By treating the weakly deformed rocks at shallow structural levels at Castle Creek as a reference and approximating the amount of displacement associated with the structural thickening at different stratigraphic levels, rocks at intermediate structural levels presently located on the north side of Roush River (Fig. 2-1, 2-3, 2-4; Box D) would restore to a position less than 10 km to the southwest. Rocks at structurally deeper levels would restore even farther to the southwest, but the amount is less well constrained. Therefore, although sedimentary environments determined at the base of the sequence near Tete Creek are presently 50 km southeast of, and 10 km structurally deeper than, the environments determined at the top of the sequence near Castle Creek, they were originally relatively more to the south and west and greater than 50 km apart (Fig. 2-6).

The dominant folding at all structural levels occurred during prograde regional metamorphism, but prior to the metamorphic peak. The relatively simply deformed top of the Kaza Group is at muscovite/chlorite zone of regional metamorphism; the grade increases gradually down-section to kyanite/staurolite zone at the base of the sequence.

Fig. 2-5 Composite vertical cross section compiled from 30 serial cross sections (a-a' to f-f' are representative topographic profiles) across areas A to F shown in Fig. 2-3. Relative positions of cross sections are determined from lineation data (see Chapter 3). Form lines are bedding traces except where indicated. Form lines outside boxes A to F are drawn from photomosaics of cliff faces. Ornamentation and symbols as in Fig. 2-3.

In spite of the structural and metamorphic overprints, stratigraphic and sedimentologic data are available throughout the sequence, although they are significantly less common and more difficult to interpret at deeper structural levels.

As in the southern Cariboo Mountains (Pell 1984 and Pell & Simony 1984), the Kaza Group in the central Cariboo Mountains has been subdivided into upper, middle, and lower units. The base of the lower Kaza Group in this area is inferred to be a westwardly dipping low-angle fault which separates an upright sequence of meta-sandstone, calc-schist, pelitic schist, and thin marble of the lower Kaza Group from an inverted sequence of pelitic schist, amphibolite, meta-sandstone (or siltstone), and marble intruded by a pre-metamorphic and pre-kinematic alkalic orthogneiss (Fig. 2-3, 2-5). The fault has not been observed directly, but is likely to be exposed in the Shelter Creek drainage north of Tete Creek and on Mica Mountain (Fig. 2-1, 2-4). The footwall sequence is tentatively correlated with the SPA ("semi-pelite" and amphibolite) and middle marble units of the Horsethief Creek Group (Murphy 1984). The top of the upper Kaza Group is transitional with the low-grade black slate and limestone of the overlying Isaac Formation of the Cariboo Group (Campbell et al 1973, Murphy & Rees 1983).

(2-5b.) Depositional model for the interpretation of the sedimentary facies

Most of the Kaza Group consists of monotonously interbedded meta-sandstone and meta-shale. The coarse clastics are interpreted as turbidites based on common graded bedding, sharp bases and tops transitional to pelitic beds, presence of variably complete Bouma

sequences, and the general lack of traction current structures. In this paper, the frequency of occurrence of certain Bouma intervals (here described using turbidite "shorthand" (Ta, Tbc, etc.), which indicates which Bouma beds are present in a single turbidite) and various other parameters (sandstone/shale ratio, bed shape, and bed thickness and grain-size trends) are used to interpret the local depositional environments of the Kaza Group (Walker 1967, 1970, 1984; Walker & Mutti 1973; Mutti & Ricchi-Lucchi 1978; Normark 1978). Fan terminology (inner, mid-, outer, suprafan lobe) is from Mutti & Ricchi-Lucchi (1978) and Normark (1978).

Although a turbidite fan model is proposed for the interpretation of the local environments of deposition, a fan configuration cannot be proven on the basis of data from the Cariboo Mountains alone. A more regionally based study is necessary to document the larger-scale configuration of Kaza/Miette Group turbidites.

(2-6) STRATIGRAPHY AND DEPOSITIONAL ENVIRONMENTS OF THE KAZA GROUP

(2-6a.) Lower Kaza Group

(2-6a.-1) Stratigraphy

The lower Kaza Group has not been examined in detail. It is best exposed in the Tete Creek Valley west of Valemount (Fig. 2-1); in this area, it consists of meta-sandstone, calcareous and pelitic schist, and thin, persistent, orange to buff marble. Overlying the lower Kaza Group are the thin- to thick-bedded, ~~coarse~~, locally pebbly meta-sandstone and phyllite/schist of the middle Kaza Group. The contact is sharp; the massive middle Kaza "grits" contrast strongly with the more

schistose character of the lower Kaza Group (Plate 2-1a).

Thin- to medium-bedded (up to 2.0 m-thick), fine-grained to pebbly meta-sandstone predominates, but the proportion of interbedded schist and marble is sufficiently high to give the unit a schistose appearance. Meta-sandstone occurs in composite, bedded units (up to 5 m-thick) with some interbeds of schist that probably represent "amalgamated" beds (Mutti & Ricchi-Lucchi 1978). Bedding is generally even and parallel with rare small channels. Graded bedding is common, and rare thin cross-laminated intervals occur at the top of the graded beds (Tace).

Schist and marble compose approximately 50% of the lower Kaza Group in the Tete Creek valley and increase towards the top. Schist is commonly slightly calcareous, with porphyroblasts of plagioclase and orthozone. Calcite-orthozone-plagioclase-garnet-biotite-hornblende calc-schists also occur, commonly in association with the thin (up to 1.5 m-thick), persistent calcitic marbles (Plate 2-1b). Schists are finely laminated, reflecting variations in graphite, calcite, or clastic quartz content. Marble is coarse-grained, generally featureless, and is restricted to the thicker schist beds.

(2-6a.-2) Depositional Environments of the Lower Kaza Group

The relatively thin-bedded and tabular shape of lower Kaza turbidites and the high percentage of schist and carbonate suggest a distal, non-channeled fan setting. Composite sandstone beds possibly represent superposed sheets of sand emitted from channels high on the fan and deposited in suprafan lobes (Normark 1978). Schist, calc-schist, and possibly marble, represent the basinal pelagic or

PLATE 2-1

- a. Contact between lower and middle Kaza Group as exposed south of Tete Creek (less prominent black dotted and dashed line). Prominent white line on right is axial surface trace of D₁ synformal syncline. White line on left is axial surface trace of D₂ antiformal anticline. Note recessive nature of calc-schists and ribbed appearance of thin marble bands (lower right).
- b. Recessive calc-schist and marble bands in lower Kaza Group north of Tete Creek. Figure in lower right for scale.
- c. Channel incised into thick-bedded coarse-grained sandstone and pebbly sandstone near the base of the middle Kaza Group.
- d. Channel incised into coarse grained sandstone near the base of the middle Kaza Group. Channel fill is rich in pelitic intraclasts which are concentrated in the upper part of the bed. Arrow indicates younging direction. Lens cap for scale.
- e. Amalgamated Tabce beds (3)' north of Raush River.
- f. Asymmetric (east-verging) fold outlined by thin marble band (tip of arrow). View is to the north; the ridge in the background is the north side of S. Kiwa Creek.

PLATE 2-1

hemipelagic background sediments overridden by the turbidites. A quiet, sub-wave base setting is inferred from the fine grain size, lack of current structures, and thinly laminated character of the schistose rocks; water depth and salinity are unknown.

(2-6b.) Middle Kaza Group

(2-6b.-1) Stratigraphy

The middle Kaza Group is a sequence of meta-sandstone, pebbly sandstone, calcareous and pelitic meta-shale, poorly sorted, and commonly intraclast-rich conglomerate, and rare thin beds of marble. The large structural thickness of Kaza Group rocks between S. Kiwa Creek and Castle Creek (Fig. 2-1, 2-3) is almost entirely composed of middle Kaza Group strata. Four subdivisions of the middle Kaza Group are recognized in this area, based on the proportions of the aforementioned rock-types. It is not known if these subdivisions persist regionally, although the middle portion of the Kaza Group described by Pell & Simony (1984) and Pell (1984) is remarkably similar.

The lowest unit is characterized by a high sandstone to shale ratio and the presence of thin limestone/marble beds within the schist units. Well graded, coarse-grained sandstone and conglomerate occur in thick-bedded (1 - 2 m-thick), amalgamated sequences with little interbedded phyllite or schist (TAamal). Pebbly sandstone or conglomerate commonly infills channels (up to 2 m deep; Plate 2-1c) eroded into underlying sandstones. Intraformational meta-shale clasts are common, concentrated at the tops of grit beds (Plate 2-1d). Both sharp and transitional contacts occur between sandstones and overlying thin beds

of phyllite or schist. Basal contacts of sandstone beds are sharp.

Laterally and vertically transitional with the coarse-grained, thick-bedded sandstone are sequences of thin-bedded (0.5 - 1.5 m-thick) and fine-grained sandstones and interbedded phyllite/schist. The sandstone to shale ratio of these sequences is lower than in the amalgamated coarse-grained sandstone sequences. Channels are less common in the thin-bedded facies, although graded beds with sharp bases are present. Cross-lamination at the tops of beds is common in this facies (Tace).

Coarse-grained, thick-bedded, amalgamated sandstones and thinner, finer-grained sandstones form laterally persistent bundles that are stratigraphically separated by units of phyllite/schist commonly containing thin marbles. Bedding thickness trends are locally present but it is not possible to confidently establish whether thinning-upward ("positive sequences", Mutti & Ricchi-Lucchi 1978) or thickening-upward ("negative sequences") predominate.

Units of schist/phyllite and thin calcitic marble range up to 15 m in structural thickness. They are laterally persistent and useful in outlining structural geometries (Plate 2-1f). Calc-schist (meta-marl) is associated with marble and is distinguished from pelitic schist by porphyroblasts of orthozoisite and plagioclase. No sedimentary structures were observed in the carbonate beds.

The lower middle unit of the middle Kaza Group is distinguished from the basal unit by a lower sandstone to shale ratio, finer grain size, and general scarcity of limestones in the schist units. Multiple fine-grained, generally featureless, 0.3 to 1 m-thick sandstones form composite beds which generally fine and thin upward and

which may be traced laterally for several kilometres. Although these units appear ~~amalgamated~~, no channels are observed and pebble- and intraclast-rich horizons are rare. Pebble-bearing beds are commonly graded and capped by thin, parallel- and cross-laminated beds (Tabce). In one location, several amalgamated Tabc(e) beds form part of a larger composite unit (Plate 2-1f). Pebble-bearing beds are more common up-section, but in general are relatively rare.

Interbedded phyllite beds are approximately the same thickness of the composite sandstone beds (5 - 20 m-thick); the overall sandstone to shale ratio approximates 1. These units contain a small percentage of thin (cm - 0.5 m-thick), fine-grained, featureless sandstone beds. Some of the thinner beds are graded, with sharp bases and tops transitional into the overlying schist. Rare carbonate-bearing beds were observed in the lower part of this unit, associated with graphitic schist.

The upper middle part of the middle Kaza Group is distinguished from the underlying unit by a significantly higher sandstone to shale ratio and features many of the characteristics of the basal portion of the middle Kaza Group. It comprises graded, coarse-grained, thick-bedded; and amalgamated sandstone (TaAmal); pebbly sandstone; thinner single-bed sandstone layers with variably complete Bouma sequences (Tabc(d)(e)); single-bed massive and graded coarse sandstone layers (Tae); and phyllite with thin (0.5 - 2 cm-thick) graded siltstone or fine-grained sandstone laminae.

Sandstones are arranged in positive, negative, and non-cyclic sequences (Fig. 2-7, in pocket). Typically, a relatively thin negative

sequence will be overlain by a thicker positive sequence. The basal parts of the positive sequences consist of amalgamated thick, graded (1 - 3 m-thick), coarse-grained, commonly pebbly and intraclast-rich sandstones. Thick beds often have erosional bases, but even with the coarsest beds, the amount of down-cutting into underlying beds is small. The base of thick-bedded sandstones commonly exhibits ball and pillow structure (Plate 2-2a).

The upper parts of the positive sequences consist of stacked individual graded turbidites (Tae, Fig. 2-7). The tops of the sequences are relatively thin-bedded, finer-grained Tabce, Tabc amalgamated, or Tabe turbidites. Interbedded phyllite ranges up to 25 m in thickness and is ubiquitously laminated with very thin (mm- to cm-scale) siltstone laminae. The top of this unit is a 25 m-thick laminated phyllite with thick beds of massive, well sorted, and occasionally graded, siltstone.

The uppermost unit of the middle Kaza Group is a distinctive, laterally persistent 200 m-thick sequence of siltstone and fine sandstone, graphitic and non-graphitic silty phyllite, and thin limestone/marble. The unit has been mapped southward from Castle Creek for approximately 15 km and has been observed to continue farther to the southwest. This unit or its facies equivalent is apparently present in the stratigraphic sections of the "grit" unit in both the Rocky Mountains and the southern Cariboo Mountains. Because of its distinctive lithofacies and importance as a marker horizon, this siltstone unit was examined in detail in a well exposed area south of Castle Creek.

The 25 m-thick laminated phyllite at the top of the underlying unit

PLATE 2-2

- a. Ball-and-pillow structure at base of massive, coarse-grained sandstone below marker unit at top of middle Kaza Group. Linear traces cutting diagonally from upper left to lower right are traces of D_2 crenulation cleavage. Photograph courtesy of G. M. Ross.
- b. Thin to medium-bedded plane-laminated siltstone unit at base of marker unit. Recessive beds are contorted and cross-laminated fine-grained calcareous sandstones and siltstones (Plate 2-2c, d). Hammer at lower center for scale. Photograph courtesy of G. M. Ross.
- c. Close-up of contorted and cross-laminated beds in the lower part of the marker unit at the top of the middle Kaza Group. Subvertical linear traces are traces of D_2 cleavage. Photo courtesy of G. M. Ross.
- d. Polished slab exposing detail of plane laminated and contorted beds in lower siltstone part of the marker unit at the top of the middle Kaza Group. Direction of overturning of contortions is southwest. Arrow in upper right indicates lens of cross-laminated quartz sandstone that is interpreted as a starved ripple. Dark clots near the contact between the contorted and plane laminated beds are calcite, chlorite, and pyrite pseudomorphs after pyrite (cubic outline preserved). Dark speckles in plane-laminated siltstone are chlorite porphyroblasts. Arrow in upper left is approximately 1.5 cm-long.
- e. Truncation surface (arrows) separating plane-laminated siltstone with laminations parallel to the surface from plane laminated siltstone with laminations inclined to surface. It is not clear if inclination of laminations is due to bedform migration or erosional incision. Inclined layer thins to the right. Photograph courtesy of G. M. Ross.
- f. Thin limestone beds near the top of the siltstone part of the marker unit at the top of the middle Kaza Group. Note lenticular nature of beds at top of photograph and lamination and cross-lamination in continuous beds in lower half of photograph.

PLATE 2-2

is overlain by a 150 m-thick unit consisting primarily of plane-laminated, green, chloritic and pyritic, fine-grained sandstone and coarse siltstone (Plate 2-2b). Plane-lamination is defined by alternating mm-scale laminae of chlorite-rich and chlorite-poor, quartz and calcite silt or fine sand (Plate 2-2d). Some laminae appear graded, with sharp bases; others have transitional tops and bottoms.

Laterally persistent beds of fine-grained, white, rusty weathering, contorted and cross-laminated, calcareous quartz-sandstone and siltstone, 5 - 15 cm-thick, with sharp, locally scoured bases, interrupt the plane laminated layers at 20 - 75 cm intervals (Plate 2-2e). Although not rigorously examined, the vergence of contorted cross-laminated horizons appears to be towards the southwest. Thin cm-scale lenses of cross-laminated fine grained white quartz sandstone resembling starved ripples occur locally between plane-laminated beds (Plate 2-2f). Low-angle internal truncation surfaces (Plate 2-3a) and wedge-outs of beds occur in the lower part of the siltstone unit.

Near the top of the siltstone unit, 5 - 10 cm-thick buff to tan limestone/marble beds, both continuous and lenticular, and 5 - 10 cm-thick, finely laminated graphitic and non-graphitic phyllites occur at 20 - 50 cm intervals (Plate 2-3b). One horizon of climbing ripples displays a scoured base near the contact of the siltstone unit with overlying phyllite (Plate 2-3c).

The carbonate-bearing upper part of the cliff-forming meta-siltstone unit is overlain by a 11 m-thick recessive phyllite and marble unit (Fig. 2-7). Laminated silty grey and black graphitic phyllites occur in the lower part of this unit and are interbedded with tabular to lenticular marble. The laminated phyllite passes up-

PLATE 2-3

- a. Climbing ripples and plane-lamination in upper siltstone part of the marker unit at the top of the middle Kaza Group. Linear traces cutting diagonally across photograph from upper left to lower right are traces of D₄ crenulation cleavage. View is approximately to the southwest. Photograph courtesy of G. M. Ross.
- b. Intraclast-bearing calcareous quartz sandstone at top of black phyllite part of the marker unit at the top of the middle Kaza Group. Coin in upper middle of photograph. Photography courtesy of G. M. Ross.
- c. Large channel incised into medium- to thick-bedded coarse-grained sandstone and pebbly sandstone near base of upper Kaza Group. Channel fill is primarily coarse-grained, thick-bedded sandstone and pebbly sandstone. Quartz vein with altered contacts in center of photograph.
- d. Trough cross-bedded coarse-grained sandstone near base of upper Kaza Group. Hammer for scale.
- e. Soft sediment folds (modified by D₂ deformation) in coarse-grained sandstone near base of upper Kaza Group. Layering is defined by carbonate content: dark layers are more carbonate rich. Vergence of folds is to the southwest. End of hammer handle visible at lower right.
- f. Cut-and-fill trough cross-beds in upper part of the upper Kaza Group. Contact of upper Kaza Group with overlying Isaac Formation is a few hundred metres to the east of the location of this photograph. Dark, subvertical linear traces are traces of D₂ crenulation cleavage. Overlying bed is approximately 75 cm-thick.

PLATE 2-3

section into approximately 7.5 m of black, graphitic phyllite with thin, black, massive and cross-laminated marble. The graphitic section is capped by a 75 cm-thick white intraclast-bearing calcareous quartz-sandstone with contorted cross-lamination. A unit of 7.5 m of grey silty phyllite with thin (5 + 10 cm-thick) siltstone and marble beds intervenes between the quartz sandstone and the coarse turbidites of the upper Kaza Group. Silty phyllites directly beneath the first turbidites are graded, signaling the return to turbiditic sedimentation.

(2-6b.-2) Depositional Environments of the Middle Kaza Group

Middle Kaza Group sandstones at all levels in the unit have several features in common: (1) organization into laterally persistent composite beds with high sandstone to shale ratios, (2) general parallel and even-beddedness, (3) rare development of channels with significant incision, (4) predominance of fining- and thinning-upward cycles, and (5) correlation of completeness of Bouma sequences with bed thickness and lower sandstone to shale ratios. These characteristics suggest sandstone lobe deposition on a smooth or weakly incised, outer mid-fan apron. Positive and negative cycles are attributed to waning and waxing of sediment supply emitted from nearby channels (cf. Mutti & Ricchi-Lucchi 1978; Walker 1984).

The basal and upper middle units of the middle Kaza Group both have high sandstone to shale ratios and common channels and thick-bedded turbidites. These two units may alternatively be interpreted as incursions of inner/upper mid-fan facies rather than suprafan lobe facies, as was suggested by Murphy and Journeay (1982). However, the

low relief relative to the apparent width of channels, the incomplete development of coarsening- and thickening-upward cycles, and the high sandstone to shale ratios do not support this interpretation. This channelized facies more likely represents deposition in the axial zones of the lobes that lie outboard of inner mid-fan channel mouths, a setting comparable to that described by Mutti & Ricchi-Lucchi (1978) and Walker (1984).

The lower middle unit of the middle Kaza Group contains a large percentage of schist, including calc-schist and marble, and thin-bedded (mm-to cm-scale) fine-grained sandstone and siltstone. These rocks are inferred to represent basinal fine-grained clastic and chemogenic(?) background sediments and "distal" turbidites, respectively. "Distal" in this sense may mean either far from channels, but topographically on the outer mid-fan apron (overbanks?), or true outer fan distal turbidites. Because of dilution by clastic sedimentation, chemogenic sediments would probably not be a significant component of the mid-fan; their presence would suggest that these turbidites were deposited on the outer fan. With the present data base, these possibilities cannot be unequivocally discriminated. Interbedded basinal units show an upward decrease in carbonate content, suggesting a change in the factors controlling carbonate production (such as turbidity, P_{CO_2} , salinity).

The upper unit of the middle Kaza Group differs from the remainder of the Kaza Group in several respects: (1) it is characterized by a suite of sedimentary structures which is distinctive from sub- and suprajacent turbidites, (2) finer grain size and better sorting, (3) quartz-rich clast composition, (4) presence of clastic calcite grains,

and (5) abundance of limestone.

The interpretation of the environment of deposition of the siltstone/fine-grained sandstone lithofacies of the marker unit hinges on the hydrodynamic interpretation of plane-lamination and the relationship of the plane-laminated beds to the contorted and cross-laminated beds. Plane lamination may be produced under suspension-settling, upper, or lower flow-regime conditions and distinguishing between them can be difficult (Harms et al. 1982). Upper flow-regime conditions seem unlikely in this case because of the lateral persistence of individual laminations and the lack of evidence of waxing or waning flow as might be expected given the 150 m thickness and large lateral extent of the unit. Suspension-settling or lower flow-regime conditions would be supported by the general lack of clearly defined tops or bottoms of laminations, in addition to the lateral persistence of laminations and the lack of evidence for waxing or waning flow.

The origin of the cross-laminated and contorted calcareous quartz sandstones and limestones is not clear. Cross-laminations and contorted cross-laminations characterize the Bouma "C" interval and possibly the plane-laminated and contorted cross-laminated layers are "B" and "C" Bouma beds, respectively, and the two together, Tbc(ê) turbidites. However, the scoured base and larger grain size of the contorted and cross-laminated beds suggest that they were not deposited continuously with the plane-laminated beds during waning-flow sedimentation, so that each bed represents a separate depositional event.

The contorted and cross-laminated beds could be interpreted as sheet-like Tc(e) turbidites that spread outward over the slightly finer-grained siltstones. However, the apparent lack of Ta and Tb beds and the lateral persistence of the current structures argues against this interpretation.

An alternative interpretation, one that seems best able to account for the observations, is that the cross-laminated and contorted cross-laminated rocks formed from intermittent traction currents. These currents would have to have been sufficiently variable in strength and competence to be able to form current structures ranging from starved to climbing ripples. They also would have to have been locally strong enough to scour the substrata and form truncation surfaces.

An environment which is consistent with the traction current interpretation of these beds is a shallow, but sub-wave base, clastic shelf across which well-sorted silicic and calcareous clastics are transported by intermittent wave, storm or density currents. The presence of clastic calcite implies transport through a carbonate-producing environment. A slight depositional slope or tilting immediately after deposition can be inferred from the consistent sense of contortion of contorted cross-laminated sandstone. The quartz-rich and well sorted nature of these rocks contrasts sharply with the poorly sorted supra- and subjacent turbidites, suggesting a higher degree of reworking.

The upper phyllite, marble, calcareous and intraclast-bearing quartz sandstone portion of the marker unit is similar in many ways to the hemipelagic "background" into which Kaza Group turbidites were deposited. Limestones/ marbles in the marker unit are commonly cross-

laminated, indicating re-sedimentation from a carbonate-producing, possibly shelf, environment. Limestone/marble elsewhere in the Kaza Group is massive, rarely exhibiting sedimentary structures. The relatively mature composition of the calcareous quartz sandstone with deformed cross-beds and calcareous intraclasts suggests that this unit is also re-sedimented, possibly from a shelf source. The re-sedimented nature of these rocks suggests a slope environment, transitional between the shelf facies inferred for the underlying siltstones and overlying turbidites of the upper Kaza Group.

(2-6c.) Upper Kaza Group

(2-6c.-1) Stratigraphy

The upper Kaza Group comprises approximately 1.2 km of coarse-grained turbiditic clastics with a high, but upwardly decreasing, sandstone to shale ratio. Its contact with the underlying "marker" unit is sharp; coarse clastics in the basal 5 m are intraclast-rich, possibly due to down-cutting into the underlying unit (Fig. 2-7). Its upper contact with the Isaac Formation of the Cariboo Group is transitional, occurring across a 250 m-thick interval (Campbell et al. 1973, Murphy & Rees 1983).

A 3 m-thick unit of grey silty phyllite with thin (5 - 10 cm-thick), graded, calcareous siltstone laminae at the top of the middle Kaza Group signals the return to turbiditic sedimentation (Fig. 2-7). A sequence of non-cyclic medium- to thick-bedded (0.5 - 1.5 m-thick) Tae, Tabe, Tac, and TaAma turbidites directly overlies the basal phyllite. Fine-grained interbeds are calcareous siltstone, silty phyllite with thin (mm-scale), graded (?) laminae, and thin, graphitic,

and calcareous phyllite with 5 - 10 cm-thick limestone/marble beds. Incised into these units are channels with relief up to 4 m, infilled with coarse-grained, medium- to thick-bedded sandstone (Plate 2-3c). A unique occurrence in the Kaza Group of this area are metre-scale trough cross-beds and southwestwardly verging slump folds (Plates 2-3d and 2-3e). Although initially interpreted as tectonic features (Murphy & Rees 1983), these folds are confined to a single horizon, have a fluidal appearance indicative of a very soft rheology, and pre-date, and are inclined to, all of the penetrative tectonic fabrics observed in this area, suggesting a penecontemporaneous origin.

The middle division of the upper Kaza has not been examined in detail, but is composed primarily of thick-bedded turbidites with a high sandstone to shale ratio, apparent thickening- and coarsening-upward trends, and local small-scale channels filled with trough cross-bedded sandstone (Plate 2-3e). Bedding thickness, sandstone to shale ratio, and presence of channels decrease upward and thickness of pelite and carbonate beds increases upward into the overlying Isaac Formation. Carbonate clasts in turbidites increase towards the top of the Kaza Group. The top of the Kaza Group has been placed at the base of a thick grey phyllite with thin, cross-laminated, sandy limestone/marble. A thin, coarsening- and thickening-upward grit unit is locally present above the phyllite, but is included in the Isaac Formation.

(2-6c, 2) Depositional Environments of the Upper Kaza Group

The lower part of the upper Kaza Group is interpreted to have formed in an upper, channelled mid-fan environment. An irregular, incised fan topography is implied by deep (up to 4 m) channels filled

49

with thick-bedded turbidites. Slump structures may reflect either the effects of local topographic relief, or a regional paleoslope. The vergence of slump folds is primarily to the southwest, similar to the contorted cross-beds in the underlying marker unit. Metre-scale trough cross-beds are interpreted as bed forms associated with channel mouth bars, following Mutti (1977).

The lateral persistence of beds, lack of large channels, and absence of slump structures in the middle part of the upper Kaza Group suggests a return to a non-channelled outer mid-fan apron environment. Channels and cut-and-fill cross beds indicate less than 1 m of local topographic relief. Thinning- and fining-upward trends are attributed to channel migration or avulsion higher on the fan (Mutti & Ricchi-Lucchi 1978, Normark 1978, Walker 1984).

The transition into the Isaac Formation is interpreted to be the result of outward progradation of slope and platform environments over the Kaza depocenter (Murphy & Rees 1983). Turbidites are gradually succeeded by finely laminated phyllite and thin, black, laminated and cross-laminated marble/limestone. Locally present at the base of the Isaac Formation are coarse limestone/marble conglomerates with clasts of limestone/marble, phyllite, calc-phyllite, blue quartz, and oolitic (?) limestone/marble, which are interpreted as slope deposits. One possibly lenticular, thickening-upward sequence of turbidites is present within the shales and limestones at the base of the Isaac Formation, possibly representing a slope channel facies.

(2-7) REGIONAL CORRELATION OF THE KAZA GROUP

(2-7a.) Distribution of sedimentary environments in the Kaza Group of the central Cariboo Mountains

The present distribution of the sedimentary environments of the Kaza Group is quite different from the distribution prior to the mid-Jurassic structural thickening. Fig. 2-8 shows the variation in depositional environments at the level of observation along the line of section in Fig. 2-4. As can be seen from the diagram, the control on the vertical variations in sedimentary environments is limited at any one locality. It is clear that more control is necessary before any conclusions can be drawn about the three-dimensional geometry of the turbidite facies.

At the level of observation along the line of section, the depositional environments are remarkably constant. The only clear departure from mid-fan sedimentation is the marker unit at the top of the middle Kaza Group. Although the environmental interpretation of this unit is not straightforward, the two environments under consideration (shallow shelf and distal turbidite) both imply perturbations of the depositional system that may be regional in extent. If region-wide, a record of these perturbations should exist in stratigraphic sections of the Kaza and Middle Miette Groups from other areas, thus providing a firm basis for regional correlation.

(2-7b.) Kaza Group of the Southern Cariboo Mountains

In the southern Cariboo Mountains, Pell (1984) subdivided the Kaza Group into upper, middle, and lower units which are similar to those described in this report. The top of the middle Kaza Group in this

Fig. 2-8 Variation of depositional environments along line of section shown in Fig. 2-4.

area is marked by a distinctive 250 m-thick, laterally persistent member comprising "pink quartzite, psammite, and pelite interbedded with green laminated pelite" at its base, "orange weathering cross-laminated calcareous psammite and green, parallel and cross-laminated silty pelite", and an overlying "rusty zone" of "interbedded calcareous psammite, rusty laminated pelites, buff sandy marble, white quartzite, black graphitic and sulfide rich schist, and buff calcareous grit".

Although the depositional environment of these rocks is not discussed, the presence of quartzite (presumably a metamorphosed quartz sandstone) is distinctive amidst the texturally and compositionally immature turbiditic feldspathic sandstone, pebbly sandstone, and conglomerate of the underlying and overlying Kaza Group. This difference can be attributed to a higher degree of reworking, as has been inferred for the quartz-rich siltstone and sandstone of the marker unit of the central Cariboo Mountains. Lateral continuity of these two units has yet to be established, but on the basis of lithostratigraphic similarity, lateral persistence, stratigraphic position, and distinctiveness of lithofacies from underlying and overlying rocks, a correlation is proposed (Fig. 2-9).

(2-7c.) Middle Miette Group of the Rocky Mountains

The middle Miette Group in the Cushing Creek area, Rocky Mountain Main Ranges consists primarily of coarse-grained turbiditic clastic rocks interpreted to have been deposited in a mid- and upper fan setting (Carey 1984). A distinctive 250 m-thick sequence of interbedded parallel laminated silty slate and gritty sandstone (base), and slate with olistostromal stromatalitic (?) limestone, cross-

Fig. 2-9 Correlation of sections of the Kaza and middle Miette Groups, Cariboo and Rocky Mountains.

laminated and micritic limestone, and fine grained calcareous sandstone (top) interrupts the upper part of the middle Miette Group. The limestone-bearing part of this unit appears on both map (Carey 1984) and cross-section (Carey & Simony 1984) as a laterally persistent, mappable unit. Although the depositional environment of this unit was not discussed, the finer grain size of the sandstones and the presence of relatively thick (individual beds up to 41 m, Carey (1984)) limestones suggest a shallower water depositional environment than that into which the underlying and overlying turbidites were deposited. For the same reasons as cited above, a correlation with the marker unit of the Cariboo Mountains is proposed (Fig. 2-9).

Carey (1984) proposed a correlation between the southern Cariboo Mountains and the Rocky Mountain Main Ranges that is similar to the correlation proposed in this study, but differs in detail. This correlation equates a distinctive 80 m-thick unit of bright green, ripple cross-laminated to parallel-laminated, silty slate 1360 m above the base of the middle Miette Group (or 600 m below the limestone - slate unit correlated in this study) with the pink and green quartzite and pelite marker of the southern Cariboo Mountains. Although not described in detail, the silty slate beds appear similar to the lower siltstone part of the marker described in this study; however, they are intimately interbedded with gritty sandstone and conglomerate and as such do not represent the dramatic break in turbiditic sedimentation such as recorded by the marker in the east-central Cariboo Mountains. For this reason, the correlation of the apparently regressive limestone and slate unit with the marker unit of the Cariboo Mountains is preferred.

(2-8) PALEOGEOGRAPHIC INTERPRETATION

Invoking Walther's Law, the continuous vertical sequence of turbidites, shelf facies clastic rocks, and slope facies shales, limestones and re-sedimented rocks implies lateral co-existence and continuity of these environments during the time of their deposition. Shelf and slope deposits are rare in the Kaza and middle Miette Groups; presumably they were more prominent to the east (up the paleoslope, Aitken & Mountjoy 1963, Carey 1984), but were either eroded in the late Proterozoic/early Paleozoic (Aitken 1969) or are still undetected in structurally lower thrust sheets not exposed at the surface. As such, the rare shelf and slope environments interfingering with the turbidites provide the only clues to the paleogeography and paleogeographic evolution of the Windermere basin in this part of the Cordillera.

Fig. 2-10 is a preliminary paleogeographic model illustrating the sedimentary environments during deposition of the Kaza Group. At time 0, the Windermere "sea" was flanked by marginal, glaciated (Eisbacher 1981) highlands which shed coarse clastic detritus directly to the shoreline. A shelf of unknown width was present offshore, incised by channels which led to the head of a slope which bounded the shelf on the west. Shelf channels fed sediment directly to the heads of slope canyons which opened into base-of-slope submarine fan complexes. The present data do not permit further characterization of this slope, but a preliminary conclusion is that rather than being a fully developed continental slope with several kilometres of relief, it was more like a pro-delta slope.

Fig. 2-10 Preliminary reconstruction of depositional environments present during the deposition of the marker unit at the top of the middle Kaza Group. See text for discussion.

At high sea level stands (time 0), poorly sorted, coarse-grained detritus was either conducted directly from its source to the turbidite depocenter with little exposure to fluvial or clastic shelf processes or reworked on the clastic shelf. With regression (time 1), the shoreline prograded outward away from the marginal highlands, and sediment underwent a greater amount of fluvial reworking before reaching the shelf. Sediment reaching the shelf would have been finer-grained, better sorted, and more quartz-rich than during high sea level stands. Traction currents on the shelf could have distributed this sediment across the shelf and onto the former site of turbidite sedimentation.

The regressive marker overlies inner fan turbidites in the Cushing Creek area (Carey 1984) and finer-grained mid-fan turbidites in the central Cariboo Mountains. The lack of a transitional slope facies in either locality suggests that the regression may have occurred rapidly, and that the slope intervening between the shelf and turbidite environments may not have been especially prominent. For shelf environments to prograde outward over the former site of turbidite deposition, the vertical distance between the two environments can not be greater than the amount of relative sea level change during regression. For example, if the regression were analogous to sea level changes during Pleistocene glaciation, the turbidite fans would have to have been less than 100 m deeper than shelf environments for the shelf environments to be able to prograde over the site of turbidite deposition.

With subsequent transgression, coarse-grained clastic sediments reaching the shelf would have been ponded high on the shelf, permitting

finer-grained clastic and carbonate sedimentation on the shelf and re-sedimentation on the slope. Sedimentation of coarse-grained turbidites would have been restored when the transgressing shoreline was re-established near the highlands flanking the basin. Possibly, river channels that conducted sediment to the shoreline during the regression provided conduits for turbidity flows to bypass the shelf and be deposited directly without significant reworking during the transgression.

With the exception of local turbidites at the base of the Isaac Formation, the transition from the upper Kaza Group and middle Miette Groups to the Isaac Formation and upper Miette Group marks the end of coarse-grained turbidite sedimentation. In the Cariboo Mountains, Kaza Group turbidites are overlain by a slope facies which strongly resembles the upper part of the marker unit. The remainder of the Isaac Formation in the Cariboo Mountains consists of black graphitic phyllite and thin limestone (Campbell et al. 1973). In the Rocky Mountains, the middle Miette Group turbidites are overlain by a sequence of green, plane-laminated slate and siltstone resembling the lower part of the marker unit (A. Carey, personal communication 1984). These data suggest that the more proximal shelf and slope facies did not prograde outward over the former location of turbidite sedimentation as far as during the sedimentation of the marker unit. A transition between a siliciclastic-dominated shelf and a shale/carbonate-dominated shelf within the Isaac/upper Miette Group is indicated.

(2-9) DISCUSSION

Although the amount of data on depositional environments of the Kaza and Middle Miette Groups is sparse, some preliminary conclusions about the evolution of the Kaza/Miette basin are possible. First of all, the lack of major variations in facies in either the east-central Cariboo Mountains or the Cushing Creek area of the Rocky Mountains (Fig. 2-1) suggests that subsidence and sedimentation were closely balanced. In the Cushing Creek area, inner fan and inner mid-fan turbidites occur in abundance throughout a thickness of almost 3 km. In the east-central Cariboo Mountains, outer mid-fan turbidites predominate throughout an unknown, but probably slightly larger original thickness. The lack of interfingering between these facies suggests that a steady-state topographic (shelf - slope - basin) configuration was sustained during the deposition of these units. This equilibrium was temporarily broken during the deposition of the marker unit and permanently broken at the onset of deposition of the Cariboo Group when shelf, slope, and terrigenous facies (Young 1979) prograded outward over the former locus of turbidite deposition.

Secondly, the eastward thinning of the "grit" unit (Fig. 2-9) suggests the presence, east of the Cushing Creek area, of a hinge line which separates more rapidly subsiding crust from less rapidly subsiding crust. Because the subsidence mechanisms of the Windermere basin are as yet unknown, the geodynamic significance of this hinge line is somewhat speculative. Perhaps the subsidence of the Windermere basin is a consequence of thermal decay of attenuated continental crust, in which case the difference in thickness across the hinge line

could be a reflection of variations in thickness of the underlying crust, i.e. the hinge line may represent the transition from cratonic to attenuated continental or transitional crust.

In the region of cross section A - A' (Fig. 2-2), the geophysically determined transition from cratonic to transitional or attenuated crust underlies the southern Rocky Mountain Trench (Monger & Price 1979, Price 1981, Price et al. 1985, Chamberlain & Lambert 1985). middle Miette Group rocks of the Cushing Creek area, which are inferred to have been deposited west of a Windermere-age basin hinge line, are currently located east of the edge of cratonic-thickness crust at the southern Rocky Mountain Trench. However, these same middle Miette Group rocks are displaced to a point approximately 100 km west of the basement beneath the southern Rocky Mountain Trench when the shortening in the Rocky Mountain Fold and Thrust Belt is restored. Therefore, it is possible that the craton edge beneath the southern Rocky Mountain Trench is the lower crustal expression of a Windermere-age basin hinge line.

Price et al. (1985) use similar reasoning to infer that the geophysically determined cratonic edge in the Kootenay arc region of southeastern British Columbia is a Paleozoic feature. With restoration of the shortening in the Rocky Mountain Fold and Thrust Belt, the basin hinge line of lower Paleozoic rocks of the Rocky Mountains displaces westward to a point in the Kootenay arc region directly above the geophysical anomalies inferred to represent the craton edge. In the region of cross section A - A' (Fig. 2-2), however, restoration of the shortening in the Rocky Mountain Fold and Thrust Belt displaces the lower Paleozoic hinge line to a point

approximately 100 km west of the geophysically determined craton edge; the craton edge in this area, therefore, is more likely to be a Windermere-age feature.

(2-10) CONCLUSIONS

- (1) The Kaza Group of the central Cariboo Mountains has been subdivided into three lithostratigraphic units. One distinctive 200 m-thick unit at the top of the middle Kaza Group appears to occur regionally and forms the basis of regional correlations.
- (2) The 200 m-thick marker at the top of the middle Kaza Group is inferred to have been deposited in a sub-wave base shelf environment. Its persistence is attributed to regional regression during which shelf facies prograded outward over the locus of turbidite sedimentation. Interbedded shelf, slope, and turbidite facies imply the presence of a shelf - slope - basin configuration during deposition of the "grit" unit.
- (3) The eastward thinning of the "grit" unit suggests the presence of a basin hinge line east of the location of the Cushing Creek area. The difference in subsidence rate across the hinge line may be due to a difference in thickness of the underlying crustal basement caused by attenuation of cratonic thickness basement during Windermere sedimentation.
- (4) If the Windermere hinge line was to the east of the lower Paleozoic hinge line, this could account for the discrepancy between

the palinspastic location of the lower Paleozoic hinge line and the geophysically determined edge of cratonic thickness basement at the Rocky Mountain Trench.

Acknowledgements I would like to thank G. Ross of Washington University (St. Louis, Missouri), J.M. Journeay of Queen's University, and C.J. Rees of Carleton for their contributions and companionship in the field. G. Ross is to be further acknowledged for a review of an early draft of this paper, sharing his ideas on the sedimentological interpretation (which differ somewhat from what is concluded in this paper) of the structures in the marker unit, and taking superb photographs. I would also like to thank A. Carey and J. Pell for sending me copies of their theses and discussions at Cordilleran Workshops. Discussions with R.L. Brown, J.A. Donaldson, L. Aspler, and R. Rainbird of Carleton University are also appreciated.

This work was supported by contracts and research agreements from the Geological Survey of Canada to R.L. Brown of Carleton University and by Natural Sciences and Engineering Research Council Operating Grant A2693, also to R.L. Brown.

CHAPTER 3

SUPRASTRUCTURE/INFRASTRUCTURE TRANSITION, EAST-CENTRAL CARIBOO
MOUNTAINS, BRITISH COLUMBIA: GEOMETRY, KINEMATICS, AND TECTONIC
IMPLICATIONS

(3-1) ABSTRACT

In the Cariboo Mountains of east-central British Columbia, syn-metamorphic D2 folds of the Proterozoic Kaza Group change from upright, open, and symmetrical, to gently inclined, tight (locally isoclinal), and southwestwardly verging. Plunge measurements and change in metamorphic grade show that the transition occurs with paleo-depth in the orogenic belt. This "suprastructure-infrastructure" transition is attributed to downwardly increasing shear strain accumulated during northeastward underthrusting of allochthonous marginal basin sediments and autochthonous distal North American continental margin sediments beneath more proximal sediments of the continental margin.

A comparison of the structural thickness of the Kaza Group with its original thickness reveals that a greater amount of bulk crustal shortening occurred at deeper structural levels than at immediately overlying shallower levels. This geometry requires that shallow structural levels to the west be separated by thrust faults from deeper structural levels. A belt of pre-metamorphic-peak, southwestwardly verging thrust faults is interpreted to be the shallow level manifestation of infrastructure shortening. Approximately 60 km of northeastward displacement of rocks at deep structural levels with respect to a fixed suprastructure can account for both the infrastructural thickening in the east-central Cariboo Mountains and the geometry of thrust faults in the Quesnel Highlands.

Although the data and interpretations presented in this paper cannot unequivocally discriminate between two tectonic models that have

been proposed to account for the kinematic evolution of the southeastern Canadian Cordillera, they impose constraints which must be satisfied by these, and other, models. The displacements determined in this study are used to define the precise structural geometries implied in the two models and to examine the geological implications of these geometries.

(3-2) INTRODUCTION

The central and southern Cariboo Mountains (Fig. 3-1) are underlain by a northwestwardly plunging structural sequence of meta-sediments of the Proterozoic Kaza Group (Campbell 1968, Campbell et al. 1973; Fig. 3-2, 3-3). Originally deposited near the base of the continental margin sequence fringing western North America (Campbell et al. 1973, Young et al. 1973, Gabrielse 1972, Wheeler et al. 1972), Kaza Group rocks were deformed and metamorphosed during early mid-Jurassic to Tertiary interaction of western North America with oceanic "suspect" terranes (Davis et al. 1978, Coney et al. 1980, Monger et al. 1982). Penetrative deformation of these rocks occurred relatively early in this interval with strains recorded before, during, and after the mid-Jurassic peak of regional metamorphism (Pigage 1977, Parrish & Wheeler 1983; Archibald et al. 1983, Nguyen et al. 1968, Campbell 1968, Campbell et al. 1973, Pigage 1978, Fletcher & Greenwood 1979, Pell & Simony 1981, Murphy & Journeay 1982, Murphy & Rees 1983). Post-metamorphic uplift of the Premier Range in the southeastern Cariboo Mountains tilted the sequence into its present northwestwardly plunging.

Fig. 3-1 Regional geological map of the Cariboo Mountains and parts of the adjacent Rocky Mountains and Quesnel Highlands (compiled and modified from Wheeler et al. 1972, Pell 1984, and Struik 1985b)

Quaternary sediments

mid-Jurassic granodiorite

Terrane I

Snowshoe Group

Cariboo Group (Cariboo Mountains); upper Miette Group (Rocky Mountains)

Kaza Group (Cariboo Mountains); middle Miette Group (Rocky Mountains)

Horsethief Creek Group (Cariboo Mountains); lower Miette Group (Rocky Mountains)

Malton gneiss

west-verging thrust fault

east-verging thrust fault

high-angle fault of unknown displacement

A. Area mapped by Carey (1984)

B. Area mapped by Pell (1984)

C. Area shown in Fig. 3-2

A - A' Line of cross section restored in Fig. 3-12 and 3-13

Fig. 3-1 Regional geological map of the Cariboo Mountains and adjacent Rocky Mountains and Quesnel Highlands.

attitude.

In mapping the Canoe River and McBride map-areas (1:250,000; Fig. 3-1), Campbell (1968) and Campbell et al. (1973) noted a distinct contrast in structural style between lower grade and higher grade rocks of the Kaza Group. Where the Group contains syn-kinematic muscovite and chlorite, it is weakly deformed, being openly to tightly folded around steep axial surfaces. In contrast, rocks that were deformed at conditions supporting the growth of garnet and higher grade minerals are tightly to isoclinally folded around gently dipping axial surfaces. Following DeSitter & Zwart (1960), Campbell (1970, 1973) and Wheeler et al. (1972) referred to the change in structural style from shallow to deep levels as a "suprastructure - infrastructure" transition and considered the difference in structural style to be a consequence of the difference in rheology between rocks being deformed under different metamorphic conditions. They thought that the total amount of shortening at all structural levels is the same and limited by the 10-20% shortening apparent in the folds of the suprastructure.

As initially outlined, this change in structural style between suprastructure and infrastructure is a tantalizing feature, potentially implying much about the kinematic evolution of this part of the Canadian Cordillera but lacking the necessary detailed documentation for a full evaluation. Of critical importance are data on: (1) the geometry, kinematics, and relative timing of strains at different structural levels, (2) the nature of the coupling between structural levels, and (3) the nature of the Kaza multilayer.

In consideration of these questions, three two-month field seasons between 1981 and 1983 were spent mapping between Castle Creek and Tete

Creek north of the Premier Range (Fig. 3-2). A series of maps and cross sections across the structural transition was prepared at 1:20,000 scale (in pocket). This scale permitted the illustration of structural geometries but precluded complete mapping of the 1400 km² region. However, the spectacular topographic relief (up to 2.5 km) of the Cariboo Mountains and lateral continuity of some of the Kaza Group lithologies made it possible to link maps and, especially, cross sections with a reasonable degree of certainty.

The purpose of this paper is to document, in detail, the nature of the structural transition from suprastructure to infrastructure and to discuss the kinematic implications of its geometry. A significant finding of this study is that the downward transition from upright to gently inclined structures involves folds with southwestward vergence. The structural geometry at even deeper structural levels is also dominated by southwestwardly verging regional-scale folds. Structures which verge away from the craton have long been recognized in the Canadian Cordillera and various models have been proposed to reconcile these structures with the more prominent (north)eastwardly verging structural features such as the Rocky Mountain Fold and Thrust Belt. These models are discussed in light of the data presented in this study.

(3-3) THE STRUCTURAL SEQUENCE OF THE KAZA GROUP

In the east-central Cariboo Mountains, a cross section to deep structural levels may be constructed from the surface geology due to the consistent northwestward plunge of structures (Fig. 3-3, 3-4, 3-5).

Fig. 3-2 Distribution of rock units in the east-central Cariboo Mountains and locations of areas (A through F) mapped in detail (in pocket).

Fig. 3-3 Longitudinal (NW-SE) cross section, from ridge north of Castle Creek to ridge south of Tete Creek, illustrating how variation in plunge of exposes deeper structural and stratigraphic levels in the southeastern Cariboo Mountains. A to F are areas shown in Fig. 3-2. The line of section passes through areas A, B, and D; data from C, E, and F are projected.

Fig. 3-4. Perspective diagram illustrating the relationship between Figs. 2, 3, and 5.

Fig. 3-5 Composite vertical cross section compiled from 30 serial cross sections (a-a' to f-f' are representative topographic profiles) of areas A through F. Relative positions of cross sections are determined from lineation data (Figs. 3, 8). Form lines are bedding traces except where indicated. Form lines outside of boxes are drawn from photomosaics of cliff faces. Ornamentation and symbols as in Fig. 3-2.

NE

COMPOSITE VERTICAL CROSS SECTION, EAST-CENTRAL CARIBOO MOUNTAINS

SW

stratigraphic facing direction

PREMIER ANTICLINORIUM

kyanite-staurolite

RAUSH

ANTICLINE

This cross section illustrates a 10 km thickness of Kaza Group rocks ranging in metamorphic grade from muscovite/chlorite zone to kyanite/staurolite zone.

The Kaza Group of the east-central Cariboo Mountains has been divided into upper, middle, and lower units following the informal stratigraphy of Pell & Simony 1984 (Chapter 2, Fig. 3-6). The upper Kaza Group consists of metamorphosed sub-feldspathic sandstone, pebbly sandstone, minor graphitic and non-graphitic phyllite (meta-shale), and rare thin marble (Murphy & Rees 1983). The middle Kaza Group consists of 5 - 20 metre-thick units of amalgamated meta-sandstone alternating with units of phyllite of approximately the same thickness; the sandstone/shale ratio is approximately 1. The middle Kaza Group is capped by a distinctive persistent 200 metre-thick unit of green laminated meta-siltstone (base), finely interlaminated grey and black phyllite, and black graphitic phyllite and marble (top). The lower Kaza Group consists of a calc-schist (meta-marl or calcareous shale), thin, relatively continuous orange to buff or grey marble, and meta-sandstone.

The top of the Kaza Group is the stratigraphic contact with the overlying Isaac Formation, which is exposed on the north side of Castle Creek (Fig. 3-2). The base of the sequence is an inferred fault separating upright rocks of the lower Kaza Group from an underlying sequence of inverted rocks which are lithologically similar to the semi-pelite and amphibolite (SPA) and middle marble units of the Horsethief Creek Group (Murphy 1984). The inferred fault is located in the western slopes of the Rocky Mountain Trench west of Valemount (Fig. 3-2).

Fig. 3-6. Stratigraphy and regional correlation of the Kaza Group of the east-central Cariboo Mountains

At all structural levels, rocks of the Kaza Group exhibit macro- and micro-structural evidence of polyphase deformation. Four sets of structures have been identified on the basis of structural superposition and relationship to the mid-Jurassic episode of metamorphic mineral growth (Fig. 3-7). Pre- to early-metamorphic D_1 structures comprise tight to isoclinal northwestwardly trending folds mappable at 1:15,000 and smaller scales, and an axial planar foliation (S_1) that is readily apparent at shallow structural levels but is almost completely transposed at deeper structural levels (Fig. 3-5, Plates 3-2a,b,c,d). Where preserved at deeper structural levels, S_1 is defined by muscovite, chlorite, and possibly biotite, implying that D_1 was not associated with high grade regional metamorphism. The amount of structural thickening due to D_1 deformation is not known, but is the small size of the structures and lack of significant metamorphism suggest that D_1 structural thickening was minor.

Syn-metamorphic D_2 structures are the dominant structures at all levels, mappable at 1:50,000 and smaller scales. D_2 deformation is responsible for the main bulk of the structural thickness across which prograde mineral isograds are printed. Randomly oriented metamorphic minerals overgrow D_2 structures implying that metamorphism outlasted D_2 deformation.

Post-metamorphic D_3 structures are found primarily at the bottom of the sequence of Kaza Group rocks, increasing in importance near the southern Rocky Mountain Trench. Post-metamorphic D_4 structures are northeastwardly trending, inclined to the northwest trend of all earlier structures. D_4 structures appear to be responsible for the large-scale variations in the magnitude of plunge of D_2 hinge lines and

SEQUENCE OF DEFORMATION AND METAMORPHISM
D_4	post- metamorphic	NE-trending upright crenulations and regional-scale warps
	D_3	late- to post- metamorphic	NW-trending, NE-verging crenulations and folds, important near faulted base of structural sequence, but dying upward; includes Premier Anticlinorium
	METAMORPHIC PEAK		
	D_2	syn- metamorphic	NW-trending, SW-verging folds; dominant at all structural levels; associated with significant structural thickening
	D_1	pre- to early- metamorphic	Small-scale, NW-trending, NE verging folds, not associated with significant structural thickening, age uncertain but may be early Jurassic, associated with the obduction of Terrane I

Fig. 3-7 Sequence of deformation and metamorphism in the east-central Cariboo Mountains. 'Time' is relative to peak of metamorphism as recorded in individual exposures.

lineations (Fig. 3-3).

For descriptive purposes, the structural sequence of the Kaza Group has been subdivided into "Suprastructure", "Infrastructure", and "Transition Zone".

(3-3a.) Suprastructure

The upper Kaza Group and part of the middle Kaza Group are represented in the suprastructure. The metamorphic grade of the suprastructure increases down-section and to the southeast from muscovite/chlorite zone to biotite zone.

The structural geometry of the suprastructure is dominated by regional-scale, open to close, northwestwardly trending and plunging, southwestwardly verging folds (Figs. 3-5 and 3-8; Murphy & Rees 1983). At the level of the contact between the Kaza Group and the Isaac Formation, the axial surfaces of these folds dip moderately to steeply to the northeast. Above this level, the axial surfaces are more nearly upright and folds more open (Campbell et al. 1973). First-order folds (wavelength > 2 km) are open to close, rotating bedding or enveloping surfaces through relatively small angles. Steep to overturned bedding is commonly found on short limbs of second-order folds; however, the suprastructure stratigraphy is dominantly upright.

Southwestwardly verging folds in the suprastructure have a spaced axial planar foliation, S_2 (Fig. 3-8, Plates 3-1a, b). S_2 fans in hinge zones and refracts at meta-sandstone - meta-shale contacts. Locally, where the angle of refraction is large and S_2 in meta-shale layers is subparallel to bedding, S_2 is folded by crenulations possibly related to further tightening of the fold. Graded bedding is clearly

Fig. 3-8 Equal-area stereoplots of D_2 planar (S_2) and linear (L_2 , L_{2e} , and F_2) fabric elements

PLATE 3-1

- a. Spaced D₂ cleavage (S₂) produced by the crenulation of chlorite - muscovite foliation (S₁). Organic-rich seams reflect dissolution of quartz on short limbs of asymmetrical crenulations. Scale bar is 0.5 mm.
- b. Randomly oriented muscovite overgrowing D₂ crenulations. Scale bar is 0.5 mm.
- c. View of top of bedding surface of pebbly sandstone. Intersection of S₂ with bedding produces linear fabric L₂. The maximum elongation direction of quartz and feldspar clasts (L_{2e}) is parallel to L₂.
- d. View (looking down-plunge, to the northwest) of mesoscopic planar fabrics. Bedding (stratigraphically upright) is defined by darker phyllite and lighter sandstone beds. Northeastwardly verging S₁ and southwestwardly verging S₂ are indicated.

PLATE 3-1

outlined by the smoothly continuous sigmoidal trace of S_2 as it passes from the coarser to the finer fraction.

In phyllites, S_2 is a crenulation cleavage, produced during the folding of an earlier chlorite-muscovite foliation (S_1 ; Plate 3-1a,b). The planar fabric itself is defined by planar concentrations of insoluble residue and micas on the short limb of asymmetric crenulations (Plate 3-1a). Mica-rich and quartz-rich segregations often result in a poorly defined compositional layering at thin-section scale. In meta-sandstones and pebbly sandstones, S_2 is defined by the dimensional preferred orientation of flattened and extended quartz, and to a lesser extent, feldspar clasts. S_2 intersects bedding (S_0) to produce an intersection lineation, L_2 which is parallel to the hinges of D_2 folds (F_2) and L_{2e} , the finite extension direction of elongate clastic grains (Plate 3-1c).

S_1 is macroscopically folded by D_2 folds and appears to be related to small-scale, tight to isoclinal, northwestwardly trending and plunging, northeastwardly verging folds (D_1) whose axial surfaces are clearly folded by D_2 folds (Fig. 3-5). On stratigraphically upright D_2 limbs, where S_2 is less pervasive, S_1 is observed to be northeastwardly verging (Plate 3-1d).

In the low-grade suprastructure, S_2 is overgrown by prograde porphyroblasts of biotite, muscovite, and chlorite (Plate 3-1b). Thus, D_1 and D_2 folds formed prior to the peak of regional metamorphism. Post- D_2 porphyroblasts and all earlier structures are deformed by a northeastwardly trending set of macroscopic crenulations with steep axial surfaces (D_4). These transverse structures may be responsible for the large-scale variations in the magnitude of plunge of D_2 and

earlier structures (Fig. 3-3, 3-8), although this relationship remains to be proven (Murphy & Rees 1983).

(3-3b.) Infrastructure

Rocks of the infrastructure belong to the middle and lower Kaza Group of Pell & Simony (1984). The metamorphic grade of the infrastructure increases down-section and to the southeast from garnet zone to kyanite/staurolite zone (Campbell 1968).

The structural geometry of the infrastructure is best described with reference to the Raush Anticline, a northwestwardly trending, gently inclined, regional-scale, southwestwardly verging anticline (Fig. 3-5). The Raush Anticline and associated parasitic structures deform bedding, an earlier foliation (S_1), and a set of smaller-scale northeastwardly verging D_1 folds (Fig. 3-5). On this basis, the Raush Anticline and associated parasitic structures are considered to be, locally at least, D_2 structures.

The Raush Anticline has been identified by the inversion of stratigraphy accompanied by the reversal of vergence of second-order, though still regional-scale, structures. The long upper limb is characterized by tight, southwestwardly verging, large-scale folds of a dominantly upright stratigraphy. The short lower limb is characterized by tight to isoclinal, northeastwardly verging folds of a dominantly inverted stratigraphy. The hinge zone, characterized by symmetrical folds with steep enveloping surfaces, has been mapped along a small portion of the ridge between Black Martin Creek and Raush River and has been observed, but not mapped, on the north wall of the Raush River Valley (Fig. 3-2).

Overtuned middle Kaza Group stratigraphy of the lower limb of the Raush Anticline has been traced to the ridge north of South Kiwa Creek (Fig. 3-2; Fig. 3-5, Location W). On the ridge south of South Kiwa Creek between Mt. Sir John Oliver and Mt. MacKenzie-Bowell (Fig. 3-2; Fig. 3-5, Location X), predominantly upright middle and lower Kaza Group stratigraphy is found. The contact between the middle and lower Kaza Group can be traced eastwardly through prominent southwestwardly verging isoclinal folds into a zone of symmetrical isoclinal folds with steep enveloping surfaces (Fig. 3-2; Fig. 3-5, Location Y). Although not completely mapped, upright lower and middle Kaza Group rocks may possibly close continuously into the short inverted limb of the Raush Anticline at this locality. However, further to the west, on the north side of South Kiwa Creek (Fig. 3-2; Fig. 3-5, Location W), layer-parallel shear zones are found within the inverted limb, suggesting that the inverted limb may be detached from the underlying upright stratigraphy to the west and southwest.

The predominantly upright panel of middle and lower Kaza Group continues to the southernmost extent of the area mapped. West of Mt. Sir John Oliver (Fig. 3-2), the panel and the southwestwardly verging folds affecting it, arch over the Premier Anticlinorium, an open and upright D_3 fold (Fig. 3-5). To the east towards the southern Rocky Mountain Trench, D_3 folds are tighter and more clearly northeastwardly verging. Just to the west of the Trench, visible from Valemount, D_2 and earlier structures steepen through the vertical in interference with D_3 structures, creating the "fan axis" of Campbell (1968) and Wheeler *et al.* (1972).

Upright lower Kaza Group rocks overlie, with apparent fault contact, an inverted sequence of amphibolite, quartz-plagioclase-biotite-garnet meta-sandstone(?), kyanite-staurolite-garnet bearing meta-pelite, and laminated marble and calc-silicate-rich rock that has been intruded by a pre-kinematic (pre- D_1) diorite (Fig. 3-5). This sequence is tentatively correlated with the semi-pelite and amphibolite (SPA) and middle marble units of the Horsethief Creek Group (Murphy 1984).

S_2 , the axial surface fabric of D_2 folds, is a crenulation cleavage. The morphology of S_2 changes with rock type and metamorphic grade. Plates 3-2a and b are from the hinge zone of a northeasterly verging second-order fold on the lower limb of the Raush Anticline. In the graphitic layer (Plate 3-2b), S_2 is clearly a crenulation cleavage, overgrown by coarse, randomly-oriented porphyroblasts of biotite and muscovite. In the less graphitic layer, the hinges of D_2 crenulations are obscured by micas aligned in S_2 , and randomly oriented, post-kinematic micas. At higher metamorphic grades, the hinges of the crenulations are entirely recrystallized in the less graphitic layers and are only visible where graphitic concentrations are found (Plate 3-2c). Recrystallized hinge zones are suggested where isolated coarse mica grains are elongate parallel to the foliation but with their 001 cleavages inclined to the foliation at an acute angle (Plate 3-2d).

In coarser clastic rocks, S_2 is defined by the plane of flattening and extension of deformed quartz and feldspar clasts. S_2 intersects S_0 in an intersection lineation (L_2) that parallels the hinges of D_2 folds (Fig. 3-8). L_2 is also parallel to the direction of extension of

PLATE 3-2

- a. Syn-metamorphic (but pre-metamorphic peak) D_2 crenulations at intermediate structural levels. Scale bar is 0.5 mm. See text for discussion.
- b. Close-up of crenulations in Plate 3-2a. Note the randomly oriented muscovite laths in both hinge and limb zones of crenulations. Scale bar is 0.15 mm. See text for discussion.
- c. Rare, relict, intermediate to deep level D_2 crenulations. Crenulations are defined by incompletely recovered polygonal arcs of muscovite, chlorite, and graphite (?). The prominent foliation in the rock (also S_2), defined by parallel laths of coarse-grained muscovite and biotite, is parallel to axial surfaces of crenulations and major southwestwardly verging folds in the region. Scale bar is 0.2 mm.
- d. At the highest metamorphic grades (deepest structural levels), D_2 crenulations are almost completely recrystallized into a foliation defined by parallel laths of muscovite and biotite. A relict hinge is possible present at the top of the photomicrograph. Inclination of basal cleavages of micas to S_2 is interpreted to be due to nucleation of coarse micas in hinge zones of crenulations. Scale bar is 0.15 mm.
- e. Upright D_4 crenulations deforming S_2 . View is to the northeast.
- f. Orthozoisite porphyroblasts aligned primarily parallel to L_2 and L_{2e} (Fig. 3-8). A few randomly oriented (post-kinematic) porphyroblasts are also visible. Length of object in lower right hand corner is approximately 3 cm.

PLATE 3-2

quartz and feldspar clasts and the preferred dimensional orientation of prismatic and tabular porphyroblasts (L_{2e}).

At all metamorphic grades, S_2 , L_2 , and L_{2e} are overgrown by randomly oriented porphyroblasts (garnet, biotite, muscovite, orthozone (Plate 3-2f), plagioclase, staurolite, and kyanite). These porphyroblasts and earlier fabrics are deformed by two sets of post-metamorphic crenulations: northwestwardly trending D_3 and northeastwardly trending D_4 (Plate 3-2e). With the exception of D_3 , the relative sequence of events: D_1 , D_2 , metamorphic peak, and post-metamorphic D_3 and D_4 , is the same as is found in the suprastructure. On this basis, southwestwardly verging, pre-metamorphic peak, D_2 folds in the suprastructure may be tentatively correlated with D_2 folds in the infrastructure. As shown in the following section, physical continuity can be established, thus confirming that southwestwardly verging D_2 folds of the suprastructure and infrastructure are, indeed, synchronous structures.

(3-3c.) Transition zone

The transition from upright to recumbent folding occurs totally within rocks of the middle Kaza Group, near the biotite and garnet mineral isograds (Fig. 3-5). To elucidate the nature of this transition, the axial surface of an individual D_2 syncline (indicated by arrow in northwest corner of Fig. 3-2; Fig. 3-5) was mapped down-section from the suprastructure. The axial surface was traced 25 km from Castle Creek southeastwardly to the first drainage northwest of Raush River (Fig. 3-2). Along its trace, the axial surface gradually decreases in dip (Fig. 3-5). The eastern limb of the syncline, which

is the short limb of a southwestwardly verging anticline/syncline pair, becomes shorter down-section as the axial surfaces of second-order folds become more gently dipping and more densely distributed. Thus, in profile (Fig. 3-5), the axial surface trace flattens downward, eventually disappearing into a zone of southwestwardly verging folds with gently dipping axial surfaces.

The western limb of the syncline in the transition zone is only weakly folded (Fig. 3-5). Individual continuous meta-sandstones and phyllites/schists may be traced for several kilometres without passing through any major folds. Those folds that are present are relatively small-scale and consistently verge to the southwest, parasitic on the upper limb of the Raush Anticline of the infrastructure. The closure of the Raush Anticline on the north side of the Raush Valley lies approximately two kilometres structurally deeper than these relatively undeformed continuous beds.

The transition from upright to near-recumbent folds is accompanied by a continuous change in the orientation and style of S_2 . S_2 is generally parallel to D_2 axial surfaces at all structural levels; as such, S_2 undergoes the same transition from steeply to gently dipping orientations. In addition, cleavage fanning and convergence in D_2 hinge zones is less apparent at depth, cleavage refraction is less apparent at depth, and the angle between cleavage and bedding on fold limbs decreases with depth.

Structure and stratigraphy may be traced continuously across the transition from suprastructure to infrastructure, uninterrupted by faults or discrete shear zones. Southwestwardly verging folds with steep axial surfaces of the suprastructure are parasitic on gently

inclined deeper level structures such as the Raush Anticline. The transition occurs over a vertical profile distance of approximately 5 km and involves a structural rotation of approximately 70° .

(3-4) KINEMATIC ANALYSIS

The structural geometry of the east-central Caribbo Mountains results from the superposition of four phases of deformation. However, the transition from upright to near-recumbent folding and the structural thickness of the Kaza Group are primarily the results of D_2 deformation. In this section, the structural geometry of D_2 deformation is approached from two perspectives: (1) the regional strain involved in the transition from upright to near-recumbent folding, and, (2) the amount and distribution of structural thickening at different structural levels.

(3-4a.) Transition from upright to near-recumbent folding

Upright, relatively symmetrical, folds and steeply dipping foliations at shallow structural levels represent approximately 10 - 20% horizontal shortening (Campbell 1973). The gently inclined folds and foliations found at deeper structural levels can not be formed purely by bulk horizontal shortening; consequently, the finite strain at deep levels includes a significant sub-horizontal shear component. The continuous overturning of folds with depth reflects the increase in importance of the shear component with depth. Indeed this gradient in shear strain is similar to that expected at the margins of shear zones of any scale (Ramsay & Graham 1970, Sanderson 1982).

The variation in the orientation of foliations in a thrust sheet or nappe undergoing combined boundary-parallel shortening and shear

strain has been modelled by Sanderson (1982). Upright and symmetrical folds are predicted for shallow structural levels where boundary-parallel shortening strain is more important than shear strain.

With depth and increasing shear strain, tighter, more asymmetrical, and progressively more overturned folds and foliations result. At very high shear strains, rotation of structures into the extensional field of the incremental strain ellipsoid causes extension of folded units and reversal of fold asymmetry (Ramsay et al. 1983).

Using the relationship, presented in Sanderson (1982), between (ϕ), the angle between the shear direction and the trace of the XY kinematic plane in the plane containing the shear direction and the normal to the shear plane, and R, the strain ratio, for different shear strains (γ) and stretches parallel to the direction of transport (α), the amount of shear strain at different structural levels has been estimated (Fig. 3-9). This exercise assumes a shear plane that was originally dipping gently to the east relative to bedding. However, ϕ is measured from the horizontal because of the near-horizontal orientation of the axial surfaces and foliations of the tightest folds in areas where post- D_2 deformation is minimal. The sense of rotation of D_2 axial surfaces and foliations suggests that the displacement or shear direction is approximately southwest, parallel to the line of section of Fig. 3-5.

Shear strains have been estimated for suprastructure shortenings of 10 and 20% ($0.9 < \alpha < 0.8$). They range from 0.15 at 10% suprastructure shortening to 3.9 at a suprastructure shortening of 20%. The total northeastward displacement of the infrastructure with respect to a fixed suprastructure ranges from 6.2 km, for suprastructure shortening

Fig. 3-9 Kinematic analysis of transition from upright folds and steeply dipping foliations to near-recumbent folds and gently dipping foliations

- a. Change-in orientation of S_2 with depth. Angle at right is the angle between the average foliation and a horizontal plane of shear.
- b. Graph showing relationship between ϕ' , the angle between a foliation and its associated plane of shear, and R , the strain ratio, for different values of the components of γ simple shear strain, and α , longitudinal elongation or shortening ($\alpha = 1$ is simple shear strain). Large dots on $\alpha = 0.8$ and $\alpha = 0.9$ curves are the values of ϕ' shown in Fig. 3-9a. From Sanderson (1982).
- c. Values of R and horizontal displacements associated with the values of γ determined in Fig. 3-9b. The heights of individual sheared boxes include vertical extension associated with α ; as such, markers would have different original heights.

trace of S_2 in plane of section

a.

b.

c.

of 10%, to 7.06 km, for a suprastructure shortening of 20%.

(3-4b.) Amount and distribution of structural thickening of the Kaza Group

In the infrastructure portion of the structural sequence of the Kaza Group, stratigraphy is duplicated by isoclinal folding around gently dipping axial surfaces. This geometry implies a significant amount of structural thickening, and therefore shortening, that is not present in the suprastructure. The small size and low metamorphic grade of pre- and post- D_2 structures suggest that most of this structural thickening was produced by D_2 deformation. If one assumes that most of the structural thickening is due to D_2 deformation, the amount and distribution of structural thickening can be estimated by a comparison of the thickness of the deformed sequence with its original stratigraphic thickness.

Although the original thickness of the Kaza Group at this locality is not precisely known, it can be estimated from the measured thicknesses of a less highly folded, continuously upright section of the Kaza Group located south of the Premier Range (Pell & Simony 1984, Pell 1984), and a stratigraphic section of the probably equivalent middle Miette Group of the Rocky Mountains (Carey & Simony 1984, Carey 1984). These two sections (Fig. 3-6) are quite similar and it is reasonable to assume that the stratigraphic thickness of the Kaza Group involved in the structural section described in this study, which lies between the other two sections, is within the range of thickness provided by the other two sections. This assumption may not be strictly correct and original thickness variations within the Kaza Group may have been present prior to deformation, although they are not

likely to be very large given the proximity of the sections to one another.

As is shown in Fig. 3-6, the structural sequence of the Kaza Group described in this report is more than double the thickness of either of the other two sections. It is also apparent that structural thickening of the Kaza Group is unevenly distributed throughout the sequence. The upper Kaza Group is similar in thickness in all three sections. In contrast, the middle Kaza Group, where involved in the Raush Anticline and related structures, is 5.5 times thicker than the other Kaza Group section and 3.7 times thicker than the middle Miette section.

In order to produce the observed distribution of structural thickening, there must have been a greater amount of crustal shortening at depth than at shallow structural levels. This difference requires that suprastructure and infrastructure become detached as shear strains cut to higher structural levels in the direction of hanging wall transport (Fig. 3-10). Kaza Group rocks to the southwest are predicted to be in reverse fault contact with either stratigraphically younger rocks as faults ramp into higher structural levels (Location B, Fig. 3-10) or with stratigraphically older rocks on the basal flat (Location A, Fig. 3-10).

The former relationship has been observed near Quesnel Lake (Fig. 3-1) where Kaza Group rocks are in the hanging wall of the structurally highest (yet identified) thrust of a southwestwardly verging, pre-metamorphic peak fold and thrust belt (Struik 1981, 1982a,b; 1983a,b). The nature of the strain in these rocks is not completely understood and the expression "fold and thrust belt" may be an oversimplification;

Fig. 3-10 Diagrammatic model illustrating the relationship between infrastructural profile thickening and thrust faulting in the Quesnel Highlands, section balanced and to scale.

however, pre-metamorphic peak, southwestwardly verging, low-angle reverse faults are documented. It is proposed that the structural thickening and shear strain evident in the Cariboo Mountains link laterally and up-section into this belt of low-angle faults.

(3-4c.) Magnitude of crustal displacements

The change in orientation of structures between suprastructure and infrastructure represents approximately 7 km of northeastward displacement of the infrastructure with respect to a fixed suprastructure (Fig. 3-9). The difference in crustal shortening between suprastructure and infrastructure requires additional northeastward displacement of the infrastructure with respect to a fixed suprastructure. To determine the magnitude of this additional displacement, the cross sectional area of the thickened middle Kaza Group in Fig. 3-5 has been restored to an idealized pre-D₂ configuration (Fig. 3-10). This scaled restoration establishes that approximately 40 km of northeastward displacement of the infrastructure with respect to a fixed suprastructure can account for the structural thickness of the middle Kaza Group. Thus, the structural geometry shown in Fig. 3-5 represents approximately 50 km of northeastward underthrusting of middle Kaza Group and structurally deeper rocks with respect to a fixed suprastructure of upper Kaza Group and younger rocks.

In the Quesnel Highlands (Fig. 3-1), the geometry of thrust slices in the hanging wall of the structurally lowest Pleasant Valley Fault records approximately 70% shortening (Struik 1980). A comparison of the deformed and palinspastic sections above the Pleasant Valley Fault

(Struik 1981) suggests that 25 - 30 km of northeastward displacement of the footwall is associated with the shortening. The displacement on the Pleasant Valley Fault is unknown due to the lack of a well-defined footwall cut-off. However, the structural topography of regional-scale culminations and depressions such as the Lanezi Arch, the Isaac Lake Synclinatorium, and the Black Stuart Synclinatorium (Fig. 3-1) may be due to changes in the trajectory of the Pleasant Valley Fault at depth. On preliminary analysis, a reasonable location of a footwall cut-off would be beneath the Lanezi Arch, suggesting a displacement of approximately 35 km on the Pleasant Valley Fault.

A total of 60 - 65 km of northeastward underthrusting is therefore associated with the fold and thrust belt in the Quesnel Highlands. The footwall of the Pleasant Valley Fault also contains southwestwardly verging folds, including spectacular folds of the previously active obduction boundary between an oceanic "suspect terrane" (Terrane 1) and North America (Rees 1981, Rees & Ferri 1983). These folds may represent partitioning of displacement into footwall strain; as such, the displacement on individual faults is expected to vary along their trajectories.

The way in which relatively discrete displacements in the Quesnel Highlands transform into structural thickening in the infrastructure further east is not known. However, both the fold and thrust belt and zone of infrastructure thickening lie at the interface between a relatively uninvolved hanging wall or suprastructure and a footwall that has been underthrust to the northeast by 50 to 65 km. The hanging wall has been thickened by successive accretion of material stripped off the footwall by underplating of large-scale infrastructure

folds and forward-stepping of shallow level thrust faults (Fig. 3-10). The fold and thrust belt and the zone of infrastructure folding may be viewed as a broad crustal shear zone accommodating shear strains associated with northeastward underthrusting.

(3-5) TECTONIC IMPLICATIONS

Mid-Jurassic, syn-metamorphic (south)westwardly verging structures have been recognized in many localities in the eastern Canadian Cordillera (Fig. 3-11). In most of these localities, these structures fold an earlier (north)eastwardly verging set of structures (D_1 ?), including the Quesnel Lake Shear Zone (Brown et al. in press; Fig. 3-1), which marks the obduction boundary between Terrane I and North America. This boundary was active after the Sinemurian (late Early Jurassic), but before metamorphism and (south)westwardly verging deformation. Additionally, in most of these localities, (south)westwardly verging structures are deformed by (north)eastwardly verging structures. These structures are late- to post-metamorphic and are spatially associated with early thrust faults which feed into the Rocky Mountain Foreland. In light of the magnitudes of (north)eastward displacements suggested for the Rocky Mountain Fold and Thrust Belt (100-200 km, Bally et al. 1966, Price & Mountjoy 1970; Price & Fermor 1981; Campbell et al. 1980), Monashee Decollement (>80 km, Read & Brown 1981; 100 km Brown et al. in press), and the Quesnel Lake Shear Zone (>60 km, Brown et al. in press), these regionally distributed (south)westwardly verging structures are somewhat enigmatic. Two tectonic models (Figs. 3-12 and 3-13) have been proposed to account for

Fig. 3-11 Areas in the southeastern Canadian Cordillera in which regional-scale southwestwardly verging structures have been described

the presence of these (south)westwardly verging structures in an orogenic belt characterized by large-scale (north)eastwardly verging structural features: (1) Obduction of Terrane I along the Quesnel Lake Shear Zone and equivalent (D_1 ?) structures is followed by the insertion of the remainder of Terrane I as a rigid indenter along the basement/cover contact of the North American continental margin prism, delaminating the cover from the basement and inducing thickening and metamorphism in advance of the rigid indenter. Continued (north)eastward displacement of the indenter on its basal shear zone results in folding of inactive southwestwardly verging structures and initiation of shortening in the Rocky Mountain Foreland (Price 1983, Archibald et al. 1983, Price, et al. 1985).

(2) Obduction of Terrane I on the Quesnel Lake Shear Zone and equivalent (D_1 ?) structures is followed by an eastward transfer of the active boundary between Terrane I and North America into the North American continental margin prism and its underlying crustal basement. The new boundary dips to the east and subsequent convergence is accommodated by underplating the transitional crust that originally underlay the continental margin prism to beneath the transitional crust still attached to cratonic North America. After crustal thickening and metamorphism, (north)eastward vergence is re-established with the formation of a new, westwardly dipping boundary. (North)eastward displacement on this boundary and associated simple shear strain results in folding of inactive southwestwardly verging structures and the initiation of shortening in the Rocky Mountain Foreland (Monger 1977b, Brown 1978, Brown 1981, Brown et al. 1984 and in press).

In both of these models, basement is shortened and thickened to

Fig. 3-12 One tectonic model for the formation of (south)westwardly verging structures in the southeastern Canadian Cordillera. In this model, (south)westwardly verging structures are produced by the insertion of a delaminating wedge into the base of the North American continental margin sequence along a (north)eastwardly verging shear zone. See text for discussion. Sections are balanced and to scale.

SCALED RECONSTRUCTIONS OF THE MIDDLE AND EARLY JURASSIC CONTINENTAL MARGINS

MODEL 1

Fig. 3-19 Alternate tectonic model for the formation of (south) westwardly verging structures in the southeastern Canadian Cordillera. In this model, (south)westwardly verging structures are formed by the underthrusting of the edge of the North American craton by the transitional or oceanic basement to the sediment prism fringing western North America. See text for discussion. Sections are balanced and to scale.

SCALED RECONSTRUCTIONS OF THE MIDDLE AND EARLY JURASSIC CONTINENTAL MARGINS

MODEL 2

accommodate supracrustal displacements. The models differ in how, and where, basement is shortened; as such, the geological signature of each model should be distinctive and recognizable in the geological record. The implications of each model can be examined by reconstructing the Cordilleran continental margin that existed just after D_2 deformation, but before shortening of the Rocky Mountain Foreland (Figs. 3-12 and 3-13). The reconstruction is based on the balanced restoration of displacements associated with shortening of the Rocky Mountain Foreland along the trajectories shown in Figs. 3-12a and 3-13a (modified from Brown et al. in press).

To construct D_2 displacement trajectories, the eastern limit of D_2 deformation was assumed to have fundamental significance in each model. In Model 1, this eastern limit is interpreted as the location of the leading edge of the delaminating wedge, where the oppositely verging shear zones bounding the wedge intersect. In Model 2, the eastern limit is interpreted to be where the trajectory of the southwestwardly verging shear strain leaves the crystalline basement and enters the supracrustal sediments. In both models, this eastern limit essentially defines a hanging wall cut-off for each model and, when combined with displacements, can be used to determine internally consistent displacement trajectories and the corresponding footwall cut-offs.

Four preliminary conclusions can be drawn from the geometries shown in Figs. 3-12 and 3-13:

- (1) The footwall cut-off of the delaminating wedge (Model 1) is constrained to lie in the basement to the North American continental margin prism. Thus, the leading edge of the rigid indenter would be composed of transitional continental margin basement, not the oceanic

suspect terrane that has been proposed for a similar geometry in the Kootenay arc of southern British Columbia (Price, 1983, Archibald et al. 1983). Displacements on the order of 150 km would be necessary to transport Terrane I rocks from their early Jurassic position outboard of the North American continental margin prism to beneath rocks presently at the Southern Rocky Mountain Trench.

(2) Model 2 predicts that the cratonic thickness basement which presently underlies the Cariboo Mountains consists partly of a slice of allochthonous transitional crust which was underplated beneath the autochthonous transitional crust still attached to the craton. If Model 2 is correct, the westward transition from 50-55 km-thick crust to 30-35 km-thick crust which occurs beneath the Cariboo Mountains and Quesnel Highlands may be related to the geometry of D_2 deformation rather than to Paleozoic continental rifting as interpreted by numerous workers (Monger & Price 1979, Price 1981, Price & Fermor 1981, Chamberlain & Lambert 1985).

(3) Model 1 predicts that the westward transition from cratonic to attenuated transitional (or oceanic?) crustal basement produced by crustal extension during the formation of the Cordilleran miogeocline lay 100 to 150 km west of the autochthonous crust which presently underlies the Southern Rocky Mountain Trench (Fig. 3-1). Model 2 predicts that a lense of 15 to 25 km-thick crustal basement underlay the thick central part of the continental margin prism, with 10 to 12 km-thick, attenuated transitional (or oceanic?) crust to the east and west. None of the transitions from cratonic to transitional or oceanic crust indicated in any of the deformed or restored sections correspond with the transition from 50-55 km-thick, low electrical conductivity

crust to 30-45 km-thick, higher conductivity crust that is observed at the Southern Rocky Mountain Trench (Price 1981, Chamberlain & Lambert 1985). The deformed and restored sections imply that this transition is due to Mesozoic tectonism, rather than extension during the formation of the Cordilleran continental margin.

(4) The difference in style and location of basement thickening of the two models would be expected to have different implications for the geological record of the Rocky Mountain Foreland. The loading of basement and thickened sediments onto a footwall still attached to the North American craton (Model 1) would be expected to depress the footwall, inducing subsidence and marine transgression in the Rocky Mountain Foreland. Alternatively, underthrusting of transitional crust beneath a hanging wall of transitional crust still attached to North America (Model 2) would be expected to induce uplift in the Rocky Mountain Foreland. The uplift in the Rocky Mountain Foreland would be somewhat counterbalanced by the loading effects of the deformed supracrustal sediments at the edge of the craton, which, as in Model 1, would be expected to depress the edge of the craton.

During the Sinemurian to Callovian interval in which D_2 was taking place, the Rocky Mountain Foreland was characterized by periods of deposition of shallow marine sediments alternating with periods of non-deposition in which little or no erosion occurred (Frebold and Tipper 1970). Thus, the Rocky Mountain Foreland ranged from weakly positive to weakly negative, apparently uninfluenced by the deformation that was occurring to the west. Such a geological history is not conclusive, but is mildly supportive of Model 2 which is expected to produce minor uplift in the Rocky Mountain Foreland.

The data and interpretations presented in this paper are useful in constraining the two models that have been proposed for the Jurassic Cordilleran convergent margin, but they cannot unequivocally discriminate between them. The two models must be refined by further detailed field work and tested mechanically and experimentally before firm conclusions can be drawn.

(3-6) CONCLUSIONS

- (1) The transition from upright to recumbent structures in the Cariboo Mountains involves syn-metamorphic, but pre-metamorphic peak, southwestwardly verging D_2 structures.
- (2) The transition is gradual and continuous and results from shear strain produced by northeastward underthrusting of the North American continental margin prism.
- (3) Structural thickening at infrastructure levels is much greater than at suprastructure levels, requiring that shallow and deep structural levels be detached southwest of this portion of the Cariboo region. A southwestwardly verging syn-metamorphic fold and thrust belt near Quesnel Lake is likely to be the shallow level manifestation of the deep level shortening evident in the east-central Cariboo Mountains.
- (4) The structural geometry and distribution of structural thickening of the Kaza Group in the east-central Cariboo Mountains can be accounted for by approximately 50 km of northeastward underthrusting of deep structural levels with respect to a fixed suprastructure.
- (5) This amount of displacement has been combined with the eastern limit of southwestwardly verging deformation to specifically define the

structural geometries implicit in the two different tectonic models which have been proposed to account for backfolds and backthrusts in the southeastern Canadian Cordillera. Both of these geometries are feasible and do not violate any geological or geophysical observations. With the present data base, it is not possible to unequivocally discriminate between the two models.

Acknowledgements I wish to thank Larry Lane, Chris Rees, and Gerry Ross of Carleton University, Murray Journeay of Queen's University, Richard Campbell of the Geological Survey of Canada, and, especially, Richard Brown of Carleton University for objective and stimulating discussions, critical reviews of early drafts of this paper, participation in the field work, and continued interest and support of the project. A large portion of this research was conducted while I was a visiting student at the Department of Geological Sciences of the University of British Columbia; I would like to thank Hugh Greenwood for the opportunity to study and use the facilities there. I am also grateful to Ed Montgomery and Gord Hodge of U. B. C. for their assistance in preparing photographic plates and to Mary Jean Hood of the Department of Continuing Education at U. B. C. for providing access to a word processor.

This project was funded by contracts and research agreements with the Geological Survey of Canada and Natural Science and Engineering Research Council operating grant A 2693, all to Richard L. Brown of Carleton University.

CHAPTER 4

LITHOSPHERIC STRENGTH AND THE KINEMATIC EVOLUTION OF THE SOUTHEASTERN CANADIAN CORDILLERA

(4-1) ABSTRACT

The kinematic evolution of the southeastern Canadian Cordillera between latitudes 52N and 54N is characterized by an eastward migration of the active boundary of convergence between North America and western oceanic terranes, and two reversals in regional structural vergence. To investigate possible mechanical controls on the kinematics of convergence, the thermal and geological structure of the Jurassic continental margin at three significant stages in its history of shortening were palinspastically restored from a balanced structure section and analyzed with the use of strength vs. depth profiles. The resulting models reproduce satisfactorily the observed distribution of brittle and ductile deformation in this part of the Canadian Cordillera and provide a mechanical basis for the localization and orientation of displacement trajectories during shortening of the continental margin. The initial eastward migration of the active boundary of convergence between North America and the oceanic terranes from outboard of the continental margin sediment prism into the thick central part of the continental margin prism can be explained as a consequence of thermal weakening of the continental margin lithosphere during the obduction of the oceanic terrane. The eastward migration of the convergence boundary into the inner part of the continental margin lithosphere is explained as a consequence of thermal weakening of the lithosphere during prograde regional metamorphism. Reversals in regional structural vergence appear to occur because the stress necessary to

continue displacement on one displacement trajectory exceeds the failure strength of a displacement trajectory that dips in the opposite direction.

(4-2) INTRODUCTION

A common observation in compressional orogenic belts is that major thrust sheets often include slices of crystalline basement and/or high grade metamorphic rocks whose prograde mineral assemblages equilibrated during faulting. Such occurrences vividly illustrate that thrust faults in shallow level foreland belts continue as relatively discrete structures to deep structural levels in the hinterland regions of the orogenic belt. This direct geometric linkage between foreland and hinterland implies that the kinematic and mechanical evolutions of foreland and hinterland regions are inextricably intertwined. As such, mechanical models to explain the kinematic evolution of orogenic belts must integrate the mechanical behaviors of rocks at shallow structural levels with the behaviors of rocks at deeper levels, including the crystalline basement.

The mechanics of compressional deformation at shallow structural levels is far better understood than that of deeper levels. The geometries of many thrust belts and accretionary prisms have been precisely documented with the use of balanced cross sections supported by drill and seismic reflection data and mechanical parameters have been measured directly in modern convergent zones. In contrast, mechanical models of deformation at deep structural levels must rely almost exclusively on surface data from uplifted metamorphic

hinterlands. These regions are often geometrically, kinematically, and thermally complex and geometric and thermal reconstructions are plagued with significant uncertainty. The approximate nature of thermal data is a major problem in light of the temperature dependence of the ductile deformation mechanisms operative at these depths.

In spite of these uncertainties, geometric and thermal models of hinterland regions have evolved to a point where preliminary investigations of the mechanical controls on hinterland evolution are possible. Balanced cross sections are being extended into hinterland regions (Elliott & Johnson 1980, Price 1981, Boyer & Elliott 1982, Coward 1983, Butler & Coward 1984, Brown et al. 1984 and in press, Monger et al. 1985, Butler 1985) and although the partitioning of displacements into translation along shear zones, vertical thickening or thinning, longitudinal elongation, and volume loss is in a preliminary stage in most orogenic belts (Butler 1985), internally consistent and retrodeformable (Suppe 1980) geometries have been constructed. As well, experimental data on the mechanical behavior of geological materials at different conditions of temperature, pressure, and strain rate have been extrapolated to lithospheric conditions and geological strain rates and summarized on lithospheric strength profiles (Goetze & Evans 1979; for a recent review, see Kirby 1983). Profiles constructed for the present day lithosphere have been somewhat successful in explaining the geophysical characteristics of different tectonic settings (Vink et al. 1984, Glazner & Bartley 1985, Ranalli & Murphy in prep.); this success in modelling current tectonic environments lends promise to the application of this technique to paleo-tectonic settings.

In this paper, lithospheric strength profiles are constructed on scaled palinspastic reconstructions of a portion of the southeastern Canadian Cordillera (between latitudes 52 and 54N) to investigate its early to late Jurassic kinematic evolution. The first part of the paper is a review of the geological relationships that form the basis for kinematic models of the southeastern Canadian Cordillera. In the second part, the geometric and thermal structures of three important stages in the shortening of the Jurassic continental margin are reconstructed from a retrodeformable cross section and various thermal considerations. In the third part, profiles of lithospheric strength vs. depth are used to model the rheological structure of the orogenic belt at these three stages; the results of this analysis are then used to evaluate the mechanical controls on the kinematic evolution of this part of the Canadian Cordillera.

(4-3) KINEMATIC HISTORY OF THE SOUTHEASTERN CANADIAN CORDILLERA (EARLY TO LATE JURASSIC)

The southeastern Canadian Cordillera is underlain by the deformed, metamorphosed, and intruded remnants of the Proterozoic to early Mesozoic continental margin prism of western North America, and allochthonous mid-Paleozoic to early Jurassic marginal basin and volcanic arc terranes (composite Terrane I of Monger et al. 1982; Fig. 4-1) which accreted to North America in the early to mid-Jurassic (Monger et al. 1982). Although middle and late Paleozoic deformation, metamorphism, and plutonism and Tertiary extensional deformation and plutonism are locally important, the patterns of deformation and

Fig. 4-1 Regional geological map of the Cariboo Mountains and adjacent Rocky Mountains and Quesnel Highlands (compiled and modified from Wheeler et al. 1972, Pell 1984, and Struik 1985b)

Fig. 4-1 Regional geological map of the Cariboo Mountains and adjacent Rocky Mountains and Quesnel Highlands.

metamorphism in the southeastern Canadian Cordillera result primarily from early Jurassic to Eocene (oblique) convergence between North America and the oceanic terranes (Monger et al. 1982). 250 to 400 km of shortening between North America and Terrane I is recorded in structures that formed during this interval (Brown et al. 1984, and in press; Fig. 4-2).

Structural and stratigraphic control has been established across the orogenic belt from autochthonous North America, in the Rocky Mountain Foreland, to the eastern edge of Terrane I, in the Quesnel Highlands (Campbell et al. 1982, Brown et al., in press; Fig. 4-1). The geological relationships are summarized in cross section A-A' (Fig. 4-2). This section incorporates surface data projected from north and south of the line of section and has been balanced at deep structural levels using a structural style inferred from observations recorded in the Monashee Mountains where deeper structural levels are exposed (Journeay 1981, 1983, 1985).

In the Quesnel Highlands, a sequence of serpentized ultramafic rocks; mid-Triassic to Early Jurassic (Sinemurian) alkalic and calc-alkalic basalt, volcanoclastic rocks, and argillite of the composite Terrane I (1, Fig. 4-2; Quesnel terrane, Struik 1984) structurally overlies quartzite, meta-pelite, amphibolite, meta-siltstone, meta-volcanic, and marble of the upper Proterozoic (?) to upper Paleozoic Snowshoe Group (2, Fig. 4-2; Barkerville terrane of Struik 1984, Kootenay terrane of Monger 1984, Price et al. 1985) along the eastwardly verging Quesnel Lake Shear Zone (Eureka thrust of Struik 1985a,c; 3, Fig. 4-2). The Snowshoe Group is intruded by mid-Paleozoic plutons (Quesnel Lake Gneiss, 350 - 380 Ma., J. Mortensen, personal

Fig. 4-2 Cross section A - A' (location indicated in Fig. 4-1) illustrating structural relationships from cratonic North America to within Terrane I. Ornamentation and symbols as in Fig. 4-1. Numbers refer to locations discussed in text. Rocky Mountain portion of the cross section is modified from Mountjoy (1980) and Campbell et al. (1982). Western Cariboo Mountains part of the section is from Rees (1981) and Rees and Ferri (1983), and modified from Struik (1980, 1982a, 1983a, and 1983b).

Fig. 4-2 Cross section A - A' (location indicated in Fig. 4-1) illustrating structural relationships from cratonic North America to within Terrane 1. Ornamentation and symbols as in Fig. 4-1. Numbers refer to locations discussed in text.

communication 1985) and records an episode of deformation that occurred before the obduction of Terrane I (Montgomery 1985, Filipone 1985). The Snowshoe Group has been correlated with rocks of the North American continental margin sequence (Campbell et al. 1973, Struik 1980, 1981), but because the continental margin rocks to the east at this latitude do not record evidence of Paleozoic plutonism and pre-early Jurassic deformation, the Snowshoe Group is considered to be "suspect" (Monger 1984, Price et al. 1985).

In the Black Stuart Synclinorium (Figs. 4-1 and 4-2), Terrane I also comprises a sequence of upper Paleozoic basalt, fragmental volcanics, and ribbon chert (Slide Mountain terrane, Struik 1984, 1985a, b). Slide Mountain terrane structurally overlies the Snowshoe Group and rocks which are clearly part of the North American continental margin sequence (Black Stuart and Cariboo Groups) along the Pundata thrust (Struik 1980). Although their hanging walls are dissimilar and their surface traces are separated by younger faults, the Pundata and Eureka - Quesnel Lake thrusts are interpreted to be part of the same fault system. The juxtaposition of Quesnel terrane and Slide Mountain terrane in the same hanging wall can be explained by a history of pre-emplacement thrust faulting and erosional truncation similar to what has been documented for Terrane I in northern British Columbia (Harms 1985) and in southern British Columbia (Klepacki & Wheeler 1985).

Terrane I, the continental margin sequence as far east as the southern Rocky Mountain Trench (6, Fig. 4-2; Fig. 4-1), and the contact between them are deformed by syn-metamorphic southwestwardly verging folds and faults (Campbell 1971; Campbell et al. 1973; Pigage 1978;

Struik 1980, 1981; Rees 1981; Murphy & Journeay 1982; Getsinger 1982; Murphy & Rees 1983; Rees & Ferri 1983; Engi 1984; Fillipone 1985; Montgomery 1985). Displacement on shallow-level thrust faults (such as the Pleasant Valley Fault of Struik 1982a; 4, Fig. 4-2) in the Quesnel Highlands is thought to be absorbed at deeper structural levels to the east in zones of ductile thickening (5, Fig. 4-2) characterized by gently inclined folds which duplicate stratigraphy across their axial surfaces (Murphy & Journeay 1982; Chapter 3). On the basis of the amount and distribution of structural thickening, southwestwardly verging structures are interpreted to represent 50 to 65 km of shortening between Terrane I and North America (Chapter 3). When this displacement is restored, the Quesnel Lake - Eureka portion of the fault system is separated from the Pundata thrust by at least 75 km and potential hanging wall cut-offs lay at least 90 km to the east from potential footwall cut-offs.

First-appearance mineral isograds are superimposed across syn-metamorphic southwestwardly verging folds of the rocks of the continental margin sequence (Chapter 3, Pell 1984, Engi 1984).

Isograds, in turn, are cross-cut by a 163 Ma (Rb/Sr whole-rock, Pigage 1977) granodiorite pluton. Thus, between the Sinemurian (the age of the youngest sediments of Terrane I) and 163 Ma, approximately 150 km of shortening between Terrane I and North America occurred. Shortening was initially accommodated on the eastwardly verging Quesnel Lake Shear Zone, and later by southwestwardly verging structures in the central and outer parts of the continental margin sequence.

Southwestwardly verging structures at the deep structural levels.

exposed in the east-central Cariboo Mountains (Fig. 4-1) are deformed by late- to post-metamorphic, northeastwardly verging folds (Murphy & Journeay 1982, Murphy & Rees 1983, Chapter 3). Although the age of these later structures is not well constrained along section A-A', they are interpreted to be correlative with northeastwardly verging deep-level shear zones such as the Monashee Decollement (Read & Brown 1981), which is locally cut by a 157 Ma. pluton (Galena Bay Stock; U/Pb zircon, Armstrong 1983). Late- to post-metamorphic eastwardly verging structures are present in the Rocky Mountain Main Ranges (Fig. 4-1; McDonough & Simony 1984, Leonard 1984, Mountjoy et al. 1985) suggesting that displacements on deeper-level shear zones such as the Monashee Decollement were transmitted eastward into the Rocky Mountain Foreland.

The Rocky Mountain Foreland (7, Fig. 4-2) is underlain by rocks of the inner portion of the North American continental margin prism and a syn-orogenic clastic wedge sequence which were both telescoped eastwardly along an interlocking array of post-metamorphic north-eastwardly verging thrust faults (Bally et al. 1966, Dahlstrom 1970, Price & Mountjoy 1970, Campbell et al. 1973, Mountjoy 1980). The age of the earliest faulting in the Rocky Mountain Foreland is not precisely known but may be mid-Jurassic, coeval with motion on Monashee Decollement. The earliest syn-orogenic terrigenous coarse clastic deposits in the Rocky Mountain Foreland are latest Jurassic (Price & Mountjoy 1970). However, the appearance of marine shales and sandstones in the lower Callovian after an extended interval of non-deposition (upper Bajocian and most of the Bathonian, Frebold & Tipper 1970) may represent a transgression induced by lithospheric flexure due to structural thickening at the edge of the craton. Two major pulses

of thrusting and uplift are recorded in the clastic wedge, one late Jurassic - early Cretaceous, and the other, late Cretaceous to Paleocene (Price & Mountjoy 1970). 125 to 150 km of shortening is recorded along section A-A' (Mountjoy 1980, Campbell et al. 1982, Brown et al. 1984 and in press).

Thus, early convergence between Terrane I and North America was accommodated along eastwardly verging faults which rooted outboard of the continental margin prism. Subsequent convergence was accommodated at first along an eastwardly dipping, westwardly verging displacement trajectory across the central and outer portions of the continental margin prism, and finally, by an eastwardly verging displacement trajectory which involved the inner part of the continental margin prism and rooted in the basement to the continental margin prism. The kinematic history of the southeastern Canadian Cordillera, therefore, is characterized by an eastward migration of the boundary of active convergence and reversals in regional structural vergence. Mechanical models to account for this kinematic history must isolate the factors which control the location and orientation of the trajectories of crustal failure. In the following sections, the continental margin of this portion of the Cordillera is restored to its structural geometry at the times when eastward migration and vergence reversal occurred. The mechanical properties of these palinspastic restorations are then analyzed using strength - depth profiles.

(4-4) THE JURASSIC CONTINENTAL MARGIN

(4-4a.) Early Jurassic (pre-Sinemurian): geometry

In the earliest Jurassic, the continental margin of this portion of North America opened westward into a marginal basin in which black shales and distal volcanic sediments accumulated (Wheeler et al. 1972). Bounding the marginal basin on the west was a volcanic arc, which was active from the mid-Triassic until the Early Jurassic (Campbell & Tipper 1972). The volcanic arc has been attributed to eastwardly dipping subduction of an ocean basin which lay to the west of the arc (Monger 1977b, Monger & Price 1979).

Two cross sections of the early Jurassic continental margin, based on two different kinematic models, are shown in Figs. 4-3a and 4-4a. Fig. 4-3a is based on a model in which southwestwardly verging structures are a result of the insertion of a rigid crustal wedge along the base of the continental margin sequence, delaminating it from its underlying basement (Archibald et al. 1983, Price 1983). Fig. 4-4a is based on a model in which southwestwardly verging structures are a result of underthrusting of transitional crust beneath the craton edge (Monger 1977b, Brown 1978, Brown 1981, Brown et al. 1984 and in press).

The dimensions and relative locations of the tectonic elements (arc, marginal basin, continental margin prism) are a consequence of the displacements determined from balancing the supracrustal part of structure section A-A' (Fig. 4-2). The topography of the sea floor above the continental margin prism is based on the necessity of a transition from the shallow marine environment of the lower Jurassic (Frebold & Tipper 1970) of the Rocky Mountain Foreland to a water depth ($D > 5$ km) appropriate for the upper Paleozoic oceanic crust of the marginal basin lying between the continental margin prism and the arc

Figs. 4-3 and 4-4 Scaled reconstructions of the Jurassic continental margin (larger scale versions in pocket)

Fig. 4-3 Model 1

Fig. 4-4 Model 2

Sections were constructed by balancing cross section A - A' using the displacement trajectories indicated in Figs. 4-3d and 4-4d.

Heavy dotted lines represent displacement trajectories for subsequent deformation; medium dashed lines represent abandoned displacement trajectories; light dot-dash lines represent isotherms in degrees Kelvin; medium dot-dash line separating Snowshoe Group from its underlying basement in Figs. 4-3c and 4-4c reflects the lack of data on the nature of basement deformation in this locality.

See text for discussion.

a. early Jurassic (pre-Sinemurian) b. early Jurassic (post Sinemurian) c. mid-Jurassic d. present

Quesnel terrane

Slide Mountain terrane

Snowshoe Group (Barkerville - Kootenay terrane

"non-suspect" (or only slightly so, Chamberlain and Lambert 1985) Proterozoic to Mesozoic rocks of the north American continental margin prism

cratonic and transitional crustal basement of North American continental margin prism

upper mantle lithosphere

MODEL 1

Fig. 4-6

MODEL 2

TERRANE 1

NORTH AMERICA

MARGINAL BASIN

CONTINENTAL MARGIN

VOLCANIC ARC

ARC OBUCTION

"BACK-THRUSTING AND -FOLDING"

PRESENT

Fig. 4-4

of Terrane I. An eastwardly sloping ocean floor on the western part of the basin is implied by the presence of arc-derived clastic sediments in the marginal basin (Campbell & Tipper 1972, C.J. Rees personal communication 1984).

With the exception of its deformed distal toe (Snowshoe Group), the doubly-tapering profile of the continental margin prism is defined by sediment thicknesses compiled from measured stratigraphic sections found along the line of section or projected into the plane of the section (Campbell et al. 1973; Pell 1982, Struik 1983a). The distal taper is defined partly by structural thicknesses of the Snowshoe Group (Struik 1982a,b; 1983a,b; 1985c) and partly by the necessity to "zero-out" Proterozoic rocks at or near the eastern limit of the upper Paleozoic oceanic basement of the allochthon. As the location of the eastern limit of upper Paleozoic basement places constraints on the location of the footwall cut-off of the Quesnel Lake Shear Zone, it was necessary to balance the shape of the taper against minimizing the displacement on the Quesnel Lake - Eureka - Pundata fault system: the shallower the taper, the larger the displacement on the fault system.

Basement is restored beneath the continental margin prism along the fault trajectories shown in Figs. 4-3d and 4-4d. The trajectories associated with eastwardly verging thrust faults feeding into the Rocky Mountain Foreland were determined from the inferred geometry of the basement thrust sheet east of the southern Rocky Mountain Trench and the magnitude of shortening determined from preliminary bed-length restoration of the supracrustal sediment prism. The trajectories of the southwestwardly verging deformation were determined by using the eastern limit of southwestwardly verging deformation as a hanging wall

cut-off and a conservative estimate of displacement of 50 km (Chapter 3).

The composition of the basement beneath the continental margin prism cannot be directly determined. Aphebian basement gneisses exposed in antiformal culminations in the Monashee Mountains are granitic to quartz monzonitic in composition (Morrison 1982, R.L. Brown and J.M. Journeay, personal communications, 1985) but its composition at deeper structural levels may only be inferred from geophysical models. Models of gravity and seismic refraction data suggest a density of 2.9 to $3.0 \times 10^3 \text{ kg m}^{-3}$ (Stacey 1973) which is typical for quartz-poor, intermediate to mafic rocks. The seismic velocity of lower crustal rocks (6.4 - 6.5 km/sec from 10 to 40 km, Chandra & Cumming 1972, Cumming 1977, summarized in Price *et al.* 1985), however, is too slow for mafic rocks (Carmichael 1982). In this analysis, the lower crust is interpreted to have the composition of a diorite. (The rheologies of rocks in the compositional range between quartz diorite and anorthosite are sufficiently similar that the results of this analysis would not be substantially different if another composition in this range were chosen.)

(4-4b.) Early Jurassic (pre-Sinemurian): thermal structure

The thermal structure of the continental margin prism and marginal basin during the Sinemurian is unknown but will be assumed to be analogous to equivalent-age modern rifted margins. The Cordilleran miogeocline records several episodes of crustal attenuation and/or rifting (Monger & Price 1979, Price & Fermor 1981, Price 1981, Pell 1982, 1984, Bond & Kominz 1984, Gabrielse 1985b), but no oceanic crust

(ophiolite) younger than the Mississippian is known (Monger 1977a, Klepacki 1983, Klepacki & Wheeler 1985). If the latest rift - drift transition occurred at the end of the Mississippian, then the Sinemurian continental margin would have been approximately 100 m.y. old.

Geotherms have been measured in deep wells drilled into the continental shelf of eastern Canada (Keen 1979). Average borehole temperature gradients measured in wells drilled into the 175 m.y. Nova Scotia margin and the 75 m.y. Labrador Sea margin are 20 K km^{-1} and 26 K km^{-1} , respectively. According to these data, the distal toe of a 100 m.y. old continental margin prism such as is found in the Canadian Cordillera during the early Jurassic would have an upper crustal geotherm of approximately 22 K km^{-1} . This geotherm would decrease towards the craton to a cratonic upper crustal geotherm, here chosen as 10 K km^{-1} .

The upper crustal geotherm in a marginal basin depends on the presence of active back-arc spreading and on proximity to the arc. 100 m.y. old oceanic crust would be expected to have a geotherm of approximately 13 K km^{-1} (Turcotte & Schubert 1982). However, the geotherm would be significantly higher if either spreading or arc volcanism were nearby. During the Sinemurian, Terrane I was an active arc terrane built on an upper Paleozoic ocean floor. The metamorphic grade of the sediments at the base of the > 12 km thick pile of arc sediments and volcanics was probably sub-greenschist before their emplacement (C. J. Rees, personal communication), suggesting maximum temperatures of 673 K to 723 K and therefore maximum geotherms between

31 and 35 K km⁻¹. An upper crustal geotherm of 30 K km⁻¹ is used in this analysis.

The thermal structure shown in Figs. 4-3a and 4-4a reflects both topography and lateral variations in geotherm. A thermal dome is found in the region of the Terrane I arc, which is topographically as well as thermally prominent.

(4-4c.) Early Jurassic (post-Sinemurian): geometry

Two cross sections of the continental margin following the emplacement of Terrane I, but before the onset of internal deformation of the continental margin by southwestwardly verging structures, are shown in Figs. 4-3b and 4-4b. The Pundata thrust and Eureka - Quesnel Lake Shear Zone are shown as a continuous fault system with a trajectory constrained by the following observations:

- (1) Snowshoe Group rocks are absent in the allochthon. This observation suggests that the décollement entered the continental margin prism above the structural level of the Snowshoe Group, possibly within the uppermost Paleozoic or Triassic sediments.
- (2) Faults in the hanging wall of the Eureka thrust imbricate the Late Triassic/early Jurassic stratigraphy (Struik 1985a). This observation suggests that the fault reached the early Jurassic sea floor, which lay an unknown, but small, distance above the upper Triassic strata.
- (3) The hanging wall contains serpentized ultramafic rocks which have been interpreted to be oceanic basement (Monger 1977a, Montgomery 1978, Rees 1981). This observation implies that the Pundata - Eureka - Quesnel Lake system cuts at a steep angle to the base of the oceanic crust. However, the slivers are quite thin, implying that the fault

has a gently dipping orientation. This paradox can be explained by an earlier episode of thrust faulting along faults which cut at a steep angle to the lower oceanic crust and upper mantle and emplaced serpentinized ultramafic rocks into shallow structural levels. Following uplift, erosional truncation, subsidence, and extrusion of arc volcanics and volcanoclastic sediments, the Pundata - Eureka - Quesnel Lake system was activated, traversing the new geometry at a lower angle than the older fault system (similar to the Whitewater Fault of Klepacki (1983), and possibly early faults described by Harms 1985).

(4) Early Jurassic clastic sediments associated with the emplacement of Terrane I have not been observed. Therefore, obduction is interpreted to have occurred sub-aqueously, accompanied by depression of the lithosphere beneath the continental margin prism.

(4-4d.) Early Jurassic (post-Sinemurian): thermal structure

The emplacement of Terrane I will disturb the equilibrium thermal structure of both Terrane I and the continental margin. The allochthon is emplaced onto significantly cooler rocks and will lose some of its heat to them. The footwall is depressed to deeper structural levels by the weight of the allochthon and will gain heat from the allochthon and the hotter surrounding rocks. The allochthon is detached from its primary heat source (arc volcanism) and would be expected to cool towards the upper crustal geotherm of the continental margin sequence. The rate of cooling is not known; the recent model of Glazner & Bartley (1985, Fig. 2) suggests that thermal re-equilibration after a tectono-thermal perturbation is almost complete within the first 10 m.y.

following the perturbation.

(4-4e.) Middle Jurassic: geometry

Two cross sections of the continental margin following the deformation of the outer and central parts of the sediment prism by southwestwardly verging structures, but before the deformation of the inner part of the prism by northeastwardly verging structures, are shown in Figs. 4-3c and 4-4c. The displacement trajectory of the southwestwardly verging structures cuts across the supracrustal realm diagonally from brittle, shallow structural levels in the west to ductile, deeper structural levels in the east. The fate of this displacement trajectory in the crystalline basement is not known; according to the two kinematic models which have been proposed to account for this deformation, the trajectory either intersects an eastwardly verging shear zone at the base of a delaminating indenter in a geometry similar to a foreland belt "triangle zone" (Model 1, Fig. 4-3c; Archibald et al. 1983, Price 1983) or it continues across the crust to a decollement at the base of the crust so that the crustal basement to the continental margin prism underplates the edge of autochthonous North American continental or transitional crust (Model 2, Fig. 4-4c, Brown et al. 1984 and in press). Fig. 4-3c incorporates geometric constraints imposed by the inferred magnitude of displacement along the southwestwardly verging trajectory; as such, the rigid indenter shown in Fig. 4-3c consists of the crustal basement to the central part of the continental margin prism as well as the oceanic rocks of Terrane I (see Chapter 3).

The suprastructure is interpreted to have become subaerially

exposed during this episode of deformation. This interpretation is based on data from the Bowser Basin, a thick sequence of fluvio-deltaic sediments presently located approximately 500 km northwest of the location of section A - A' (Eisbacher 1974, 1977; Fig. 1-1). Although not presently located near section A-A', the Bowser basin lies west of the continuations of major dextral strike-slip faults which were active in the Cretaceous and Tertiary. These faults may account for a cumulative displacement of over 1000 km (Gabrielse 1985a); displacement of this order of magnitude would mean that the Bowser Basin would have been located to the west of the Quesnel Highlands before strike-slip faulting and would have been intersected by the continuation of section A-A' (Eisbacher 1974).

Mid-Jurassic sediments in the Bowser Basin were derived from a coeval westwardly verging thrust belt which presently lies to the east and northeast of the basin (Eisbacher 1974). As just discussed, this thrust belt, which involves rocks of Terrane I, would have been continuous with the Quesnel Highlands thrust belt during the Jurassic. If so, the Bowser Basin can be interpreted as a foreland basin, receiving sediments from a sub-aerially exposed westwardly verging foreland thrust belt and its associated ductilely/thickened hinterland.

(4-4f.) Middle Jurassic: thermal structure

Shortening of the continental margin sequence and its allochthonous cover results in significant structural thickening, alteration of geotherms, and metamorphism. Petrologic data from mineral assemblages that were quenched after structural thickening, but before 163-Ma, are available from the Cariboo Mountains and Quesnel

Table 1: Summary of geothermobarometric data used to constrain the mid-Jurassic geotherms

1. Western Cariboo Mountains, near contact between Kaza and Snowshoe Groups (Fletcher & Greenwood 1979)

$$\begin{array}{cc} \frac{T \text{ (K)}}{1023 \pm 150} & \frac{P \text{ (MPa)}}{953 \pm 30} \end{array}$$

Average geotherm: 28.5 K km⁻¹

2. Western Cariboo Mountains, near contact between Kaza Group and Snowshoe Group (Engi 1984)

$$\begin{array}{cc} \frac{T \text{ (K)}}{778 \text{ (average)}} & \frac{P \text{ (MPa)}}{500 \text{ (average)}} \end{array}$$

Average geotherm: 29 K km⁻¹

$$\begin{array}{cc} \frac{T \text{ (K)}}{853 \text{ (average)}} & \frac{P \text{ (MPa)}}{500 \text{ (average)}} \end{array}$$

Average geotherm: 34 K km⁻¹

3. Southern Cariboo Mountains, sillimanite and co-existing kyanite and staurolite isograds within Kaza and Horsethief Creek Groups (Pell 1984)

$$\begin{array}{cc} \frac{T \text{ (K)}}{530 \text{ (average)}} & \frac{P \text{ (MPa)}}{470 \text{ (average)}} \end{array}$$

Average geotherm: 31 K km⁻¹

$$\begin{array}{cc} \frac{T \text{ (K)}}{655 \text{ (average)}} & \frac{P \text{ (MPa)}}{660 \text{ (average)}} \end{array}$$

Average geotherm: 28 K km⁻¹

Highlands (Table 1). Most of the data are from the high grade rocks near the base of the continental margin prism; no data exists for regions to the east or west. All of the petrologic data suggest that geotherms in the range of 28 to 35 K km⁻¹ were present in the crust affected by the southwestwardly verging structures. Because of uncertainty about the relative rates of uplift and upward relaxation of isotherms, these geotherms are considered to be minimum estimates.

The eastern limit of structural thickening by southwestwardly verging structures is presently at or near the southern Rocky Mountain Trench (Fig. 4-1, 4-2). East of this limit, the isotherms should only be slightly disturbed by the deformation to the west. On the basis of the stable mineral assemblage of paragonite - chloritoid - carbonate, Carey (1984) estimates temperatures between 573 K and 673 K, and pressures between 400 MPa and 500 MPa for low grade rocks in the Rocky Mountain Main Ranges. This mineral assemblage probably quenched at a later time than those upon which Table 1 is based, but it provides some evidence of disturbance of the geotherm in the Rocky Mountain Main Ranges.

A thermal saddle separates the ductilely thickened infrastructure from its foreland area to the west. This saddle is primarily an artifact of the distribution of structural thickening shown in the cross section. The amount of structural thickening in the area of the thermal saddle is poorly constrained and may actually be larger than shown. An increase in the amount of ductile thickening in this area would reduce the amplitude of this thermal saddle.

(4-5) MECHANICAL PROPERTIES OF THE JURASSIC CONTINENTAL MARGIN

(4-5a.) Analytical method

In this section, the restorations of the Jurassic continental margin are analyzed with the use of strength vs. depth profiles. Profiles are constructed at several places along the restored cross sections and correlated laterally to produce a two-dimensional model of the margin's rheological structure. The reliability of the rheological models is then tested by comparing the predicted variations of mechanical behavior along displacement trajectories with the observed variations. Finally, the models are used to evaluate the importance of the rheological structure on the kinematic evolution of this part of the Canadian Cordillera.

(4-5b.) Calculations of lithospheric strength

Strength vs. depth profiles show the variations of the experimentally predicted yield strength (brittle or ductile) present in a lithologic column with a given geotherm, pore pressure, and strain rate. The profiles are constructed by calculating the brittle and ductile (creep) strengths of the rocks in the lithologic column and plotting the lower of the two values against the depths at which the rocks occur. If the brittle strength is lower, the rock would be expected to fail in a brittle manner and its strength is the maximum stress difference that the rock could support at the point of fracture. If the ductile strength is lower, the rock would be expected to fail in a ductile manner and its strength is the stress difference necessary to achieve a given strain rate.

Brittle strengths are calculated with the expression for linear frictional failure, written in terms of stress difference, overburden pressure, and pore fluid pressure (Sibson 1974)

$$(\sigma_1 - \sigma_3) \geq \beta \rho g z (1 - \lambda)$$

where σ_1 and σ_3 are the maximum and minimum principal compressive stresses; β is a factor with the value 3, 0.75, and 1.2 for thrust, normal and strike-slip faults, respectively; ρ is the rock density, g is the acceleration of gravity, z is depth, and λ is the pore fluid factor (ratio of pore fluid pressure to overburden pressure). In this analysis, brittle strengths are calculated using a rock density of $2.75 \times 10^{-3} \text{ kg m}^{-3}$; hydrostatic pore pressure ($\lambda = 0.36$), the general condition assumed for most of the upper crust with the exception of sedimentary basins (Brace & Kohlstedt 1980, Vink et al. 1984); and $\beta = 3.0$. These strengths are considered to be maximum values due to the expected presence of weak horizons (locally overpressured zones ($\lambda > 0.36$) or clay-rich horizons) which control displacement trajectories.

When ductilely deformed in the laboratory, practically all crustal and upper mantle rocks follow a flow law of the type (Kirby 1983)

$$\dot{\epsilon} = A(\sigma_1 - \sigma_3)^n \exp(-Q/RT)$$

where σ_1 and σ_3 are the maximum and minimum principal compressive stresses, respectively; $\dot{\epsilon}$ is the strain rate; A , n , and Q are approximately stress- and temperature-independent (for crustal and upper mantle temperatures) material parameters; R , the gas constant; T , the temperature (K). To calculate ductile strengths, this expression is re-written as

$$(\sigma_1 - \sigma_3) = \left(\frac{\dot{\epsilon}}{A} \right)^{1/n} \exp(Q/nRT)$$

and solved for the stress difference.

The values of A, n, and Q used in the calculations are determined by the rock-type and the temperature is determined directly from the reconstructed geotherms. It is impossible to precisely determine the strain rate; indeed, it is likely that the strain rate is both laterally and vertically variable due to variations in the volume of rock in which the initial displacement is distributed. However, a range of average strain rates can be estimated from the average velocity of convergence between North America and Terrane I. 120 to 175 km of displacement between North America and Terrane I occurred between approximately 187 Ma and 163 Ma, yielding an average velocity of 1.5 to 2.3 x 10⁻¹³ km sec⁻¹. If displacement is initially distributed over a shear zone thickness of 0.5 km, these velocities represent a shear strain rate of 3.0 to 4.6 x 10⁻¹³; if the shear zone is 10 km-thick the shear strain rate is 1.5 to 2.3 x 10⁻¹⁴. (These strain rates apply only in the case of a through-going simple shear zone; if the initial displacement feeds into a blind zone, the strain rate will vary from 0 upwards to the value of the shear strain rate.) Ductile strengths are calculated for strain rates ranging from 10⁻¹³ to 10⁻¹⁵ s⁻¹.

Experimental data are available for a limited number of rock-types; as such, where data are unavailable for a specific rock-type, a close approximation is used. Terrane I is composed primarily of basaltic volcanics and volcanic sediments and an underlying gabbroic oceanic crustal basement. However, data are available only for

compositionally equivalent, but texturally dissimilar, diabase. As a consequence, Terrane I, from the surface downward to the base of its oceanic crust, is modelled with the rheology of diabase. The North American continental margin sequence is composed of a variety of sedimentary rocks but at the scale of this analysis, can be divided into a basal unit, whose initial rheology is probably controlled by coarse-grained quartzo-feldspathic clastic rocks (Windermere "grit" unit) and an overlying mixed unit whose initial rheology is probably controlled by quartz-rich clastic rocks. To account for the quartz-rich nature of the supracrustal sediments, an appropriate model rheology would be that of "wet" (containing variable amounts of H_2O) quartz; however, to account for the feldspar-rich basal unit, the supracrustal sediments are modelled with the rheology of a "wet" granite. The rheologies of "wet" quartz and granite are sufficiently similar that the conclusions would not be substantially different if the supracrustal sediments were modelled with a two-layer quartz-over-granite model rather than the single-layer "wet" granite. Due to the lack of data on diorites, the basement to the continental margin sequence is modelled with an anorthosite rheology. Experimental data on diorites are rare; however, plagioclase-rich rocks have similar creep parameters and, as long as the mafic minerals in the diorite do not interlock into a framework, a diorite should behave similarly to an anorthosite. The upper mantle underlying the whole region is modelled with the rheology of olivine. The creep parameters of the model rock-types are summarized in Table 2.

Creep parameters are often only approximately reproducible; as such, the resulting strengths are considered to be order-of-magnitude

Table 2: Creep parameters for rock types used in analysis
(from Kirby 1983)

	<u>Terrane I</u>		
	<u>A</u>	<u>n</u>	<u>Q</u>
supracrustal sediments and volcanics (basaltic average composition; modelled as diabase)	$3.2 \cdot 10^6$	3.4	260
crustal basement (basaltic average composition)		as above	
upper mantle (ultramafic composition; modelled as olivine)	$4.0 \cdot 10^{15}$	3.0	540
<u>North America</u>			
supracrustal sediments (modelled as "wet granite")	10^2	1.9	137
crustal basement (intermediate average composition; modelled as anorthosite)	$1.3 \cdot 10^6$	3.2	238
upper mantle (ultramafic composition; modelled as olivine)	$4.0 \cdot 10^{15}$	3.0	540

estimates. As well, Newtonian flow (strain rate linearly proportional to the stress) and various softening mechanisms (pressure solution, recrystallization, reaction softening, and shear heating) may pertain under certain conditions, so the calculated strengths for a given strain rate are considered to be maximum values.

(4-5b.) Results

The evolution of the rheological structure of the continental margin during the Jurassic is shown in Figs. 4-5, 4-6, 4-7, 4-8, 4-9, and 4-10 (small-scale versions in text; larger scale versions in pocket):

(4-5b.-1) Early Jurassic (pre-Sinemurian, Fig. 4-5 and 4-6).

With the thermal structures shown in Figs. 4-3 and 4-4 and the range of strain rates under consideration, the thick central part of the continental margin prism is composed of a lithospheric multilayer of strong (brittle) and weak (ductile) layers. The multilayer is bounded on the west by a 100 km wide region of 20 - 30 km-thick brittle transitional and oceanic lithosphere and on the east by 40 - 55 km-thick brittle transitional and continental lithosphere. The lower limit of brittle behavior rises to the west from 60 km beneath the craton to 32 km at the edge of transitional crust.

Due to the elevated geotherms of the arc terrane, a single brittle - ductile transition exists at the base of the volcanic and sediment piles. To the east, and lower geotherms, a wedge of brittle oceanic mantle separates a thin wedge of ductile oceanic lower crust from ductile oceanic upper mantle. Further to the east, the oceanic lithosphere is brittle to approximately 25 km.

Figs. 4-5 through 4-10 Rheological evolution of the Jurassic continental margin (larger-scale versions are found in pocket)

Fig. 4-5 Early Jurassic (pre-Sinemurian); Model 1

Fig. 4-6 Early Jurassic (pre-Sinemurian); Model 2

Fig. 4-7 Early Jurassic (post-Sinemurian); Model 1

Fig. 4-8 Early Jurassic (post-Sinemurian); Model 2

Fig. 4-9 Middle Jurassic; Model 1

Fig. 4-10 Middle Jurassic; Model 2

- a. Distribution of brittle and ductile regions: B, brittle; D, ductile; M, Moho; light dotted line at boundary between brittle and ductile realms is location of brittle - ductile transition for strain rate of 10^{-14} s^{-1} ; heavy dotted lines are displacement trajectories for subsequent deformation. Section is balanced and to scale.
- b. Representative curves of lithospheric strength vs. depth used in the construction of cross section shown in a. The area under the curves ($A = \int_0^A \sigma dz$) is the strength of the lithospheric column. The ductile strength curves are for a strain rate of 10^{-14} s^{-1} . The larger scale versions in the pocket show the variations in ductile strength for the range of strain rates from 10^{-13} to 10^{-15} s^{-1} (stippled areas).

See text for discussion.

RHEOLOGICAL STRUCTURE: EARLY JURASSIC (pre-Sinemurian) (MODEL 1)

Fig. 4-5

RHEOLOGICAL STRUCTURE: EARLY JURASSIC (pre-Sinemurian) (MODEL 2)

Fig. 4-6

RHEOLOGICAL STRUCTURE: EARLY JURASSIC (post-Sinemurian)

(MODEL 1)

Fig. 4-7

RHEOLOGICAL STRUCTURE: EARLY JURASSIC (post-Sinemurian) (MODEL 2)

Fig. 4-8

RHEOLOGICAL STRUCTURE: MIDDLE JURASSIC (MODEL 1)

Fig. 4-9

RHEOLOGICAL STRUCTURE: MIDDLE JURASSIC (MODEL 2)

Fig. 10

(4-5b.-2) Early Jurassic (post-Sinemurian, Figs. 4-7 and 4-8)

Heating of the outer part of the continental margin prism causes the multilayer of brittle and ductile zones beneath the thick central part of the continental margin prism (Figs. 4-5 and 4-6) to extend westward to beneath the leading edge of the allochthon (Figs. 4-7 and 4-8). The two ductile layers of the multilayer merge westward at the expense of the intervening brittle layer in the upper part of the transitional crustal basement of the continental margin prism. This new "rheological stratigraphy" has a significantly lower strength than the pre-emplacement rheological structure.

(4-5b.-3) Middle Jurassic (Figs. 4-9 and 4-10)

With the thermal structure shown in Figs. 4-3 and 4-4, the deformed part of the continental margin is composed of a 10 km thick upper brittle layer and an underlying ductile middle and lower crust and upper mantle. A thin wedge of brittle dioritic basement intrudes into the ductile material at the cratonward edge of the orogen; this wedge enlarges to the east to a point where the cross section is brittle from the surface to a depth of approximately 33 km. The strength of the hot, deformed lithosphere is significantly lower than its pre-D₂ strength.

(4-6) COMPARISON OF MODEL RHEOLOGICAL STRATIGRAPHY WITH OBSERVED PATTERNS OF DEFORMATION

The reliability of the rheological models can be determined by comparing the variations in structural style observed in the Quesnel Highlands and the Cariboo Mountains with the distribution of brittle

and ductile regions along the restored displacement trajectories (Figs. 4-5 to 4-10):

(4-6a.) Early Jurassic (pre-Sinemurian, Figs. 4-5a and 4-6a)

At the western end of Figs. 4-5a and 4-6a, the trajectory of the Quesnel Lake - Eureka - Pundata thrust system lies within ductile rocks of Terrane I oceanic crust (Location 1, Figs. 4-5a and 4-6a). It crosses a brittle - ductile transition within oceanic crust and then follows a westwardly dipping horizon across the brittle realm to the free surface (Location 2, Figs. 4-5a and 4-6a). Thus, in the hanging wall immediately above the fault, brittle structures would be expected to occur from the hanging wall cut-off westward along the fault for approximately 70 km, and ductile fabrics would be expected to occur along the remainder of the fault trajectory. Footwall rocks immediately below the fault would be expected to be brittly deformed between the hanging wall cut-off and the original brittle - ductile transition.

According to Struik (1985a,c), the Pundata thrust is a discrete fault, separating brittle rocks. In contrast, the Eureka thrust and the Quesnel Lake Shear Zone are marked by mylonitic and cataclastic hanging wall and footwall rocks (Struik 1985a,b; Rees & Ferri 1983). Within a body of Quesnel Lake Gneiss deformed in the Quesnel Lake Shear Zone, potassium feldspar megacrysts in the part of the gneiss which restores furthest to the east deform by cataclasis and fracturing (C.J. Rees, personal communication). In the part of the gneiss which restores further to the west, potassium feldspar megacrysts are internally strained and partly dynamically recrystallized (Rees & Ferri

1983). Struik (1985a, p. 270) reports that Snowshoe Group meta-sediments are "invariably sheared at the contact (with the allochthon), from mylonite to variably cataclastic."

The presence of mylonitic and otherwise ductily deformed footwall rocks is not predicted by the model. This may indicate that the model upper crustal geotherm for this area ($20-25 \text{ K km}^{-1}$) is too low. The distribution of brittle and ductile regions is sensitive to changes in geotherm; an increase of only 5 K km^{-1} is sufficient to induce ductile behaviour in the footwall. Alternatively, failure may have been initially brittle, but heating of the footwall by shear heating or rapid heat transfer from the hanging wall during its depression to deeper structural levels may have induced ductile failure. With instantaneous heat transfer, depression of the footwall approximately 8 km would be sufficient for ductile behaviour to occur.

(4-6b.) Early Jurassic (post-Sinemurian, Figs. 4-7a and 4-8a)

From east to west within the supracrustal rocks, the trajectory of D₂ structures follows the weak ductile layer at the base of the thick central part of the continental margin sequence (Location 1, Figs. 4-7a and 4-8a), crosses a brittle - ductile transition in the distal continental margin sediments immediately below the allochthon (Location 2, Figs. 4-7a and 4-8a), merges with the Quesnel Lake - Eureka - Pundata thrust trajectory and then crosses the brittle realm of the allochthon to the free surface (Location 3, Figs. 4-7a and 4-8a). In Model 1, the supracrustal trajectory intersects a westwardly dipping, eastwardly verging shear zone that crosses the brittle upper basement and "roots" in the ductile lower crust (Location 4, Fig. 4-7a). In

Model 2, the D_2 displacement trajectory passes into brittle basement rocks and then "roots" in a ductile layer at the base of the crust (Location 4, Fig. 4-8a). The fate of these trajectories at depth to the east or west is not known, but, ultimately, they must root in zones of plate consumption (subduction zone) or formation (spreading center).

The predicted transition from brittle to ductile behaviour in the continental margin sediments beneath the allochthon corresponds quite closely with field observations. In the Quesnel Highlands, structurally low thrust faults such as the Pleasant Valley Fault separate thrust sheets exhibiting mesoscopic fabrics which record significant ductile strain (Struik 1980, Klepacki 1981). In one location, the Pleasant Valley Fault is tightly folded (Mt. Watt, Struik 1983). Both of these observations suggest a location near a brittle - ductile transition.

The relatively young and structurally high rocks involved in thrusting in the Quesnel Highlands are only weakly deformed in the Cariboe Mountains to the east. Here, older and structurally deeper rocks are folded into a structural sequence that is significantly thicker than the original stratigraphic thickness (Murphy 1985 and Chapter 3). Thickening occurs primarily at deeper structural levels by duplication of stratigraphy across gently dipping axial surfaces; D_2 faults or discrete shear zones are conspicuously absent in this area.

According to Figs. 4-7a and 4-8a, all of Terrane I is expected to fail in a brittle manner. In the Black Stuart Synclinorium, the allochthon is not highly deformed (Struik 1985b). Thrust faults in the underlying continental margin sediments merge into a roof thrust at the

base of the allochthon, suggesting that the allochthon in this area was strong enough to guide stresses to a weaker location. In the footwall of the Pleasant Valley Fault, however, the allochthon, its basal decollement, and the underlying continental margin sediments are folded into tight to isoclinal D_2 folds (Rees & Ferri 1983). The inconsistency between the model and the observed geology is possibly due to heating of the allochthon during deformation.

(4-6c.) Middle Jurassic (Fig. 4-9a and 4-10a):

The fault trajectory associated with the initial mid-Jurassic northeastwardly verging displacement enters ductile crystalline crust at a point directly beneath the eastern edge of the obducted and deformed allochthon (Location 1, Figs. 4-9a and 4-10a). It climbs upward to the east across ductile transitional crust and follows the unconformity at the base of the deformed continental margin sediment prism until it reaches a ramp into a higher level in the deformed sediments (Location 2, Figs. 4-9a and 4-10a). It remains at this level as it passes eastward from ductile, deformed continental margin sediments to ductile, undeformed continental margin sediments (Location 3, Figs. 4-9a and 4-10a), crosses a brittle - ductile transition (Location 4, Figs. 4-9a and 4-10a), and finally reaches the free surface along a poorly constrained ramp - flat trajectory.

A transition similar to the one at Location 3 (Figs. 4-9a and 4-10a) occurs across the Southern Rocky Mountain Trench northwest of Valemount. Ductilely deformed rocks exhibiting early southwestwardly verging (D_2) and later northeastwardly verging (D_3) fabrics occur in the Cariboo Mountains west of the Trench; east of the Trench in the

Rocky Mountains, equivalent age rocks are also ductilely deformed, but record only the later northeastwardly verging fabrics. Further to the east in the Rocky Mountains, ductile deformation is only locally present, suggesting that a lateral brittle - ductile transition such as is shown at Location 4 (Figs. 4-9a and 4-10a) also occurs.

(4-7) IMPLICATIONS OF THE RHEOLOGICAL STRUCTURE OF THE LITHOSPHERE FOR THE KINEMATIC EVOLUTION OF THE SOUTHEASTERN CANADIAN CORDILLERA

The correlation between model predictions and geological observations is sufficiently strong as to warrant the following first-order conclusions about the relationship between the rheological structure of the lithosphere and its deformation:

(1) The trajectories of lithospheric failure pass from deep structural levels to the free surface across lithosphere that is weak relative to surrounding lithosphere. (Compare Profile A with Profiles B, C, D, and E in Figs. 4-5b and 4-6b; Profiles B, C, and D with Profiles A and E in Figs. 4-7b and 4-8b; Profiles B, C, and D with Profiles A and E in Figs. 4-9b and 4-10b.)

(2) Rheological stratification of the lithosphere exerts a profound control on displacement trajectories as they pass from depth to the free surface. Displacement trajectories at deep structural levels follow weak ductile horizons and boundaries between ductile rocks of contrasting strengths, such as the Moho, and cross strong brittle layers in such a way that the contribution of the strength of the brittle layers to the overall strength of the trajectory is minimized. Displacement trajectories appear to follow a ramp - flat profile across

a rheologically stratified lithosphere in much the same way as brittle thrusts at shallow structural levels cross a stratigraphic multilayer.

(3) The strength of lithospheric columns changes during their tectono-thermal evolution. Changes in strength are large enough to induce migration of the locus of lithospheric failure into newly weakened lithosphere. In this regard, the eastward migration of the active boundary between North America and Terrane I is explained as a consequence of thermal weakening during overthrusting and regional metamorphism.

Figure 4-11 compares the strengths of abandoned displacement trajectories with the strengths of the new trajectories at the times when re-locations occur. (The strength profiles shown in Fig. 4-11 are measured along displacement trajectories rather than vertically and assume that the displacement trajectories initiated simultaneously along their whole lengths. In view of the common occurrence of blind thrusts, back thrusts, and zones of ductile thickening, this assumption is not strictly valid. As such, the comparison is purely qualitative.) In both cases, the new displacement trajectory has a lower strength than the the abandoned one; in this regard, the eastward migration of the active boundary of convergence between North America and Terrane I is explained as a consequence of thermal weakening of the lithosphere in regions during overthrusting and regional metamorphism.

(4) The models suggest a relationship between surface topography and the orientation (and therefore vergence) of displacement trajectories. The west-dipping, east-verging trajectory of the Quesnel Lake - Eureka-

Fig. 4-11 Comparison of strengths of abandoned displacement trajectories with strengths of new displacement trajectories at the time of changeover. Strengths are considered to be a maximum because it is unlikely that the whole displacement trajectory will be active simultaneously.

Left: Comparison of the strength of trajectory of Quesnel Lake - Eureka - Pundata thrust system at the time of its abandonment (open circles), with the strength of new displacement trajectories (closed circles; DL = Model 1, UT = Model 2).

Right: Comparison of the strength of Model 1 and Model 2 displacement trajectories at the time of abandonment (closed circles; DL = Model 1, UT = Model 2), with the strength of 1 (open diamonds), the earliest northeastwardly verging displacement trajectory which feeds into the Rocky Mountain Foreland (near location 1, Figs. 4-9 and 4-10)

Pundata thrust system formed beneath the east-dipping surface slope of Terrane I. A west-dipping surface overlies much of the west-verging D_2 displacement trajectory as it passes through the continental margin prism. Where it crosses the allochthon, the surface is sub-horizontal. The trajectory of middle Jurassic east-verging deformation underlies a broad east-dipping surface slope; however, at this time, clastic detritus is eroding from the orogenic belt and is being transported to the west into the Bowser Basin suggesting that both east- and west-dipping surface slopes may have been present.

This relationship may have a dynamic basis: it is possible that the unidirectional horizontal component of gravitational force created by the topographic slope would augment the horizontal compressive tectonic "push" such that the resolved shear stress on a surface (or zone) dipping in the opposite direction to the surface slope would be higher than on a surface (or zone) dipping in a direction parallel to the surface slope. If so, failure would be more likely to occur on a west-dipping surface than on an east-dipping surface.

(5) Integration of (3) and (4) yields the following preliminary mechanical model of the tectonic evolution of this part of the southeastern Canadian Cordillera. Shortening between Terrane I and North America initiated in the weak lithosphere beneath Terrane I due to the elevated arc geotherms in the early Jurassic. Displacement on the Quesnel Lake - Eureka - Pundata thrust system continued until the allochthon encountered the west-dipping surface slope above the continental margin prism. The stress necessary to sustain motion on the original decollement exceeded the yield strength of a displacement.

trajectory across the subjacent lithosphere which became thermally weakened during overthrusting of Terrane I. The eastward dip of the supracrustal portion of the new displacement trajectory may have been due to the dynamic influence of the overlying west-dipping surface slope. Continued shortening was accommodated on the east-dipping trajectory until the stress necessary to sustain motion exceeded the yield strength of a displacement trajectory which traversed the lithosphere most strongly affected by regional metamorphism. The westward dip of the new trajectory may have been a consequence of the dynamic influence of the surface slope, although this relationship can not be clearly demonstrated.

(6). The rheological models do not give sufficient insight into the continuation of the D_2 supracrustal displacement trajectory into the crustal basement and to thereby discriminate between the two tectonic models which have been proposed to account for the southwestwardly vergence of D_2 deformation. When the experimental uncertainties of the empirical flow laws are considered, the crustal portions of the displacement trajectories of each model have approximately the same strength (Fig. 4-11). The displacement trajectory of Model 2 must cross brittle, strong upper mantle beneath the craton before it can root in zones of mantle flow; in contrast, the displacement trajectory of Model 1 accesses weak upper mantle just a short distance to the west, beneath the deformed allochthon. As such, it appears that the weaker Model 1 displacement trajectory will fail at a lower stress than the displacement trajectory of Model 2.

(4-8) DISCUSSION

The rheological models in this paper are a preliminary attempt at providing a mechanical rationale for the kinematic evolution of a part of the southeastern Canadian Cordillera. In spite of the various uncertainties (geotherms, composition of the lower crust, nature and geometry of strain at deep structural levels) and the imprecise nature of the empirical flow laws, the models correspond quite well to the observed and inferred geology summarized on the palinspastic reconstructions.

The scaled models affirm the general conclusion of Armstrong & Dick (1974) that brittle - ductile transitions control the location of decollements during the deformation of crystalline basement rocks. However, temperature-independent (except at very high temperatures where the strength of most rock types are all equally small) rheological transitions between rock types with significantly different ductile failure strengths, such as is found at the Moho, are also likely to be factors in determining displacement trajectories. Rheological stratification is a function of both temperature and rock type.

The models do not attempt to incorporate the effects of erosion. The geological evidence suggests that uplift of the orogenic belt through sea level commenced at or near the end of D_2 deformation. As such, erosion would not have any effect on the early history of convergence. During the shortening in the Rocky Mountain Fold and Thrust Belt, erosion of thickened radioactive crust may have retarded relaxation of the isotherms in such a way that the newly thickened

crust would be initially stronger after overthrusting (Glazner & Bartley (1985). Hardening of this sort may provide the impetus for back-stepping of displacements into regions of the lithosphere that are weak relative to the thickened toe region. Erosion would also more directly affect the kinematic evolution of the thrust belt, through its influence on the angle of taper of thrust sheets (Elliott 1976, Chapple 1978, Davis et al. 1983, Willemin 1984).

Rigorous thermal models are also not incorporated in this analysis. Instead the models appeal to geological and petrological evidence which suggests that geotherms in the central part of the continental margin sequence rose from pre-kinematic (early Jurassic, pre-Sinemurian) values ranging from 10 to 25 K km⁻¹ to post-D₂ values of 20 to 35 K km⁻¹. The source of this elevated heat flow is not known; two possible sources are upward heat transport by magmatic intrusion, and an increase in the radioactive heat production due to structural thickening of the feldspar-rich (and therefore radioactive) clastic unit near the base of the Windermere sequence.

The conclusion of this paper that Cordilleran lithosphere was weaker immediately following both the obduction of Terrane I and the internal deformation and metamorphism of the continental margin prism follows directly from the built-in conditions of sub-aqueous obduction (no erosion) and an increasing geotherm (during regional metamorphism). Accordingly, this conclusion differs from the conclusion of Glazner & Bartley (1985) that approximately 20 Ma are necessary for thermal weakening of structurally thickened (overthrust) lithosphere to occur. Their conclusion is based on a model which incorporates erosion and assumes that isotherms rebound to their pre-kinematic value after

overthrusting. This model clearly does not apply in the southeastern Canadian Cordillera where both high-grade regional metamorphism and ductile deformation occurred within 20 m.y. after the initial emplacement of Terrane I.

Acknowledgements. This paper represents the results of a very stimulating interchange of ideas with Giorgio Ranalli, whose contributions are most gratefully acknowledged. Discussions with Murray Journeay of Queen's University, Chris Rees of Carleton University, and Rambeau Parrish of the Geological Survey of Canada were most helpful in guiding my thinking on the thermal evolution of orogenic belts; however, any misconceptions expressed in this paper are solely the responsibility of the author. This work was supported by grants and research agreements from the Geological Survey of Canada and by Natural Sciences and Engineering Research Council Operating Grant A2693; all to R.L. Brown of Carleton University.

CHAPTER 5

SUMMARY OF CONCLUSIONS

SUMMARY OF CONCLUSIONS

- (1) The Kaza Group of the central Cariboo Mountains has been subdivided into three lithostratigraphic units. One distinctive 200 m-thick unit at the top of the middle Kaza Group appears to be present regionally and forms the basis of regional correlations.
- (2) The 200 m-thick marker at the top of the middle Kaza Group is interpreted to have been deposited in a sub-wave base shelf environment. Its persistence is attributed to regional regression during which shelf facies prograded outward over the locus of turbidite sedimentation. Interbedded shelf, slope, and turbidite facies imply the presence of a shelf - slope - basin configuration during deposition of the "grit" unit.
- (3) The eastward thinning of the "grit" unit suggests the presence of a basin hinge line east of the location of the Cushing Creek area. The difference in subsidence rate across the hinge line may be due to a difference in thickness of the underlying crustal basement caused by attenuation of cratonic thickness basement during Windermere sedimentation.
- (4) A location of the Windermere hinge line to the east of the lower Paleozoic hinge line would account for the discrepancy between the palinspastic location of the lower Paleozoic hinge line and the geophysically determined edge of cratonic thickness basement at the Rocky Mountain Trench.
- (5) The transition from upright to recumbent structures in the Cariboo

Mountains involves syn-metamorphic, but pre-metamorphic peak, southwestwardly verging D_2 structures.

(6) The transition is gradual and continuous and results from a component of shear strain produced by northeastward underthrusting of the North American continental margin prism.

(7) Structural thickening by duplication of stratigraphy across the axial surfaces of tight to isoclinal folds occurs primarily at deep structural levels and represents approximately 75% horizontal shortening. Shortening at shallow structural levels is approximately 10 to 20%; this difference requires that shallow and deep structural levels be detached in regions to the southwest of this portion of the Cariboo Mountains. A southwestwardly verging syn-metamorphic fold and thrust belt near Quesnel Lake is likely to be the shallow level manifestation of the difference in the amount of shortening at different structural levels.

(8) The combination of shear strain and structural thickening represents between 50 and 75 km of northeastward underthrusting of deep structural levels with respect to a fixed suprastructure.

(9) This amount of displacement has been combined with the eastern limit of southwestwardly verging deformation to specifically define the structural geometries implicit in the two different tectonic models which have been proposed to account for backfolds and backthrusts in the southeastern Canadian Cordillera. Both of these geometries are feasible and do not violate any geological or geophysical observations.

With the present data base, it is not possible to unequivocally discriminate between the two models.

(10) The rheological structure of the lithosphere exerts a fundamental control on the location of its failure. The boundaries of active convergence in the Canadian Cordillera traverse lithosphere that is weak relative to adjacent lithosphere.

(11) The strength of lithosphere of the Canadian Cordillera changes during its tectono-thermal evolution. Changes in lithospheric strength due to overthrusting or prograde metamorphism are sufficient to relocate the boundary of active convergence.

(12) The rheological models suggest a relationship between surface topography and the orientation (and therefore vergence) of displacement trajectories. This relationship may have a dynamic basis: it is possible that the unidirectional horizontal component of gravitational force created by the topographic slope would augment the horizontal compressive tectonic "push" such that the resolved shear stress on a surface (or zone) dipping in the opposite direction to the surface slope would be higher than on a surface (or zone) dipping in a direction parallel to that of the surface slope.

(13) Integration of (11) and (12) yields the following preliminary mechanical model of the tectonic evolution of this part of the southeastern Canadian Cordillera. Shortening between Terrane I and North America initiated in the weak lithosphere beneath Terrane I due to the elevated arc geotherms in the early Jurassic. Displacement on the Quesnel Lake - Eureka - Pundata thrust system continued until the

allochthon encountered the west-dipping surface slope above the continental margin prism. The stress necessary to sustain motion on the original decollement exceeded the yield strength of a displacement trajectory across the subjacent lithosphere which became thermally weakened during overthrusting of Terrane I. The eastward dip of the supracrustal portion of the new displacement trajectory may have been due to the dynamic influence of the overlying west-dipping surface slope. Continued shortening was accommodated on the east-dipping trajectory until the stress necessary to sustain motion exceeded the yield strength of a displacement trajectory which traversed the lithosphere most strongly affected by regional metamorphism. The westward dip of the new trajectory may have been a consequence of the dynamic influence of the surface slope, although this relationship can not be clearly demonstrated.

(14) The rheological models do not give sufficient insight into the continuation of the D_2 supracrustal displacement trajectory into the crustal basement and to thereby discriminate between the two tectonic models which have been proposed to account for the southwestwardly vergence of D_2 deformation. When the experimental uncertainties of the empirical flow laws are considered, the crustal portions of the displacement trajectories of each model have approximately the same strength. The displacement trajectory of Model 2 must cross brittle strong upper mantle beneath the craton before it can root in zones of mantle flow; in contrast, the displacement trajectory of Model 1 accesses weak upper mantle a short distance to the west, beneath the deformed allochthon. As such, it appears that the weaker Model 1

displacement trajectory will fail at a lower stress than the displacement trajectory of Model 2.

REFERENCES

- Aitken, J.D., 1969. Documentation of the sub-Cambrian unconformity, Rocky Mountain Main Ranges, Alberta. *Can. J. Earth Sci.* 6: 193-200
- Aitken, J.D. and Mountjoy, E.W., 1963. Early Cambrian and late Precambrian paleocurrents, Banff and Jasper National Parks. *Bull. Can. Petr. Geol.* 11: 161-168
- Archibald, D.A., Glover, J.K., Price, R.A., Farrar, E., and Carmichael, D.M., 1983. Geochronology and tectonic implications of magmatism and metamorphism, southern Kootenay arc and neighboring regions, southeastern British Columbia: Part I. *Can. J. Earth Sci.* 20: 1891-1913
- Armstrong, R.L., 1983. Cordilleran S- and I-type granites: indicators of lithosphere thickness. *Geol. Assoc. Can. Progr. Abstr.* 8: 3
- Armstrong, R.L. and Dick, H.J.B., 1974. A model for the development of thin overthrust sheets of crystalline rock. *Geology* 2: 25-40
- Armstrong, R.L., Eisbacher, G.H., and Evans, P.D., 1982. Age and stratigraphic-tectonic significance of Proterozoic diabase sheets, MacKenzie Mountains, northwestern Canada. *Can. J. Earth Sci.* 19, 316-323
- Bally, A.W., Gordy, P.L., and Stewart, G.A., 1966. Structure, seismic data, and orogenic evolution of the southern Canadian Rocky Mountains. *Bull. Can. Petr. Geol.* 14: 337-381
- Bond, G.C. and Kominz, M.A., 1984. Construction of tectonic subsidence curves for the early Paleozoic miogeocline, southern Canadian Rocky Mountains: Implications for subsidence mechanisms, age of breakup, and crustal thinning. *Bull. Geol. Soc. Am.* 95: 155-173
- Boyer, S.E. and Elliott, D., 1982. Thrust systems. *Bull. Ame. Assoc. Petr. Geol.* 66: 1196-1230
- Brace, W.F. and Kohlstedt, D.L., 1980. Limits on lithospheric stress imposed by laboratory experiments. *J. Geophys. Res.* 85: 6248-6252
- Brown, A., 1968. McBride map-area (93H), British Columbia: Structural Study. *Geol. Surv. Can. Paper* 68-1A: 20-21
- Brown, R.L., 1978. Structural evolution of the southeastern Canadian Cordillera: a new hypothesis. *Tectonophysics* 48: 133-151
- Brown, R.L., 1981. Metamorphic complex of SE Canadian Cordillera and relationship to foreland thrusting. In: Coward, M.P. and McClay, K. (eds.) *Thrust and Nappe Tectonics*. *Geol. Soc. Lond.*

- Brown, R.L., Journeay, J.M., Lane, L.S., Murphy, D.C., and Rees, C.J., 1984. Obduction, backfolding, and piggyback thrusting in the metamorphic hinterland of the southeastern Canadian Cordillera. (abstract) Colloque associe du C. N. R. S. organise avec le concours de la region Midi Pyrenees: Chevauchement et Deformation. p. 11. Toulouse, France (Full paper accepted for publication, Journal of Structural Geology)
- Brown, R.L. and Read, P.B., 1983. Shuswap terrane of British Columbia: a Mesozoic "core complex". *Geology* 11: 164-168
- Brown, R.L. and Tippet, C.R., 1978. The Selkirk fan structure of the southeastern Canadian Cordillera. *Bull. Geol. Soc. Am.* 89: 548-558
- Butler, R.W.H., 1985. Thrust tectonics: a personal view. *Geol. Mag.* 122: 223-232
- Butler, R.W.H. and Coward, M.P., 1984. Geological constraints, deep geology, and structural evolution of the NW Scottish Caledonides. *Tectonics* 3: 347-365
- Campbell, K.V., 1971. Metamorphic petrology and structural geology of the Crooked Lake area, Cariboo Mountains, British Columbia. unpublished Ph.D. thesis, University of Washington, Seattle, Washington, U. S. A. 192 p.
- Campbell, K.V. and Campbell, R.B., 1970. Quesnel Lake map-area, British Columbia (93A). *Geol. Surv. Can. Paper* 70-1A: 32-35
- Campbell, R.B., 1968. Canoe River, British Columbia. *Geol. Surv. Can. Map* 15-1967
- Campbell, R.B., 1970. Structural and metamorphic transitions from infrastructure to suprastructure, Cariboo Mountains, British Columbia. In: *Structure of the Canadian Cordillera* (edited by J. O. Wheeler) *Geol. Assoc. Can. Spec. Paper* 6: 67-72
- Campbell, R.B., 1973. Structural cross-section and tectonic model of the southeastern Canadian Cordillera. *Can. J. Earth Sci.* 10: 1607-1620
- Campbell, R.B. and Tipper, H.W., 1972. Geology of the Bonaparte lake map-area. *Geol. Surv. Can. Mem.* 363, 100 pp.
- Campbell, R.B., Mountjoy, E.W., and Struik, L.C., 1982. Structural cross-section through south central Rocky and Cariboo Mountains to Coast Range. *Geol. Surv. Can. Open File* 844
- Campbell, R.B., Mountjoy, E.W., and Young, F.G. 1973. Geology of McBride map-area, British Columbia. *Geol. Surv. Can. Paper* 72-35
- Carey, A., 1984. Geology of Late Proterozoic Miette Group, Southern Main Ranges, Cushing Creek Area, British Columbia. unpublished

3

MICROCOPY RESOLUTION TEST CHART
NATIONAL BUREAU OF STANDARDS
STANDARD REFERENCE MATERIAL 3010a
(ANSI and ISO TEST CHART No. 2)

- M.Sc. thesis, University of Calgary, Calgary, Alberta, 119 pp.
- Carey, A. and Simony, P.S., 1984. Structure and stratigraphy of the late Proterozoic Miette Group, Cushing Creek area, Rocky Mountains, British Columbia. *Geol. Surv. Can. Paper* 84-1A: 425-428
- Carmichael, R.S., 1982. CRC Handbook of Physical Properties of Rocks CRC Press, Boca Raton, Fla.
- Carpe, A., 1925. Climbs in the Cariboo Mountains and the Northern Gold Range, British Columbia. *The Alpine Journal* 37: 64-67
- Chamberlain, V. and Lambert, R. St.J., 1985. Cordilleria, a newly defined Canadian microcontinent. *Nature* 314: 707-713
- Chamberlin, T.C., 1929. Exploration of the Cariboo Mountains of British Columbia. *Bull. Geog. Soc. Philadelphia* 25: 58-76
- Chandra, N.N. and Cumming, G.L., 1972. Seismic refraction studies in western Canada. *Can. J. Earth Sci.* 9: 1099-1109
- Chapple, W.M., 1978. Mechanics of thin-skinned fold-and-thrust belts. *Bull. Geol. Soc. Am.* 89: 1189-1198
- Coney, P.J., Jones, D.L., and Monger, J.W.H., 1980. Cordilleran suspect terranes. *Nature* 288: 329-333
- Coward, M.P., 1983. Thrust tectonics, thin skinned or thick skinned, and the continuation of thrusts to deep in the crust. *J. Struct. Geol.* 5: 113-123
- Coward, M.P. and Butler, R.W.H., 1985. Thrust tectonics and the deep structure of the Pakistan Himalaya. *Geology* 13: 417-420
- Cumming, W.B., 1977. Crustal structure from a seismic refraction profile across southern British Columbia. Unpublished Ph. D. thesis, University of British Columbia, Vancouver, British Columbia, Canada, 82 pp.
- Dahlstrom, C.D.A., 1970. Structural evidence of the eastern margin of the Canadian Rocky Mountains. *Bull. Can. Petr. Geol.* 18: 332-406
- Davis, D., Suppe, J., and Dahlen, F.A., 1983. Mechanics of fold-and-thrust belts and accretionary wedges. *J. Geophys. Res.* 88: B1153-1172
- Davis, G.A., Monger, J.W.H., and Burchfiel, B.C., 1978. Mesozoic construction of the Cordilleran "collage", central British Columbia to central California. In: Howell, D.G. and McDougall, K.A. (eds.) *Mesozoic Paleogeography of the Western United States*, Soc. Econ. Paleont. Miner. Paleogeography Symposium 2: 1-32
- DeSitter, L.V. and Zwart, H.J., 1960. Tectonic development in supra-

- and infrastructures of a mountain chain. Proceedings of the 21st International Geological Congress 18: 248-256, Copenhagen
- Dickinson, W.R., 1974. Plate tectonics and sedimentation. In: Tectonics and sedimentation (edited by W. R. Dickinson), Society of Economic Paleontologists and Mineralogists Special Publication 22: 1-28
- Duchesne, R.G. and Simony, P.S., 1984. Structural evolution and metamorphism of the southern Cariboo Mountains near Blue River, British Columbia. Geol. Surv. Can. Paper 84-1A: 91-94
- Eisbacher, G.H., 1974. Evolution of successor basins in the Canadian Cordillera. In: Dott, R. and Shaver, J., Modern and ancient geosynclinal sedimentation. Soc. Econ. Paleont. Minér. Spec. Pub. 19: 274-291
- Eisbacher, G.H., 1977. Mesozoic-Tertiary basin models for the Canadian Cordillera and their geological constraints. Can. J. Earth Sci. 14: 2414-2421
- Eisbacher, G.H., 1981. Sedimentary tectonics and glacial record in the Windermere Supergroup, MacKenzie Mountains, northwestern Canada. Geol. Surv. Can. Paper 80-27, 40 p.
- Elliott, D., 1976. The motion of thrust sheets. J. Geophys. Res. 81, B949-963
- Elliott, D. and Johnson, M.R.W., 1980. Structural evolution in the northern part of the Moine thrust belt, NW Scotland. Trans. Roy. Soc. Edin.: Earth Sci. 71: 69-96
- Engi, J., 1984. Structure and metamorphism north of Quesnel Lake and east of Niagara Creek, Cariboo Mountains, British Columbia. unpublished M.Sc. thesis, University of British Columbia, Vancouver, British Columbia, 137 pp.
- Evenchick, C.A., Parrish, R., and Gabrielse, H., 1984. Precambrian gneiss and late Proterozoic sedimentation in north-central British Columbia. Geology 12: 233-237
- Fillipone, J., 1985. Structure and metamorphism at the western margin of the Omineca Belt near Boss Mountain, east-central British Columbia. unpublished M.Sc. thesis, University of British Columbia, Vancouver, British Columbia, 141pp.
- Fletcher, C.J.N. and Greenwood, H.J., 1979. Metamorphism and structure of the Penfold Creek area, near Quesnel Lake, British Columbia. J. Petr. 20: 743-794
- Frebold, H. and Tipper, H.W., 1970. Status of the Jurassic in the Canadian Cordillera of British Columbia, Alberta, and southern Yukon. Can. J. Earth Sci. 7: 1-21

- Fyles, J.T., 1964. Geology of the Duncan Lake area, British Columbia. Brit. Col. Dept. Mines Petr. Res. Bulletin 49, 87 pp.
- Fyles, J.T., 1967. Geology of the Ainsworth-Kaslo area, British Columbia. Brit. Col. Dept. Mines Petr. Res. Bulletin 52, 125 pp.
- Fyles, J.T. and Eastwood, G.E.P., 1962. Geology of the Ferguson area, Lardeau district, British Columbia. Brit. Col. Dept. Mines Petr. Res. Bulletin 45, 92 pp.
- Fyles, J.T. and Hewlett, C.G., 1959. Stratigraphy and structure of the Salmo lead-zinc area. Brit. Col. Dept. Mines Petr. Res. Bulletin 41, 162 pp.
- Gabrielse, H., 1972. Younger Precambrian of the Canadian Cordillera. Am. J. Sci. 272: 521-536
- Gabrielse, H., 1985a. Major dextral transcurrent displacements along the Northern Rocky Mountain Trench and related lineaments in north-central British Columbia. Bull. Geol. Soc. Am. 96: 1-19
- Gabrielse, H., 1985b. Miogeoclinal in the early history of the Canadian Cordillera. Geol. Soc. Am. Prog. Abstr. 17: 355
- Getsinger, J., 1982. Metamorphism and structure of Three Ladies Mountain area, Cariboo Mountains, British Columbia. Geol. Surv. Can. Paper 82-1A: 317-320
- Glazner, A.F. and Bartley, J.M., 1985. Evolution of lithospheric strength after thrusting. Geology 13: 42-45
- Glover, J.K., 1978. Geology of the Summit Creek map-area, southern Kootenay arc, British Columbia: unpublished Ph. D. thesis, Queen's University, Kingston, Ontario, 143 pp.
- Glover, J.K. and Price, R.A., 1976. Stratigraphy and structure of the Windermere Supergroup, southern Kootenay arc, British Columbia. Geol. Surv. Can. Paper 76-1B: 21-23
- Goetze, C. and Evans, B., 1979. Stress and temperature in the bending lithosphere as constrained by experimental rock mechanics. Geophys. J.R. Astr. Soc. 59: 463-478
- Harms, J.C., Southard, J.B., and Walker, J.C., 1982. Structures and sequences in clastic rocks. Soc. Econ. Paleon. Miner. Short Course 9
- Harms, T., 1985. Pre-emplacement thrust faulting in the Sylvestre allochthon, northeast Cry Lake map-area, British Columbia. Geol. Surv. Can. Paper 85-1A: 301-304
- Hillhouse, J.W., 1977. Paleomagnetism of the Triassic Nikolai

- Greenstone, McCarthy, Quadrangle, Alaska. Can. J. Earth Sci. 14: 2578-2592
- Holway, E.W.D., 1917. The Cariboo Mountains. Can. Alp. J. 8: 30-33
- Hoy, T., 1974. Structure and metamorphism of Kootenay arc rocks around Riondel, British Columbia. unpublished Ph. D. thesis, Queen's University, Kingston, Ontario. 201 pp.
- Hoy, T., 1976. Calc-silicate isograds in the Riondel area, southeastern British Columbia. Can. J. Earth Sci. 13: 1093-1104
- Hoy, T., 1977. Stratigraphy and structure of the Kootenay arc in the Riondel area, southeastern British Columbia. Can. J. Earth Sci. 14: 2301-2315
- Irving, E., 1979. Paleopoles and paleolatitudes of North America and speculation about displaced terranes. Can. J. Earth Sci. 16: 669-694
- Irving, E., Monger, J.W.H., and Yole, R.A., 1980. New paleomagnetic evidence for displaced terranes in British Columbia. In: The continental crust and its mineral deposits. Geol. Assoc. Can. Spec. Paper 20: 441-456
- Irving, E. and Yole, R.A., 1972. Paleomagnetism and the kinematic history of mafic and ultramafic rocks in fold mountain belts. (Canadian) Earth Phys. Br. Pub. 42: 87-95
- Journey, J.M., 1981. Basement-cover relationships and structural evolution of the north-central Frenchman's Cap dome, Shuswap complex, southeastern British Columbia. Geol. Assoc. Can. Prog. Abstr. 6: A29
- Journey, J.M., 1983. Progressive metamorphism and inverted regional metamorphism associated with Mesozoic emplacement of the Shuswap-Monashee complex, SE British Columbia. Geol. Assoc. Can. Progr. Abstr. 15: 606
- Journey, J.M., 1985. Thermo-tectonic evolution of the Monashee Complex: a basement duplex, Omineca belt, British Columbia. Geol. Soc. Am. Progr. Abstr. 17: 364
- Journey, J.M. and Brown, R.L., 1985. Internal strain and kinematic evolution of the Monashee Complex: a basement duplex, British Columbia. Geol. Soc. Am. Progr. Abstr. 17: 364
- Keen, C.E., 1979. Thermal history and subsidence of rifted continental margins - evidence from wells on the Nova Scotian and Labrador shelves. Can. J. Earth Sci. 16: 505-522
- Kirby, S.H., 1983. Rheology of the lithosphere. Rev. Geophys. Space Phys. 21: 1458-1487

- Klepck, D., 1981. The Little River fault - new low-angle boundary of the northern Shuswap Complex, Quesnel Lake, British Columbia. Geol. Assoc. Can. Progr. Abstr. 6: A32
- Klepck, D.W., 1983. Structural and structural relations of the Milford, Kaslo, and Slocan Groups, Roseberry Quadrangle, Lardeau map-area, British Columbia. Geol. Surv. Can. Paper 83-1A: 229-233
- Klepck, D.W. and Wheeler, J.O., 1985. Stratigraphic and structural relations of the Milford, Kaslo, and Slocan Groups, Goat Range, Lardeau and Nelson map-areas, British Columbia. Geol. Surv. Can. Paper 85-1A: 277-286
- Leclair, A.D., 1983. Stratigraphy and structural implications of the central Kootenay arc rocks, southeastern British Columbia. Geol. Surv. Can. Paper 83-1A: 235-240
- Leonard, R., 1984. Metamorphism, structure, and stratigraphy around the Mount Blackman Gneiss, British Columbia. Geol. Surv. Can. Paper 84-1A: 121-124
- McDonough, M.R. and Simony, P.S., 1984. Basement gneisses and Hadrynian metasediments near Bulldog Creek, Selwyn Range, British Columbia. Geol. Surv. Can. Paper 84-1A: 99-102
- Mereu, R.F., Majumdar, S.C., and White, R.E., 1977. The structure of the crust and upper mantle under the highest ranges in the Canadian Rockies from a seismic refraction survey. Can. J. Earth Sci. 14: 196-208
- Miller, F.K., McKee, E.H., and Yates, R.G., 1973. Age and correlation of the Windermere Group in northeastern Washington. Bull. Geol. Soc. Am. 84: 3723-3730
- Monger, J.W.H., 1977a. Ophiolitic assemblages in the Canadian Cordillera. In: Coleman, R.G. (ed.) North American ophiolites. Oreg. Dept. Geol. Miner. Ind., Bulletin 95: 59-65
- Monger, J.W.H., 1977b. Upper Paleozoic rocks of the Canadian Cordillera and their bearing on Cordilleran evolution. Can. J. Earth Sci. 14: 1832-1859
- Monger, J.W.H., 1984. Cordilleran tectonics: a Canadian perspective. Bull. Soc. Geol. France 26: 255-278
- Monger, J.W.H., Clowes, R.M., Price, R.A., Simony, P.S., Riddihough, R.P., Woodsworth, G.J. with contributions from Currie, R.R., Hoy, T., Snively, P.D., and Yorath, C.J., 1985. Continent - Ocean Transect B2, Juan de Fuca Plate to Alberta. Geological Society of America, Decade of North American Geology Transect Series
- Monger, J.W.H. and Price, R.A., 1979. Geodynamic evolution of the

Canadian Cordillera - progress and problems. Can. J. Earth Sci. 16: 770-791

- Price, R.A., Monger, J.W.H., and Muller, J.E., 1981. Cordilleran cross-section, Calgary to Victoria. In: Thompson, R.I. and Cook, D.G. (eds.) Field guide to geology and mineral deposits. Annual Meeting Geol. Assoc. Can., Min. Assoc. Can., Can. Geophys. Un., Calgary
- Price, R.A., Monger, J.W.H., and Roddick, J.A., 1985. Cordilleran cross-section, Calgary to Vancouver. Field Trip 3, Fieldtrip Guidebook for 1985 meeting of the Cordilleran section of the Geol. Soc. Am., Vancouver, British Columbia
- Monger, J.W.H., Price, R.A., and Tempelman-Kluit, D.J., 1982. Tectonic accretion and the origin of the two major metamorphic and plutonic belts in the Canadian Cordillera. *Geology* 10: 70-75
- Monger, J.W.H., Souther, J.G., and Gabrielse, H., 1972. Evolution of the Canadian Cordillera: a plate tectonic model. *Am. J. Sci.* 272: 577-602
- Montgomery, J.R., 1985. Structural relations of the Quesnel Lake Gneiss, Isosceles Mountain area, southwest Cariboo Mountains, British Columbia. unpublished M.Sc. thesis, University of British Columbia, Vancouver, British Columbia, 101 pp.
- Montgomery, Scott L., 1978. Structural and metamorphic history of the Lake Dinford map area, Cariboo Mountains, British Columbia: ophiolite obduction in the southeastern Canadian Cordillera. unpublished M. Sc. thesis, Cornell University, Ithaca, New York, 170 pp.
- Morrison, M.L., 1982. Structure and petrology of the Malton Gneiss Complex. unpublished Ph. D. thesis, University of Calgary, Calgary, Alberta, 314 pp.
- Mountjoy, E.W., 1962. Mount Robson (SE) map-area. *Geol. Surv. Can. Paper* 61-31, 114 pp.
- Mountjoy, E.W., 1970. Geology of the Main Ranges between Tete Jaune Cache and Jasper. In: Edmonton Geological Society Field Conference Guidebook
- Mountjoy, E.W., 1971. Mount Robson map-area, Alberta and British Columbia. *Geol. Surv. Can. Paper* 71-1A: 227-232
- Mountjoy, E.W., 1976a. Edson to Valemount - The Rocky Mountain Front and Main Ranges along highway 16 in the Jasper area. *Geol. Assoc. Can. Edmonton Meeting Field Trip Guidebook* C-11: 4-42
- Mountjoy, E.W., 1976b. Medicine Lake map-area (83C/13, east and west). *Geol. Surv. Can. Open File* 371

- Mountjoy, E.W., 1980. Mount Robson, Alberta and British Columbia. Geol. Surv. Can. Map 1499A
- Mountjoy, E.W. and Price, R.A., 1984a. Jasper (83D/16) geological maps and cross-sections, 1:50,000 scale. Geol. Surv. Can. Open File 1038
- Mountjoy, E.W. and Price, R.A., 1984b. Amethyst Lakes (83D/9) geological maps and cross-sections; 1:50,000 scale. Geol. Surv. Can. Open File 1075
- Mountjoy, E.W., Forest, R., and Leonard, R., 1985. Structure and stratigraphy of the Miette Group, Selwyn Range, between Ptarmigan and Hugh Allan Creeks, British Columbia. Geol. Surv. Can. Paper 85-1A: 485-490
- Murphy, D.C., 1983. Suprastructure/Infrastructure transition in the southern Cariboo Mountain, Omineca Crystalline Belt, British Columbia. Geol. Soc. Am. Progr. Abstr. 15: 324
- Murphy, D.C., 1984. A note on faulting in the southern Rocky Mountain Trench between McBride and Canoe Reach, British Columbia. Geol. Surv. Can. Paper 84-1A: 627-630
- Murphy, D.C., 1985. Kinematic and tectonic implications of the vertical transition from upright to near-recumbent folds, central Cariboo Mountains, British Columbia. Geol. Soc. Am. Prog. Abstr. 17: 372
- Murphy, D.C., in press. Suprastructure/Infrastructure transition, east-central Cariboo Mountains, British Columbia: Geometry, kinematics, and tectonic implication. J. Struct. Geol.
- Murphy, D.C. and Journeay, J.M., 1982. Structural style in the Premier Range, Cariboo Mountains, southern British Columbia: preliminary results. Geol. Surv. Can. Paper 82-1A: 289-292
- Murphy, D.C. and Rees, C.J., 1983. Structural transition and stratigraphy in the Cariboo Mountains, British Columbia. Geol. Surv. Can. Paper 83-1A: 245-252
- Mutti, E., 1977. Distinctive thin-bedded turbidite facies and related depositional environments in the Eocene Hecho Group (south central Pyrenees, Spain). Sedimentology 24: 107-131
- Mutti, E. and Ricci-Lucchi, F., 1978. Turbidites of the Northern Apennines: Introduction to facies analysis. Int. Geol. Rev. 20: 125-166 (translation by T. H. Nilsen of: Le torbiditi dell'Apennino settentrionale: introduzione all'analisi di facies. Memorie della Societa' Geologica Italiana 1972: 161-199)
- Nguyen, K.K., Sinclair, A.J., and Libby, W.G., 1968. Age of the northern part of the Nelson batholith. Can. J. Earth Sci. 5: 955-957

- Normark, W.R., 1978. Fan valleys, channels, and depositional lobes on modern submarine fans: characters for recognition of sandy turbidite environments. *Bull. Am. Assoc. Petr. Geol.* 62: 912-931
- North, F.K. and Henderson, G.C.L., 1954. The Rocky Mountain Trench. *Alb. Soc. Petr. Geol. Guidebook*, 4th Annual Field Conference, Banff-Golden - Radium: 82-100
- Okulitch, A.V., 1974. Stratigraphy and structure of the Mt. Ida Group, Vernon, Seymour Arm, Bonaparte Lake, and Kettle River map-areas, British Columbia. *Geol. Surv. Can. Pap.* 74-1A: 25-30
- Oxburgh, E.R. and Turcotte, D.L., 1974. Thermal gradients and regional metamorphism in overthrust terranes with special reference to the eastern Alps. *Schweizerische Mineralogische und Petrografische Mitteilungen* 54: 641-662
- Packer, D.R. and Stone, D.B., 1974. Paleomagnetism of Jurassic rocks from southern Alaska. *Can. J. Earth Sci.* 13: 75-83
- Palmer, H., 1918. Early explorations in British Columbia for the Canadian Pacific Railroad. *Bull. Geog. Soc. Philadelphia* 14: No. 3
- Parrish, R. and Wheeler, J.O., 1983. U-Pb zircon ages of the Kuskanax batholith, southeastern British Columbia. *Can. J. Earth Sci.* 20: 1751-1756
- Pell, J., 1982. New correlations of Hadrynian strata, southern Canadian Cordillera. *Geol. Assoc. Can. Progr. Abstr.* 7: 73
- Pell, J., 1984. Stratigraphy, structure, and metamorphism of Hadrynian strata in the southeastern Cariboo Mountains, British Columbia. unpublished Ph.D. thesis, University of Calgary, Calgary, Alberta, 184 pp.
- Pell, J. and Simony, P.S., 1981. Stratigraphy, structure, and metamorphism in the southern Cariboo Mountains, British Columbia. *Geol. Surv. Can. Paper* 81-1A: 227-230
- Pell, J. and Simony, P.S., 1984. Stratigraphy of the Hadrynian Kaza Group between the Azure and North Thompson Rivers, Cariboo Mountains, British Columbia. *Geol. Surv. Can. Paper* 84-1A: 95-98
- Pigage, L.C., 1977. Rb-Sr dates for granodiorite intrusions in the northeastern margin of the Shuswap Metamorphic Complex, Cariboo Mountains, British Columbia. *Can. J. Earth Sci.* 14: 1690-1695
- Pigage, L.C., 1978. Metamorphism and deformation on the northeast margin of the Shuswap Metamorphic Complex, Azure Lake, British Columbia. unpublished Ph. D. thesis, University of British Columbia, Vancouver, British Columbia

- Price, R.A., 1981. The Cordilleran fold and thrust belt in the southern Canadian Rocky Mountains. In: Thrust and Nappe Tectonics. Geol. Soc. Lond. 427-448
- Price, R.A., 1982a. Mid-Proterozoic to Oligocene Cordilleran tectonic evolution, northeastern Washington and adjacent British Columbia. Geol. Soc. Am. Prog. Abstr. 14: 225-226
- Price, R.A., 1982b. The geotectonic significance of the Cordilleran Foreland Fold and Thrust Belt of Canada. Geol. Soc. Am. Progr. Abstr. 14: 593
- Price, R.A., 1983. The Rocky Mountain Belt of Canada: Thrust Faulting, tectonic wedging, and delamination of the lithosphere. Geol. Assoc. Can. Prog. Abstr. 8: A55
- Price, R.A. and Fermor, P.R., 1981. Three sections through the southern part of the Rocky Mountain Thrust and Fold Belt in southern Canada. Geol. Assoc. Can. Prog. Abstr. 6: A47
- Price, R.A. and Mountjoy, E.W., 1970. Geological structure of the Canadian Rocky Mountains between Bow and Athabasca - a progress report. In: Wheeler, J.O. (ed.) Structure of the Southern Canadian Cordillera. Geological Association of Canada Special Paper 6: 7-25
- Raesside, R. and Simony, P.S., 1983. Stratigraphy and deformational history of the Scrip Nappe, Monashee Mountains, British Columbia. Can. J. Earth Sci. 20: 639-650
- Ramsay, J.G. and Graham, R.H., 1970. Strain variations in shear belts. Can. J. Earth Sci. 7: 786-813
- Ramsay, J.G., Casey, M., and Kligfield, R., 1983. Role of shear in the development of the Helvetic fold-and-thrust belt of Switzerland. Geology 11: 439-442
- Ranalli, G., 1984. Rheological stratification of the continental lithosphere and its tectonic consequences. Abstracts, 27th International Geological Congress (Moscow) 3: 371
- Ranalli, G. and Murphy, D.C., in prep. Rheological stratification of the lithosphere. Tectonophysics
- Read, P.B. and Brown, R.L., 1981. Columbia River fault zone: southeastern margin of the Shuswap and Monashee complexes, southern British Columbia. Can. J. Earth Sci. 18: 1127-1145
- Read, P.B. and Wheeler, J.O., 1976. Geology, Lardeau west-half, British Columbia. Geological Survey of Canada Open File Map 432
- Reading, H.G.(editor), 1978. Sedimentary Environments and Facies Elsevier, New York 569 p.

- Rees, C.J., 1981. Western margin of the Omineca Belt at Quesnel Lake, British Columbia. *Geol. Surv. Can. Paper* 81-1A: 223-226
- Rees, C.J. and Ferri, F., 1983. A kinematic study of mylonitic rocks in the Omineca-Intermontane Belt tectonic boundary in east-central British Columbia. *Geol. Surv. Can. Paper* 83-1B: 121-125
- Rees, C.J., Irving, E., and Brown, R.L., 1985. Secondary magnetization of Triassic-Jurassic volcanoclastic rocks from the Quesnel terrane, British Columbia. *Geophys. Res. Lett.* 12: 498-501
- Reesor, J.E., 1973. Geology of the Lardeau map-area, east half, British Columbia. *Geol. Surv. Can. Memoir* 292, 78 p.
- Roeder, D.H., 1973. Subduction and orogeny. *J. Geophys. Res.* 78: 5005-5024
- Sanderson, D.J., 1980. The transition from upright to recumbent folding in the Variscan fold belt of SW. England: a model based on the kinematics of simple shear. *J. Struct. Geol.* 1: 171-180
- Sanderson, D.J., 1982. Models of strain variations in nappes and thrust sheets: a review. *Tectonophysics* 88: 201-233
- Sears, J.W. and Price, R.A., 1978. The Siberian connection - a case for Precambrian separation of the North American and Siberian platforms. *Geology* 6: 267-270
- Sibson, R.H., 1974. Frictional constraints on thrust, wrench, and normal faults. *Nature* 249: 542-544
- Simony, P.S., Ghent, E.D., Craw, D., Mitchell, W., and Robbins, D.B., 1980. Structural and metamorphic evolution of the northeast flank of the Shuswap Complex, southern Canoe River area, British Columbia. *Geol. Soc. Am. Mem.* 153: 445-461
- Slind, O.L. and Perkins, G.D., 1966. Lower Paleozoic and Proterozoic sediments of the Rocky Mountains between Jasper, Alberta, and Pine River, British Columbia. *Bull. Can. Petr. Geol.* 14: 442-468
- Stacey, R.A., 1973. Gravity anomalies, crustal structure, and plate tectonics in the Canadian Cordillera. *Can. J. Earth Sci.* 10: 615-628
- Steckler, M.S. and Watts, A.B., 1978. Subsidence of the Atlantic-type continental margins off New York. *Earth Plan. Sci. Lett.* 41: 1-13
- Stewart, J., 1972. Initial deposits in the Cordilleran geosyncline: evidence of a late Precambrian (<850 m. y.) continental separation. *Bull. Geol. Soc. Am.* 83: 1345-1360
- Stewart, J., 1976. Late Precambrian evolution of North America: plate tectonics implication. *Geology* 4: 11-15

- Struik, L.C., 1980. Geology of the Barkerville-Cariboo River area, central British Columbia: unpublished Ph. D. thesis, University of Calgary, Calgary, Alberta
- Struik, L.C., 1981. A re-examination of the type area of the Devonian-Mississippian Cariboo Orogeny, central British Columbia. *Can. J. Earth Sci.* 18: 1767-1775
- Struik, L.C., 1982a. Wells (93H/4), Spectacle Lakes (93H/3), Swift River (93A/13), and Cariboo River (93A/14) map-areas. *Geol. Surv. Can. Open File* 858
- Struik, L.C., 1982b. Snowshoe Formation (1982), central British Columbia. *Geol. Surv. Can. Paper* 82-1B: 117-124
- Struik, L.C., 1983a. Quesnel Lake (93A/10) and part of Mitchell Lake map-areas. *Geol. Surv. Can. Open File* 962
- Struik, L.C., 1983b. Spanish Lake (93A/11) and adjoining map-areas. *Geol. Surv. Can. Open File* 920
- Struik, L.C., 1984. Stratigraphy of Quesnel terrane near Dragon Lake, Quesnel Lake map-area, central British Columbia. *Geol. Surv. Can. Paper* 84-1A: 113-116
- Struik, L.C., 1985a. Pre-Cretaceous terranes and their thrust and strike-slip contacts, Prince George (east half) and McBride (west half), British Columbia. *Geol. Surv. Can. Paper* 85-1A: 267-272
- Struik, L.C., 1985b. Dextral strike-slip through Wells Gray Provincial Park, British Columbia. *Geol. Surv. Can. Paper* 85-1A: 305-310
- Struik, L.C., 1985c. Thrust and strike-slip faults bounding tectono-stratigraphic terranes, central British Columbia. *Guidebook to Field Trips, 1985 Meeting, Cordilleran section Geol. Assoc. Can./Geol. Soc. Am.*: 14-1 - 14-7
- Suppe, J., 1980. A retrodeformable cross-section of northern Taiwan. *Proc. Geol. Soc. China* 23: 46-55
- Tipper, H.W. and Smith, P.L., 1985. Paleobiogeography and displaced Jurassic terranes. *Geol. Soc. Am. Prog. Abstr.* 17: 413
- Turcotte, D.L. and Schubert, G., 1982. Geodynamics: Applications of Continuum Physics to Geological Problems. John Wiley and Sons.
- Van der Voo, R., Jones, M., Eberlein, G.D., and Churkin, M. 1980. Paleozoic paleomagnetism and northward drift of the Alexander terrane, southeastern Alaska. *J. Geophys. Res.* 85: 5281-5286
- Vink, G.E., Morgan, W.J., and Zhao, W.-L., 1984. Preferential rifting of continents: a source of displaced terranes. *J. Geophys. Res.* 89: 10,072-10,076

- Walker, R.G., 1967. Turbidite sedimentary structures and their relationship to proximal and distal depositional environments. *J. Sed. Petr.* 37: 25-43
- Walker, R.G., 1970. Review of the geometry and facies organization of turbidites and turbidite-bearing basins. In: Lajoie, J. (ed.), *Flysch sedimentology in North America* Geol. Assoc. Can. Special Paper 7: 219-251
- Walker, R.G., 1984. Turbidites and associated coarse clastic rocks. In: Walker, R.G. (ed.) *Facies models*, second edition. *Geoscience Canada Reprint Series 1*: 171-188
- Walker, R.G. and Mutti, E., 1973. Turbidite facies and facies associations. In: Bouma, A.H. and Middleton, G.V. (eds.) *Turbidites and deep water sedimentation*, Soc. Econ. Paleo. Miner. Pacific Section Short Course, Anaheim, California, 119-157
- Wheeler, J.O., 1965. Big Bend map-area, British Columbia (82N, east half). *Geol. Surv. Can. Paper* 63-32, 37 pp.
- Wheeler, J.O., Campbell, R.B., Reesor, J.E., and Mountjoy, E.W., 1972. Structural style of the Southern Canadian Cordillera. *Guidebook X01-A01*, 24th International Geological Congress, Montreal, Canada
- Willemin, J.H., 1984. Erosion and the mechanics of shallow level foreland thrusts. *J. Struct. Geol.* 6: 425-432
- Young, F.G., 1972. Stratigraphy of the upper Miette Group, central Rocky Mountains. *Geol. Surv. Can. Paper* 72-1A: 235-236
- Young, F.G., 1979. The lowermost Paleozoic McNaughton Formation and equivalent Cariboo Group of eastern British Columbia: piedmont and tidal complex. *Geol. Surv. Can. Bulletin* 288, 60 pp.
- Young, F.G., Campbell, R.B., and Poulton, T.P., 1973. The Windermere Supergroup of the southeastern Canadian Cordillera. *Proceedings of the Belt Symposium*, Moscow Idaho

1 OF/DE

MIDDLE KAZA GROUP

MID-FAN TURBIDITES BELOW MARKER UNIT

MARKER UNIT

TA

TAE (siltst.)

E+vtTAs (calc.) grey

grey silty phyllite and

calc. quartz sandston

black (graphitic) phy

thin dark grey to bla

grey (musc.-rich), si

grey limestone beds

20F/DE

UPPER KAZA GROUP

MARKER UNIT

30F/DE

210

200

190

180

170

160

TAamal

TAE

TA

TAE

TAamal

TA

TAamal

TAamal

TAE

TAamal

TAE

E+vtAs

TAE

TABE

TAE

TA

TAamal

TAE

TAB

TAE

TABE

TA

TAaBE

TA

TAaBCE

TABE

TAE

TA

TABC

ITABCamal

TAamal

TACE

TAE

TA

TA

E+vtAs

150

140

130

120

110

100

plane-laminated,
thin beds of white,
fine-grained quartz

Hi

Isaac Formation

4 OF/DE

150

140

130

120

110

100

5 OF/DE

160

TAg

TA

E+vtAs

TAE

150

E+vtAs

TAE

TAg

TA (siltst. at top)g

TA

TA siltst.

TAg

TA

140

TAamal

TAE

TA

TA

TA

TAB

130

TABE

TAB

TA

TAamal

TA

TAB

TA

TAg

120

E+vtAs

TAamal

TABE

TAB

TAg

E+vtAs

TAg

TAg

TA

TA

110

TAg

TAg

100

90

80

70

60

50

plane-laminated, gre
thin beds of white, c
fine-grained quartz
1 m. (locations of s
are diagrammatic)
phyllite near top

gently inclined trunc
(locations diagr

100

90

80

70

60

50

plane-laminated, green (chloritic) siltst. with
thin beds of white, contorted cross-laminated
fine-grained quartz sst. at intervals of less than
1 m. (locations of sst., lst., and grey phyllite beds
are diagrammatic); thin beds of grey lst. and silty
phyllite near top

gently inclined truncation surfaces
(locations diagrammatic)

EXPLANATION

phyllite (meta-shale)

siltstone (siltst.)

limestone

6 OF/DE

8
OF/DE

EXPLANATION

phyllite (meta-shale)

siltstone (siltst.)

contorted, cross-laminated calcareous quartz sandstone

sandy turbidites:

TA, TACE, TB, etc.: Bouma beds present

g, graded; intra, intraclast-rich; amal, amalgamated;

vt, very thin (cm-scale)

amalgamated beds of coarse-grained, commonly pebbly sandstone, multiple scour surfaces, commonly intraclast-rich

S_e

incised surface

direction of fining and thinning of beds

Isaac Formation

upper Kaza Group

middle Kaza Group

9 OF/DE

Hi

Isaac Formation

Hku

upper Kaza Group

Hkm

middle Kaza Group

10
OF/DE

110F/DE

120F/DE12

1 OF/DE

MODEL 1

TERRANE I

NORTH AMERICA

VOLCANIC ARC

MARGINAL BASIN

CONTINENTAL MARG

20F/DE

MODEL 1

NORTH AMERICA

CONTINENTAL MARGIN

MARGINAL BASIN

ARC OBDUCTION

"BACK-THRUSTING AND -FOLDING"

subsidence

ARC OBDUCTION

b.

"BACK-THRUSTING AND -FOLDING"

subsidence

c.

PRESENT

d.

30F/DE

ARC OBDUCTION

"BACK-THRUSTING AND -FOLDING"

PRESENT

40F/DE 4

A

B

C

D

E

Fig 44

10F/DE

MODEL 2

TERRANE I

NORTH AMERICA

VOLCANIC ARC

MARGINAL BASIN

CONTINENTAL MARG

ARC OBDUCTION

"BACK-THRUSTING AND FOLDING"

B C D E

MODEL 2

NORTH AMERICA

CONTINENTAL MARGIN

20F/DE

MARGINAL BASIN

ARC OBUCTION

"BACK-THRUSTING AND -FOLDING"

minor uplift

3 OF DE

ARC OBDUCTION

"BACK-THRUSTING AND -FOLDING"

PRESENT

4 OF / DE 4

A

B

C

D

E

10F/DE

RHEOLOGICAL STRUCTURE: EARLY JURASSIC (pre-Sinemurian) (MODEL 1)

20F/DE

RHEOLOGICAL STRUCTURE: EARLY JURASSIC (pre-Sinemurian) (MODEL 1)

$\log(\dot{\epsilon}_1 - \dot{\epsilon}_2)$ (MPa)

3 OF/DE

$\log(\sigma_1 - \sigma_3)$ (MPa)

40F/DE 4

10F/DE

RHEOLOGICAL STRUCTURE: EARLY JURASSIC (pre-Sinemurian) (MODEL 2)

20F/DE

RHEOLOGICAL STRUCTURE: EARLY JURASSIC (pre-Sinemurian) (MODEL 2)

$\log(\sigma_1 - \sigma_3)$ (MPa)

30F/DE

$\log(\sigma_1 - \sigma_3)$ (MPa)

40F/DE4

A B C D E

10F/DE

RHEOLOGICAL STRUCTURE: EARLY JURASSIC (post-Sinemurian) (MODEL 1)

20F/DE

RHEOLOGICAL STRUCTURE: EARLY JURASSIC (post-Sinemurian) (MODEL 1)

$\log(\sigma_1 - \sigma_3)$ (MPa)

30F/DE

40F/DE4

1 OF/DE

RHEOLOGICAL STRUCTURE: EARLY JURASSIC (post-Sinemurian) (MODEL 2)

20F/DE

RHEOLOGICAL STRUCTURE: EARLY JURASSIC (post-Sinemurian) (MODEL 2)

$\log(\sigma_1 - \sigma_3)$ (MPa)

$\log(\sigma_1 - \sigma_3)$ (MPa)

40E/DE 4

z (km)

10F/DE

RHEOLOGICAL STRUCTURE: MIDDLE JURASSIC (MODEL 1)

RHEOLOGICAL STRUCTURE: MIDDLE JURASSIC (MODEL 1)

20F/DE

a.

b.

3 OF/DE

$\log(\sigma_1 - \sigma_3) \text{ (MPa)}$

40F/DE 4

10F/DE

RHEOLOGICAL STRUCTURE: MIDDLE JURASSIC (MODEL 2)

RHEOLOGICAL STRUCTURE: MIDDLE JURASSIC (MODEL 2)

20F/DE

$\log(\sigma_1 - \sigma_3)$ (MPa)

4

MICROCOPY RESOLUTION TEST CHART
NATIONAL BUREAU OF STANDARDS
STANDARD REFERENCE MATERIAL 1010a
(ANSI and ISO TEST CHART No. 2)

a.

b.

$\log(\sigma_1 - \sigma_3)$ (MPa)

4 OF/DE 4

10F/DE

STRUCTURAL GEOMETRY OF THE AREA BETWEEN CASTLE CREEK AND

CARIBOO MOUNTAINS, BRITISH COLUMBIA

(North Half)

20F/DE

1
EK A EK AND MICA MOUNTAIN

3 OF/DE

4 OF/DE

M

B

A

M

O

U

N

5'0F/DE

Castle Creek

EXPLANATION

46
MOUNTAIN

6 OF/DE

Hkm

Hku

Icefield

Hkm

trace of major stratigraphic contact (observed, inferred, covered)

7
OF/DE

120°

9

7000

8
OF/DE

9 OF/DE

EXPLANATION

Hi

Isaac Formation

Hku

upper Kaza Group

Hkm

middle Kaza Group

(coarse stipple ornaments laterally continuous sandstone sequences used to outline structural geometry)

----- trace of major stratigraphic contact (observed, inferred, covered)

----- bedding surface trace

↑----- axial surface trace, antiform (D_2 , except where indicated)

↓----- axial surface trace, synform (D_2 , except where indicated)

↗₆₇ ↘₆₅

strike and dip of bedding: upright, younging direction observed, inferred

↘₅₇ ↗₉

strike and dip of bedding: overturned, younging direction observed, inferred

↑
|
↑

line of cross section

CROSS SECTIONS

100F/DE

11 OF/DE

9
7000
6000

12
OF/DE

120

Hkm

Hkm

trace of major stratigraphic contact (observed, inferred, covered)

bedding surface trace

axial surface trace, antiform (D_2 , except where indicated)

axial surface trace, synform (D_2 , except where indicated)

strike and dip of bedding: upright, younging direction observed, inferred

strike and dip of bedding: overturned, younging direction observed, inferred

line of cross section

CROSS SECTIONS

topographic profile

trace of bedding in plane of cross section

Elevations are in feet above sea level.

13 OF/DE

Scale 1:20,000

14 OF/DE

53°
120° 05'

150F/DE

Fig P-2

1 OF/DE

8000

8000

8000

20F/DE

STRUCTURAL GEOMETRY OF THE AREA BETWEEN

CARIBOO MOUNTAINS,

(South H)

BETWEEN CASTLE CREEK AND MICA MOUNTAIN

INS, BRITISH COLUMBIA

3 OF DE

th Half)

4 OF/DE

50F/DE

Raush River

6000

6000

6 OF DE

PREMIER

7 OF DE

CLINO

80F/DE

119° 30'

Hhtc

Hhtc

Hkl

Hkm

Hhtc

9/DE

EXPLANATION

Hkm

middle Kaza Group

Hkl

lower Kaza Group

og

dioritic ortho-gneiss

Hhtc

Horsethief Creek Group:

m, marble, a, amphibolite-bearing meta-sandstone and shale

(coarse stipple ornaments laterally continuous sandstone sequences
used to determine structural geometry)

— ····· trace of major stratigraphic contact (observed, inferred, covered)

— ····· bedding surface trace

↑
↓
— ····· axial surface trace, antiform

↑
↓
— ····· axial surface trace, anticline with overturned limb (D₂, except where indicated)

↑
↓
— ····· axial surface trace, synform

↑
↓
— ····· axial surface trace, syncline with overturned limb (D₂, except where indicated)

Raush Anticline

PREMIER

P R E M I E R

Hkm

Hkm

Black Martin
Creek

10 OF/DE

ANTICLINOR

Icefield

R I B O O

121

FRASER R.

120°

ROCKY MOUNT
(MAIN RANGE)

- ····· trace of major stratigraphic contact (observed, inferred, covered)
- ····· bedding surface trace
- axial surface trace, antiform
- axial surface trace, anticline with overturned limb (D_2 , except where indicated)
- axial surface trace, synform
- axial surface trace, syncline with overturned limb (D_2 , except where indicated)

strike and dip of bedding: upright, younging direction observed, inferred

strike and dip of bedding: overturned, younging direction observed, inferred

line of cross section

CROSS SECTIONS

topographic profile

trace of bedding in plane of cross section

Elevations are in feet above sea level.

13
OF/DE

14
OF/DE

NORMORIUM

MOUNTAINS)

Scale 1:20,000

15

0F/0

EXPLANATION

(to accompany maps of mesoscopic structures)

10F/DE**—STRATIGRAPHY—**

Hi	Isaac Formation
Hku	upper
Hkm	middle
Hkl	lower
	} Kaza Group
Hhtc	Horsethief Creek Group

----- major stratigraphic contact (observed, inferred, covered)

—STRUCTURAL FEATURES—

----- axial surface trace; phase of deformation indicated by the number of slashes (observed, inferred, covered)

----- trace of moderately to gently dipping fault (teeth on upper plate, observed, inferred, covered)

MINOR STRUCTURES

strike and dip of axial surfaces of D_2 crenulations or foliations; open symbols indicate eastwardly verging cleavage/bedding relationship; solid symbols indicate westwardly verging cleavage/bedding relationship; half-solid symbols indicate symmetrical cleavage/bedding relationship

strike and dip of axial surfaces of D_3 crenulations

strike and dip of axial surfaces of D_4 crenulations

trend and plunge of D_2 intersection lineation or hinge line; barb indicates direction of vergence of cleavage/bedding relationship; full arrowhead indicates symmetrical fold or cleavage/bedding relationship; unbarbed, vergence not recorded

Hi Isaac Formation

Hku upper

Hkm middle Kaza Group

Hkl lower

Hhtc Horsethief Creek Group

----- major stratigraphic contact (observed, inferred, covered)

— STRUCTURAL FEATURES —

— / — / — / — axial surface trace; phase of deformation indicated by the number, of slashes (observed, inferred, covered)

----- trace of moderately to gently dipping fault (teeth on upper plate; observed, inferred, covered)

MINOR STRUCTURES

67 78
strike and dip of axial surfaces of D_2 crenulations or foliations; open symbols indicate eastwardly verging cleavage/bedding relationship; solid symbols indicate westwardly verging cleavage/bedding relationship; half-solid symbols indicate symmetrical cleavage/bedding relationship

79
strike and dip of axial surfaces of D_3 crenulations

74
strike and dip of axial surfaces of D_4 crenulations

6 3 7
6 9 6
4 4
6
trend and plunge of D_2 intersection lineation or hinge line; barb indicates direction of vergence of cleavage/bedding relationship; full arrowhead indicates symmetrical fold or cleavage/bedding relationship; unbarbed, vergence not recorded

20F/DE2

Fig. P-4

10E/DE

MESOSCOPIC STRUCTURES: AREA 'A'

(location of Area A shown
in Figure 3-2)

poles to axial surfaces of D_4 crenulations

+ hinge lines of D_2 folds

• D_2 intersection and extension lineations

(Symbols explained on accompanying sheet)

poles to axial surfaces of D_2 crenulations

20F/DE2

MESOSCOPIC STRUCTURES: AREA 'B'

(location of Area B shown
in Figure 3-2)

1
OF/DE

(Symbols explained on accompanying sheet)

kilometres

D₂ intersection and extension lineations

poles to axial surfaces of
D₂ crenulations

20F/DE2

poles to axial surfaces of
D₄ crenulations

10F/DE

MESOSCOPIC STRUCTURES: AREA C

(location of Area C shown
in Figure 3-2)(Symbols explained on
accompanying sheet)

Fig P-9

MESOSCOPIC STRUCTURES: AREA F

(location of Area F shown
in Figure 3-2)

1 OF DE

(Symbols explained on accompanying sheet)

AREA F

WN

20F/DE

(Symbols explained on accompanying sheet)

Tete Creek

Hhtc

Hkm

Hkl

• D₂ intersection and extension lineations
+ D₂ hinge lines

poles to D₂ foliations

poles to axial surfaces of
D₃ crenulations

poles to axial surfaces of
D₄ crenulations

30E/DE

Fig P-8

1 OF/DE

MESOSCOPIC STRUCTURES: AREA 'E'

(location of Area E shown
in Figure 3-2)

2 OF 11

(Symbols explained on accompanying sheet)

5 OF 5

MICROCOPY RESOLUTION TEST CHART
NATIONAL BUREAU OF STANDARDS
STANDARD REFERENCE MATERIAL 1010a
(ANSI and ISO TEST CHART No. 2)

+ hinge lines of D_2 folds
 • D_2 intersection and extension lineations

poles to axial surfaces of D_2 crenulations

poles to axial surfaces of D_3 crenulations

3 OF DE

MESOSCOPIC STRUCTURES: AREA D
(location of AREA D shown in
Figure 3-2)

poles to axial surfaces of
D₂ crenulations

poles to axial surfaces of
D₄ crenulations

+ hinge lines of D₂ folds

• D₂ intersection and extension
lineations

(Symbols explained on accompanying
sheet)

poles to axial surfaces of
D₃ crenulations

205/DE2

END

FIN

7 04 87