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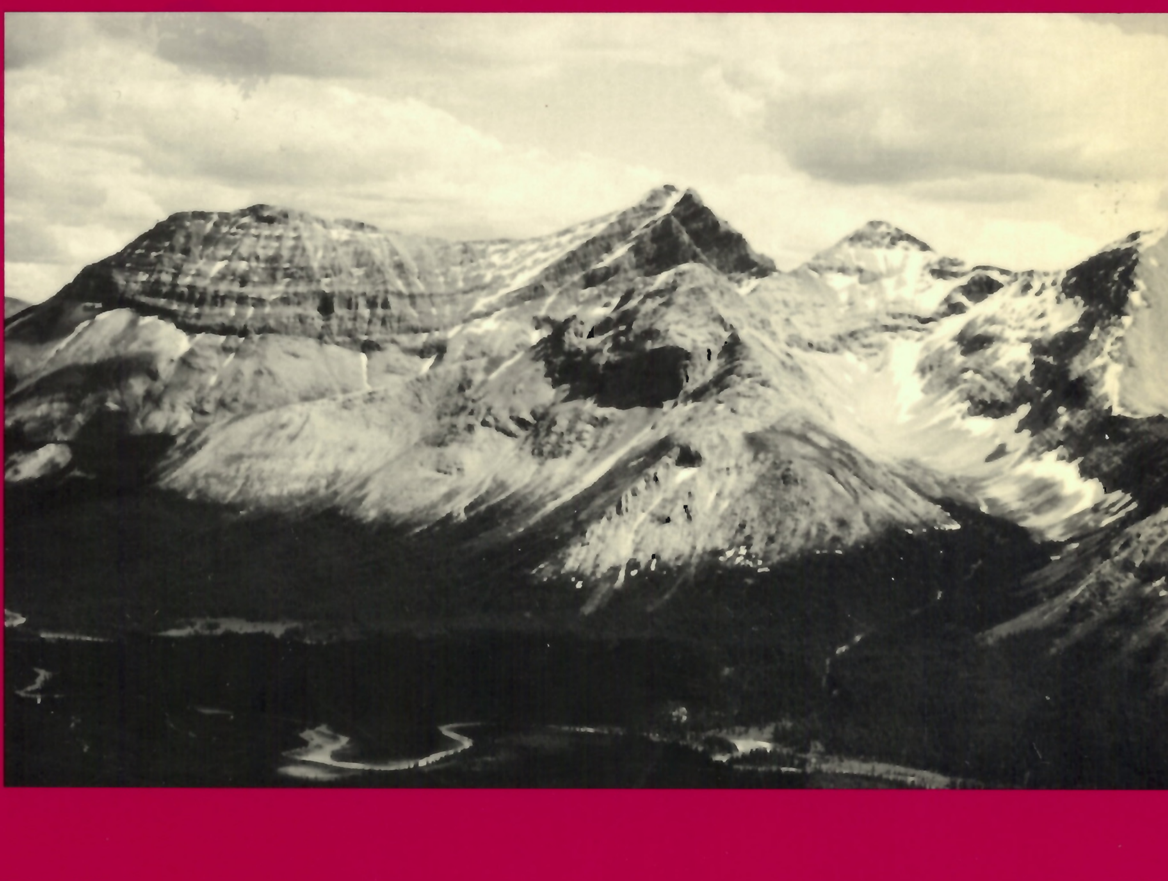
MEMOIR 425

STRATIGRAPHY, TECTONIC EVOLUTION AND STRUCTURAL ANALYSIS OF THE HALFWAY RIVER MAP AREA (94 B), NORTHERN ROCKY MOUNTAINS, BRITISH COLUMBIA



R.I. Thompson

1989

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PREFACE

The decision to update the geology of the Halfway River map area was triggered by the discovery of carbonate-hosted lead and zinc occurrences near Robb Lake in 1971. Initially, studies focused on the nature and distribution of carbonate-to-shale facies transitions in Silurian and Devonian strata, but soon expanded to include the stratigraphy and structure of the entire Paleozoic succession. The revised map permitted new stratigraphic and structural interpretations: the spatial relationship between metal occurrences and facies transitions became obvious; and cross-section interpretation of map patterns in terms of "thin-skinned tectonics" drew attention to the substantial gas potential within thrust panels of Devonian carbonate strata buried beneath the western part of the Foothills.

The updated database has proven useful. It helped guide Cominco geologists to one of the world's largest niobium occurrences – the Aley Deposit.

The Halfway River map area is underlain by thrust faulted and folded Upper Proterozoic, Paleozoic and Mesozoic strata. This report contains a systematic description of stratigraphic units, paleogeographic reconstructions for Paleozoic time, descriptions and analyses of structural style, tectonic reconstructions, and a description of important mineral and hydrocarbon occurrences. An updated 1:250,000 compilation map is supplemented by eight 1:50,000 preliminary maps and nine structural cross-sections.

Elkanah A. Babcock
Assistant Deputy Minister
Geological Survey of Canada

PRÉFACE

La découverte en 1971 de venues plombifères et zincifères contenues dans des roches carbonatées, à proximité de Robb Lake, a été à l'origine de la décision de mettre à jour la géologie de la carte de Halfway River. Initialement, les études portaient principalement sur la nature et la distribution des transitions de faciès des roches carbonatées aux schistes argileux dans les strates siluriennes et dévoniennes, mais se sont rapidement élargies de façon à inclure la stratigraphie et la structure de toute la succession paléozoique. La carte révisée a permis de formuler de nouvelles interprétations stratigraphiques et structurales: la relation spatiale entre les venues métallifères d'une part et les transitions de faciès d'autre part est devenue évidente; et l'interprétation de coupes de terrain d'après les structures cartographiées, du point de vue de la "tectonique mince", a mis en relief le potentiel notable en gaz naturel des nappes de charriage composées des strates carbonatées du Dévonien et enfouies sous de la partie ouest des Foothills.

La mise à jour de la base de données s'est avérée utile. Elle a aidé les géologues de la Cominco à trouver l'une des venues les plus vastes de niobium au monde – le gisement d'Aley.

Le sous-sol de la carte de Halfway River contient des strates du Protérozoïque supérieur, du Paléozoïque et de Mésozoïque, plissées et traversées par des failles chevauchantes. Dans ce rapport, on décrit de façon systématique les unités stratigraphiques et les reconstructions paléogéographiques du Paléozoïque, puis l'on décrit et l'on analyse le style structural, ainsi que les reconstructions tectoniques, et l'on décrit les importantes venues minéralisées et d'hydrocarbures. Une carte de compilation à l'échelle de 1:250 000, mise à jour, est complétée par huit cartes préliminaires à l'échelle de 1:50 000 et neuf coupes structurales.

Elkanah A. Babcock, sous-ministre adjoint
Commission géologique du Canada

CONTENTS

Figures

Tables

Maps

STRATIGRAPHY, TECTONIC EVOLUTION AND STRUCTURAL ANALYSIS OF THE HALFWAY RIVER MAP AREA (94 B), NORTHERN ROCKY MOUNTAINS, BRITISH COLUMBIA

Abstract

The Halfway River map area encompasses a segment of the Canadian Rocky Mountains between the Peace (lat. 56°N) and Halfway (lat. 57°N) rivers. On the east is the Alberta Plateau, a region of flat-lying Cretaceous strata having little relief; bisecting the map area from south to north is the Foothills structural subprovince, a belt of folded Carboniferous through Lower Cretaceous strata forming long parallel ridges breached by east-flowing streams; on the west is the Rocky Mountains structural subprovince, consisting of folded and thrust faulted Upper Proterozoic through Carboniferous strata sculpted into jagged peaks. Relief within the Foothills and Rocky Mountains subprovinces averages 1000 m.

Passive margin sedimentation persisted from the Late Proterozoic until the Late Jurassic. Alpine orogenesis began late in the Jurassic and lasted until the early Tertiary. Thrusts and detached folds formed along the eastern margin of the fold and thrust belt and migrated eastward ahead of the advancing deformation front.

The passive margin (miogeocline) succession is about 12 000 m thick and represents a westward thickening wedge of shelf carbonates that "shaled out" into off-shelf, fine grained, clastic sediments. The overlying synorogenic foreland basin succession has a maximum thickness of 4000 m within an eastward tapering wedge made up of deltaic and prodeltaic marine shale, siltstone and sandstone with minor conglomerate.

Supracrustal shortening across the area is about 60 km; including 25 to 30 km within the chevron and box folds of the Foothills subprovince.

There are 11 producing gas fields east of the Foothills subprovince. Coal is known within Cretaceous strata, lead and zinc occur in brecciated Silurian and Devonian carbonate strata, and a carbonatite deposit occurs within Lower Ordovician carbonate strata.

Résumé

La région couverte par la carte de Halfway River englobe un segment des Rocheuses canadiennes entre les rivières Peace (latitude 56°N) et Halfway (latitude 57°N). À l'est, se trouve le plateau de l'Alberta, région de strates crétacées pratiquement horizontales, sans reliefs; la sous-province structurale des Foothills, zone de strates plissées du Carbonifère au Crétacé inférieur qui forment de longues crêtes parallèles interrompues par des cours d'eau de direction est, divise en deux la carte, du sud au nord; à l'ouest, se trouve la sous-province structurale des Rocheuses, composée de strates du Protérozoïque supérieur au Carbonifère, plissées et traversées par des failles chevauchantes, et sculptées en arêtes. À l'intérieur des sous-provinces des Foothills et des Rocheuses, les reliefs atteignent en moyenne 1000 m.

La sédimentation typique de marge passive a persisté du Protérozoïque supérieur au Jurassique supérieur. L'orogenèse alpine a commencé vers la fin du Jurassique, et a duré jusqu'au Tertiaire inférieur. Des charriages et plis décollés se sont formés le long de la marge est de la zone de plissements et charriages, et ont migré vers l'est en avant du front de déformation en progression.

La succession de la marge passive (miogéoclinal) a une puissance d'environ 12 000 m et représente un prisme s'épaississant vers l'ouest, composé de roches carbonatées de plate-forme qui faisaient progressivement place à de minces strates de sédiments clastiques fins au large de la plate-forme. La succession susjacente et synorogénique du bassin d'avant-pays, atteint une épaisseur maximum de 4000 m à l'intérieur d'un biseau qui s'amincit vers l'est et se compose de schistes argileux marins deltaïques et prodeltaïques, de microgrès et de grès, accompagnés d'un peu de conglomérat.

Dans toute la région, le raccourcissement supracrustal est d'environ 60 km; y compris 25 à 30 km à l'intérieur des plis en accordéon et plis coffrés de la sous-province des Foothills.

À l'est de la sous-province des Foothills, existent 11 champs gazéifères producteurs. On sait qu'il existe de charbon à l'intérieur des strates crétacées, du plomb et du zinc à l'intérieur des strates carbonatées bréchiformes du Silurien et du Dévonien, et un gisement de carbonatite à l'intérieur des strates carbonatées de l'Ordovicien inférieur.

Summary

The Halfway River map area includes a segment of the northern Canadian Rocky Mountains between the Peace (56°N) and Halfway rivers (57°N). The Foothills and Rocky Mountains structural subprovinces are bounded by the undeformed Alberta Plateau on the east and by the Northern Rocky Mountain Trench on the southwest. The geological history of the map area typifies that for the Canadian Rocky Mountains: extending- and passive-margin type sedimentation persisted from the Late Proterozoic until Late Jurassic; alpine orogenesis followed, causing the sequential west to east development of detached thrusts and rootless folds; orogenic activity ceased early in the Tertiary; and isostatic uplift and erosion persists to the present.

The stratigraphic succession can be divided into two parts: a continental terrace wedge (miogeocline) derived mainly from, and deposited onto, the subsiding and intermittently extending western margin of continental North America; and a succeeding prism of synorogenic clastic strata eroded from the terrace wedge during mountain building, and deposited into a marginal inland sea on the east. Aggregate thickness of the continental terrace wedge is in the order of 12 000 m; at any one location 8000 m is a likely maximum. The oldest exposed strata are Upper Proterozoic clastics and carbonate deposited on an unstable (extending?) crust. During most of the Paleozoic, thick sequences of shallow water shelf carbonate and laterally adjacent, off-shelf clastics accumulated on a gently subsiding crust. From Middle Ordovician until Middle Devonian time, the boundary between shelf and off-shelf strata outlined the southeastward closing Ospika Embayment, the southern boundary of which was the northern flank of Peace River Arch. Middle to Late Devonian time saw a rearrangement of paleogeographic patterns: a major shale transgression overstepped the shelf, pushing the western limit of shelf carbonate deposition far to the east. As carbonates were deposited westward during the Early Carboniferous, the Ospika Embayment ceased to exist, and the Peace River Arch inverted into a depositional trough or embayment. During Late Carboniferous through Early Triassic time, deposition of fine grained clastics predominated; during the Middle and Late Triassic, shallow water carbonates were deposited westward for one last time before transgression of Jurassic seas and the deposition of shale. The first stratigraphic evidence of orogenic activity farther west is preserved in uppermost Jurassic and Lower Cretaceous sandstone; this sandstone formed part of what became a vast deltaic complex fed by rivers from the south and west. Sediment continued to pour eastward, episodically, from the deforming Cordillera into a marginal drainage system and inland sea. As deformation progressed eastward, the clastic wedge prograded ahead of it, as did the basin axis.

In Halfway River map area, the foreland fold and thrust belt comprises two structural subprovinces: Foothills, and Rocky Mountains. The Foothills subprovince is characterized by chevron- and box-style folds with few attendant thrusts; the Rocky Mountains subprovince is structurally diverse, containing large, imbricated thrust sheets, intricate folds, and steeply dipping extension and contraction faults. The transition from the mainly folded Foothills to the thrust faulted and folded Rocky Mountains occurs within a regional detachment zone of Upper Devonian and Lower Carboniferous shale called the Besa River Formation. Displacement at depth across "nonsurfacing" thrusts affecting lower and middle Paleozoic carbonate strata is compensated for by folding within Carboniferous and younger strata closer to the surface. The nonsurfacing thrusts are termed "blind"; they played an important role in the structural development of the region.

Shortening across the Foothills subprovince approximates 25 to 30 km; total shortening across the map area approximates 60 km.

Structural trends diverge northward from Peace River: the Foothills and eastern Rocky Mountains bear north-northwest, the western Rocky Mountains northwest. The locus of divergence is in line with pre-Silurian north-northwest trending faults and dike swarms found within the Tuchodi Anticline in Tuchodi Lakes map area, 150 km farther north. These older features suggest a period of crustal extension and development of basement ramps that later influenced the trend of detached structures formed over them.

There are 11 producing gas fields located east of the Foothills subprovince beneath the Alberta Plateau. Traps are mainly stratigraphic within Cretaceous (Fort St. John Group, Minnes Group), Triassic (Baldonnel, Charlie Lake, and Liard formations) and Carboniferous (Prophet Formation) strata. Additional exploration potential lies along the western margin of the Foothills subprovince where mid-Devonian barrier facies carbonate (Dunedin Formation) is repeated at depth across blind thrusts.

Carbonate-hosted lead-zinc showings in the form of vein and breccia fillings, replacements and fine disseminations, are known from Silurian and Devonian successions; only the Robb Lake occurrence has economic potential. The mineralization is interpreted as postdepositional, associated with the mobilization of metallic ions by fluids generated during shale diagenesis and dewatering.

A niobium-bearing carbonatite complex intrudes Lower Ordovician carbonate rocks (Kechika Formation). It consists of a carbonatite core (dolomite, apatite, pyrite, phlogopite, calcite) surrounded by an amphibolite ring (riebeckite, arfvedsonite, acmite). The age of the carbonatite complex is Early Carboniferous.

Coal potential is slight. Flooding of Peace Reach has made the coal-bearing Cretaceous Bullhead Group inaccessible in that area. Farther north the group contains little or no coal.

Sommaire

La carte de Halfway River comprend un segment du nord des Rocheuses canadiennes, entre les rivières Peace (56°N) et Halfway (57°N). Les Foothills et les sous-provinces structurales des Rocheuses sont limités par le plateau non déformé de l'Alberta à l'est et par le sillon des Rocheuses septentrionales au sud-ouest. L'histoire géologique de cette carte est représentative des Rocheuses canadiennes; la sédimentation typique d'une marge en extension et d'une marge passive a duré du Protérozoïque supérieur au Jurassique supérieur; l'orogenèse alpine lui a succédé, causant d'ouest en est le développement séquentiel de nappes de charriage et de plis sans racines; l'activité orogénique s'est terminée au début du Tertiaire; le soulèvement isostatique et l'érosion se poursuivent encore maintenant.

On peut subdiviser la succession stratigraphique en deux parties: un biseau composé d'une terrasse continentale (miogéoclinal) dérivé principalement de, et déposé sur, la marge ouest du continent nord-américain en subsidence et de façon intermittente en extension; à ce terrain succédait un prisme de strates clastiques synorogéniques, enlevé par l'érosion du biseau de terrasse continental durant la formation des montagnes puis déposé dans une mer intérieur marginale à l'est. L'épaisseur cumulative de la terrasse continentale en biseau est de l'ordre de 12 000 m; en n'importe quel endroit, il est probable que l'épaisseur maximum est de 8000 m. Les strates les plus anciennes exposées sont des roches clastiques et carbonatées du Protérozoïque supérieur déposées sur une croûte instable (en expansion?). Durant la majeure partie du Paléozoïque, d'épaisses séquences de roches carbonatées de plate-forme peu profonde et de roches clastiques latéralement adjacentes, situées au large de la plate-forme, se sont accumulées sur une croûte soumise à une lente subsidence. De l'Ordovicien moyen au Dévonien moyen, la limite entre les strates de la plate-forme et les strates du large a délimité le rentrant d'Ospika qui se fermait vers le sud-est, et dont la limite sud était constituée par le flanc nord de l'arche de Peace River. Durant de Dévonien moyen à supérieur, les structures paléogéographiques se sont remodelées: une importante transgression avec dépôt de schistes argileux a progressivement gagné la plate-forme, poussant la limite ouest de sédimentation des roches carbonatées de plate-forme loin à l'est. À mesure que les roches carbonatées se déposaient vers l'ouest durant le Carbonifère inférieur, le rentrant d'Ospika a cessé d'exister, et l'arche de Peace River s'est transformée en un sillon ou une baie dans lesquels s'est effectuée la sédimentation. Du Carbonifère supérieur au Trias inférieur, a dominé la sédimentation de roches clastiques fines; durant le Trias moyen et le Trias supérieur, des roches carbonatées d'eau peu profonde se sont déposés vers l'ouest une dernière fois avant la transgression des mers jurassiques et la sédimentation des schistes argileux. Le premier indice stratigraphique d'une activité orogénique plus à l'ouest est conservé dans les grès du sommet du Jurassique et les grès du Crétacé inférieur; ce grès constitue une partie de ce qui est devenu un vaste complexe deltaique alimenté par des cours d'eau en provenant du sud et de l'ouest. Les sédiments ont continué à s'écouler épisodiquement vers l'est à partir de la Cordillère en voie de déformation jusque dans un réseau hydrographique marginal et dans une mer intérieure. À mesure que la déformation progressait vers l'est, le prisme de sédiments clastiques a progressé en avant de celle-ci, de même que l'axe du bassin.

Sur la carte de Halfway River, la zone de plissement et de charriage d'avant-pays comprend deux sous-provinces structurales: les Foothills et les montagnes Rocheuses. La sous-province des Foothills est caractérisée par des plis en accordéon et plis coffrés accompagnées d'un petit nombre de charriages; la sous-province des Rocheuses est structuralement diversifiée, et contient de vastes nappes de charriage imbriquées, des plis complexes, et des failles d'extension et de compression de fort pendage. La transition des Foothills principalement plissés aux Rocheuses plissées et traversées par des failles chevauchantes existe à l'intérieur d'une zone régionale de décollement des schistes argileux du Dévonien supérieur et du Carbonifère inférieur, appelée formation de Besa River. Le déplacement survenu en profondeur à travers les charriages "n'apparaissant pas en surface", et touchant les strates carbonatées du Paléozoïque inférieur et moyen, est compensé par le plissement des strates carbonifères et plus récentes, à plus grande proximité de la surface. Les charriages n'apparaissant pas en surface sont appelés "aveugles"; ils ont joué un rôle important du point de vue du développement structural de la région.

La raccourcissement survenu dans la sous-province des Foothills est d'environ 25 à 30 km; la raccourcissement total dans l'ensemble de la carte se rapproche de 60 km.

Les axes structuraux divergent vers le nord à partir de Peace River: les Foothills et l'est des Rocheuses ont une direction nord-nord-ouest, l'ouest des Rocheuses une direction nord-ouest. Le lieu géométrique de divergence est aligné avec les failles pré-siluriennes d'orientation générale nord-nord-ouest, et avec les essaims de dykes que l'on rencontre à l'intérieur de l'anticlinal de Tuchodi sur la carte des lacs Tuchodi, à 150 km plus au nord. Ces structures plus anciennes suggèrent l'existence d'une période d'expansion crustale et de développement de rampes sur le socle, qui plus tard ont influencé la direction générale des structures décollées qui se sont formées au-dessus de ces rampes.

Il existe 11 champs pétrolifères productifs à l'est de la sous-province des Foothills au-dessous du plateau de l'Alberta. Les pièges sont principalement de type stratigraphique à l'intérieur des strates crétacées (groupe de Fort Saint-John, groupe de Minnes), triasiques (formations de Baldonnel, de Charlie Lake et de Liard) et carbonifères (formation de Prophet). Il existe d'autres cibles potentielles d'exploration le long de la marge ouest de la sous-province des Foothills où les carbonates d'un faciès de barrière, d'âge dévonien moyen (formation de Dunedin) se répètent en profondeur sur toute l'étendue des charriages aveugles.

Les venues plombifères et zincifères, contenues dans des roches carbonatées et se présentant sous forme de remplissages de filons et de brèches, de substitutions et de fines disséminations, ont été observées dans des successions siluriennes et dévoniennes; seule la venue minéralisée de Robb Lake présente un potentiel économique. On a interprété la minéralisation comme étant post-sédimentaire, et associée à la mobilisation d'ions métalliques par des fluides générés durant la diagenèse et la déshydratation des schistes argileux.

Le complexe à carbonatite niobifère est intrusif dans des roches carbonatées de l'Ordovicien inférieur (formation de Kechika). Il est composé d'un noyau de carbonatite (dolomie, apatite, pyrite, phlogopite, calcite), environné d'un anneau amphibolitique (riebeckite, arfvedsonite, acmite). Le complexe à carbonatite date du Carbonifère inférieur.

Le potentiel houiller est faible. L'inondation de la zone de Peace Reach a rendu inaccessible dans cette région le groupe crétacé de Bullhead, qui contient du charbon. Plus au nord, le groupe contient peu ou pas de charbon.

CHAPTER 1

INTRODUCTION

The Halfway River map area covers a narrow segment of the Canadian Rocky Mountains between the Peace (lat. 56°00'N) and Halfway (lat. 57°00'N) rivers. It is a landscape of rugged peaks and smooth, rounded ridges cut by steep-walled, flat-bottomed valleys. To the east is the topographically subdued Alberta Plateau, and to the west, forming a discrete linear boundary, is the Rocky Mountain Trench (Fig. 1).

The Canadian Rocky Mountains physiographic region forms part of the much longer Cordilleran Foreland Fold and Thrust Belt, which extends from the Yukon-Alaska boundary around the arc of the Mackenzie Mountains and south to California, thence east across Arizona and New Mexico, and south into Mexico (King, 1959). The fundamental stratigraphic and structural elements of this 10 000 km long belt change little along its strike. Sedimentary rocks were deposited along the margin of ancestral North America during Proterozoic, Paleozoic and early Mesozoic time. In response to a succession of tectonic episodes lasting from latest Jurassic through early Tertiary time, they were displaced relatively eastward in a series of overlapping thrust sheets and detached (rootless) fold complexes.

The geology of the Halfway River map area provides a synoptic view of foreland fold and thrust belt evolution. Across this narrow segment of the Canadian Rocky Mountains one may reconstruct the Proterozoic through Cretaceous stratigraphic record, and decipher the structural linkages manifest in the lateral transition from undeformed, flat-lying Cretaceous rocks in the east, to metamorphosed, folded and thrust-faulted Proterozoic strata in the west.

The record of exploration within the Halfway River map area dates back to the late 1700's, when the Peace River served as an important transportation link for fur traders travelling from the east into north-central British Columbia (MacKenzie, 1801). The Peace River valley also provided a first glimpse of the geology of the region, "noted in passing", by members of early geological expeditions travelling the major river systems of northeastern British Columbia (Selwyn, 1877; Dawson, 1881; McConnell, 1896). By 1910, interest in the economic potential of coal seams outcropping along the Peace River canyon brought Provincial and Federal Government geologists to the area (Galloway, 1913, 1924; McLearn, 1918, 1921, 1923). These preliminary studies were the beginnings of more-or-less continuous geological interest, shared by industry and government, that focused primarily on oil, gas and coal potential, as well as Mesozoic stratigraphy and biostratigraphy.

Limited accessibility confined most geological studies to the Peace River and its major tributaries. The physical barriers associated with geological exploration of rugged mountainous terrains were not overcome until the late 1950's, when major petroleum exploration companies, pioneering the use of helicopters for mapping purposes, embarked on systematic reconnaissance mapping of the Northern Rocky Mountain region.

Irish (1970) began reconnaissance mapping of the Halfway River map area in 1959. Four field seasons, the first two supported logistically by pack horses and the final two by helicopter, completed the job, and led to the first published

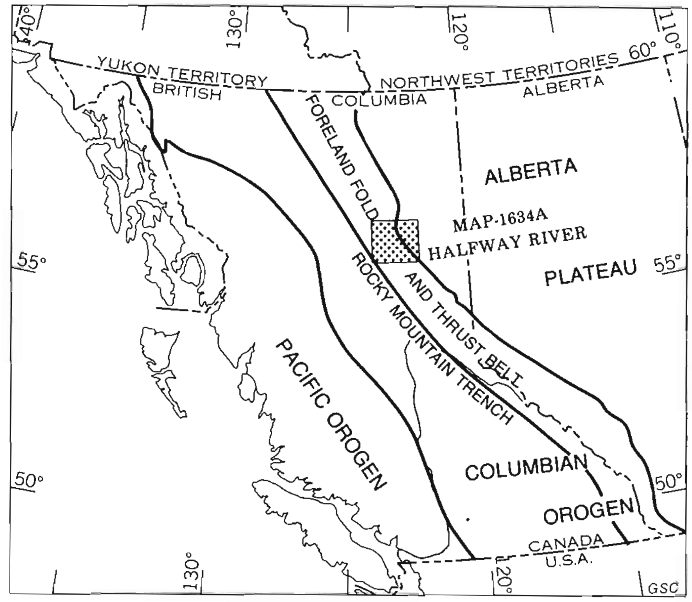

__Figure 1.__ Location map of Halfway River map area and its position relative to the major geological subdivisions of the Canadian Cordillera. The Columbia and Pacific orogens each comprise a metamorphic-plutonic core zone flanked by belts in which the tectonic transport is away from the cores. The Halfway River map area spans a narrow portion of the Foreland fold and thrust belt on the eastern flank of the Columbia orogen core zone.

geological map of the Halfway River area. It is an excellent piece of work, and within 12 months of its publication, had been used to practical advantage by mineral explorationists who adopted it as the geological base for mineral exploration that culminated in the discovery of lead-zinc occurrences near Robb Lake in late August, 1971. This discovery sparked a staking and exploration rush that lasted until 1974. By that time, a host of new geological problems, requiring investigation at a more refined scale, needed attention. This study was initiated on that basis, and focused, initially, on the stratigraphy and structure of Paleozoic strata around the Robb Lake lead-zinc occurrences. A primary objective was to assess the relationship, if any, between regional facies patterns and mineralization.

By 1975, the interest of mining companies in the Halfway River map area had waned; several mineral occurrences were known but only the Robb Lake occurrences held promise. Exploration emphasis swung from minerals to hydrocarbons – the potential for deep Foothills-type gas reservoirs was being re-evaluated. Producing gas fields already existed on the Alberta Plateau along the eastern boundary of the Halfway River map area, and long range exploration programs sought to evaluate the potential for deformed gas-bearing strata at deeper levels beneath the Foothills and eastern margin of the Rocky Mountains. Interest peaked in 1979-1980. One deep exploration well was drilled in 1981-1982, but most exploration acreages await preliminary evaluation. High exploration costs and a sagging gas market combined to stall exploration and evaluation activity.

Unlike the work of Irish, which was begun with an essentially blank geological database, this study benefited from his work as well as that of Taylor (1983, 1979), Taylor and Stott (1979, 1973), and Stott et al. (1983) who, between 1962 and 1974, headed a reconnaissance mapping operation that covered most of the northern Rocky Mountains region.

This report is based on fieldwork done in 1975 and 1976. Mapping was restricted to Triassic and older strata, primarily in the western part of the map area. These data are presented at 1:50 000 scale. The 1:250 000 compilation is an amalgamation of the 1:50 000 scale data with existing 1:250 000 maps showing the distribution of Jurassic and Cretaceous strata (Stott, 1967b).

Accessibility

Few roads extend into the Halfway River map area. The Alaska Highway cuts across the northwestern corner, and poor quality dirt roads extend from it to nearby ranches, but these roads do not reach the Foothills, and are of little use to geological operations. Seismic trails provide winter access by tracked vehicle along the Halfway River to Robb Lake, and along portions of Chowade, Cypress and Graham rivers. Horse trails are maintained along most major rivers by hunting guides and ranchers living in the region.

Most geological work is done from base camps located within the Foothills or Rocky Mountains, using helicopter and fixed-wing aircraft support. Mackenzie, British Columbia, is the usual supply centre for aircraft and material used by mining exploration companies. Petroleum exploration companies supply from Dawson Creek, British Columbia, or Fort St. John, Alberta. Float equipped aircraft can land on Lady Laurier Lake, Robb Lake, and Peace Reach (flooded Peace River); dirt landing strips are found at Christina Falls on the Graham River, at the head of Lady Laurier Lake, and on the Halfway River 10 km east of Robb Lake – none is necessarily maintained.

Acknowledgments

A successful field operation requires hard work, tolerance and thoughtful cooperation by field party members. The success of this operation was assured by the dedicated participation and excellent support provided by the following persons: D. Noakes and R. Day (geological assistants, 1975) S. Preston (camp cook, 1975), S. Trollope and N. Godfrey (geological assistants, 1976) and G. Wilson (camp cook, 1976). Alpine Helicopters Ltd. provided expert pilots and mechanics, and helicopters kept in excellent operating condition. Northern Thunderbird Airlines Ltd. provided prompt and reliable fixed-wing support.

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CHAPTER 2

A SYNOPSIS OF NORTHERN ROCKY MOUNTAIN GEOLOGY

Introduction

The thick succession of thrust-faulted and folded sedimentary rocks within the Halfway River map area formed part of an extensive westward-thickening prism deposited at the western margin of the North American craton beginning in the Middle Proterozoic and ending in the early Tertiary. Despite a long and complex depositional history, the sedimentary prism may be divided into two fundamental and sequential parts, each of which reflects a depositional history associated with a particular tectonic setting. The lower, older part comprises a broad, continental terrace wedge of mature, shallow-water clastic and carbonate rocks that change facies westward into a thinner succession of fine grained, deeper water clastic rocks. This is the miogeocline assemblage (Figs. 2, 3) – the sedimentary record of long-lived crustal stability characterized by the deposition of regionally continuous litho- and time-stratigraphic rock units on a gently subsiding cratonic margin.[1] By contrast, the overlying, younger part of the sedimentary prism consists of syntectonic clastic rocks generated during brief bursts of tectonic activity, when mountains were formed and partially eroded. This is the clastic wedge assemblage – the sedimentary record of tectonic events within the eastern part of the Canadian Cordillera from Late Jurassic through Early Eocene time (Price, 1973). Hence, the sedimentary prism that makes up the eastern part of the Canadian Cordillera records two essential phases of geological evolution: a "pre-orogenic" miogeoclinal phase that spanned more than 1500 million years; and a brief synorogenic phase, perhaps little more than 100 million years in duration, when sedimentary debris was shed into a marginal foreland basin from the uplifted and eroding miogeocline.

The term miogeocline, as it is used here, includes the broad, shallow water, continental terrace wedge as well as the deeper water shaly facies that accumulated as an outboard extension from the terrace wedge, perhaps on the continental rise.[2] Along most parts of the Canadian Cordillera, shallow-water carbonate units can be traced westward into fine grained clastic facies dominated by shale (graptolitic), siltstone, chert and some volcanics (Gordey, 1978; Cecile, 1982; Thompson, 1976; Gabrielse, 1974; Aitken, 1971). The characteristics and positions of these transitions vary with time such that shallow water limestones and dolostones are overlain by deeper water shales and vice versa.

Miogeoclinal strata thicken from an aggregate of 1500 m on the east, adjacent to the thin, cratonic platform succession, to a maximum of 10 000 to 15 000 m on the west near the edge of the continental terrace. These aggregate thicknesses are remarkably consistent along strike despite marked variations in the thickness of individual time stratigraphic assemblages (Gabrielse, 1974), and suggest that the character of the underlying crust and mechanisms that controlled subsidence did not vary significantly along the strike of the miogeocline.

Normal thickness of continental lithosphere results in too much buoyancy to sustain the long-lived, gentle subsidence required for the deposition of a rock prism having the dimensions of a miogeocline (Watts and Ryan, 1977). Alternatives are: deposition on oceanic crust, on physically altered continental crust, or on a combination of both. Studies using models of rifted continental margins by Keen et al. (1981) suggest that a miogeocline, like that now accumulating along the Atlantic continental margin of North America, is deposited on stretched continental lithosphere (Fig. 2). The stretching – analogous to thermal necking – occurs during the initial phase of intracontinental rifting above a hot spreading centre. Eventually, the "necked" crust separates and an ocean basin floored by new oceanic crust intervenes. The thinned continental crust subsides, partly because of the accumulating sedimentary load, and partly because of the effects of cooling (increase in density) as the rifted margin travels farther from the spreading centre. Hence the miogeocline is deposited on a subsiding shelf of cooling, attenuated, continental crust that represents a broad transition zone up to 200 km wide, between continental crust of normal thickness, called the platform, and thin, more dense, oceanic crust. Differential subsidence at the platform-shelf boundary produces a flexure called the hinge line. Price (1981) has suggested a similar model for the southern part of the Canadian Cordillera. He argues that the miogeocline was deposited on attenuated lithosphere and/or oceanic crust and that the hinge line followed the present western limb of the Purcell Anticlinorium. All of the miogeoclinal rocks now east of that trend have been transported laterally away from the subsiding shelf up and onto the platform.

Using the "Atlantic-type" continental margin as a physical model, the primary facies belts of the platform and the miogeocline (i.e. subsiding shelf) are described as follows (Fig. 2). The thin veneer of mature clastic and carbonate rocks deposited on the platform (normal thickness continental crust) will be called the platform facies. The continental terrace wedge of shallow water clastics and carbonates that overlies the subsiding shelf and forms the eastern part of the miogeocline will be called the shelf facies. The succession of laterally equivalent, generally fine grained clastic rocks beyond the shelf facies will be called the off-shelf facies. The transition from shelf facies to off-shelf facies can be abrupt or gradual, high relief or low relief, and it may shift in position through time. Whether or not most of the off-shelf facies is deposited on the subsiding shelf is not clear. At some point it overlaps the transition from subsiding shelf (i.e. attenuated continental crust) to oceanic crust.

Exposed strata in the Halfway River map area span Late Proterozoic (Hadrynian) through Early Cretaceous time, and have an aggregate thickness of 8000 m. The map area straddles the facies transition from shelf carbonates to off-shelf clastics (Fig. 3). A thick succession of synorogenic sandstone, siltstone and shale of the clastic wedge assemblage is preserved across the eastern half of the map area.

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[1]Periods of block faulting in the Late Proterozoic and in the Devonian interrupted subsiding margin type deposition.
[2]This is a departure from the usage of Dietz and Holden (1966) who applied the term miogeocline to the continental terrace wedge only.

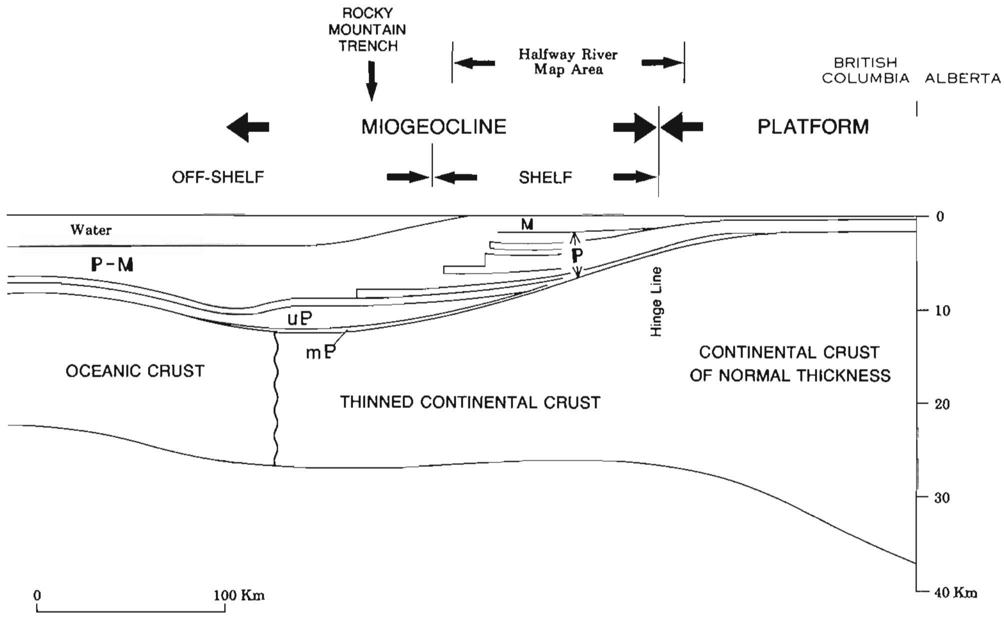


Figure 2. Diagrammatic model of a proto-Pacific continental margin showing distribution of platform, shelf and off-shelf facies relative to the extended crust of a passive, Atlantic-type, continental margin (5x vertical exaggeration).

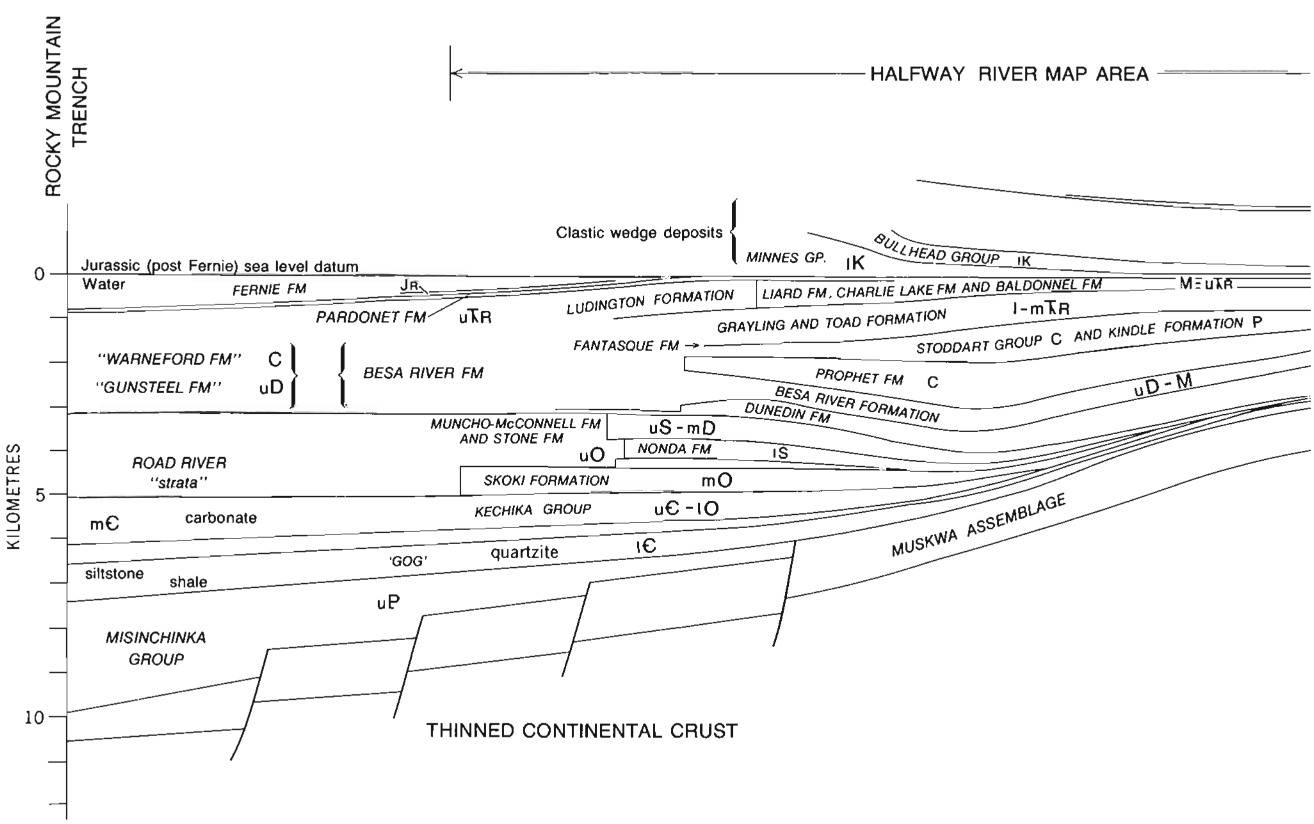


Figure 3. Generalized east-west cross-section showing the distribution of major lithostratigraphic units, the positions of major facies transitions, and the position of the Halfway River area in terms of the platform and miogeocline assemblages (5x vertical exaggeration).

In mid-Ordovician time, the boundary separating off-shelf from shelf facies extended southeastward across Ware (94 F) and Mesilinka (94 C) map areas into the west central part of the Halfway River map area, where it delineated the initial stage of the Ospika Embayment.[1] The embayment transgressed the shelf in a step-wise fashion until much of the western part of the map area was covered by off-shelf shale and siltstone (Fig. 4). The Peace River Arch, a westward projection of the carbonate shelf, formed the southern flank of the embayment. It persisted as a relatively positive, east-west trending tectonic feature, from mid-Cambrian through Late Devonian time, and played a major role in controlling sedimentary patterns across the miogeocline and a large segment of the platform.

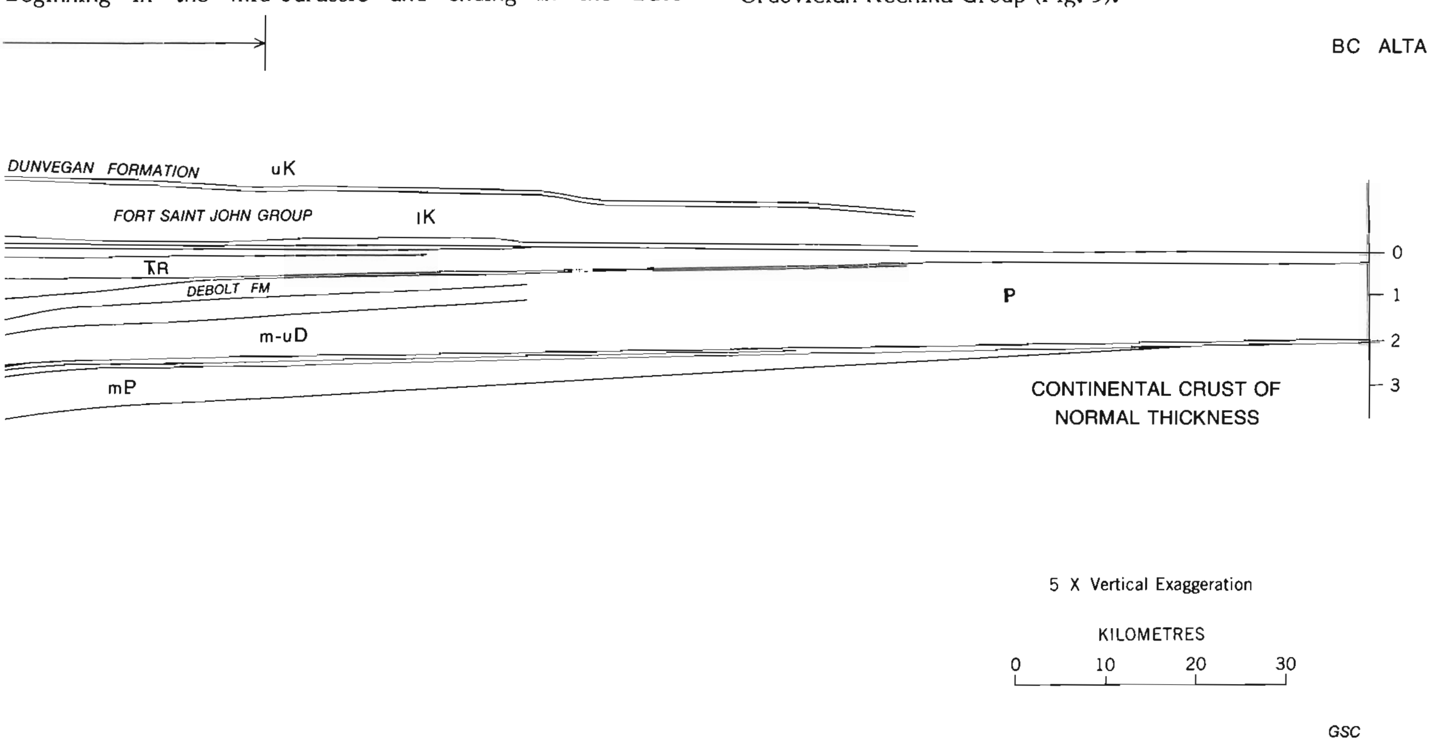
At the end of the Late Devonian, a major transgression of off-shelf facies pushed the limits of shelf carbonate deposition far to the east. Early in the Mississippian, when the next regressive pulse of shallow water carbonates overstepped the off-shelf belt, the Ospika Embayment and Peace River Arch were no longer in existence, and the transition from off-shelf clastic to shelf carbonate facies followed an undeflected southeast trend across the Halfway River map area. This depositional trend continued until Late Jurassic time, when the polarity of sediment source and transport was reversed, and large volumes of synorogenic clastic debris were shed from the then active Cordilleran orogen.

More than half of the Canadian Cordillera consists of volcanic, plutonic and sedimentary rocks that never were part of the miogeocline-clastic wedge prism (Fig. 5). This western "eugeosynclinal-type" region is a composite of smaller independent crustal blocks or terranes that evolved as separate "pieces" elsewhere on the Pacific plate. Beginning in the mid-Jurassic and ending in the Late Cretaceous or early Tertiary, these "pieces" collided with and were welded onto the western margin of ancestral North America (Monger and Price, 1979; Monger et al., 1982; Tempelman-Kluit, 1979; Fig. 5). Relative westward movement of the North American plate with respect to the ancient Pacific plate was a possible driving mechanism for deformation within the miogeocline; but details of the crustal mechanics have yet to be deciphered. Hence, the deformed miogeocline and overlying clastic wedge constitute only part of the history of Cordilleran evolution (Fig. 5).

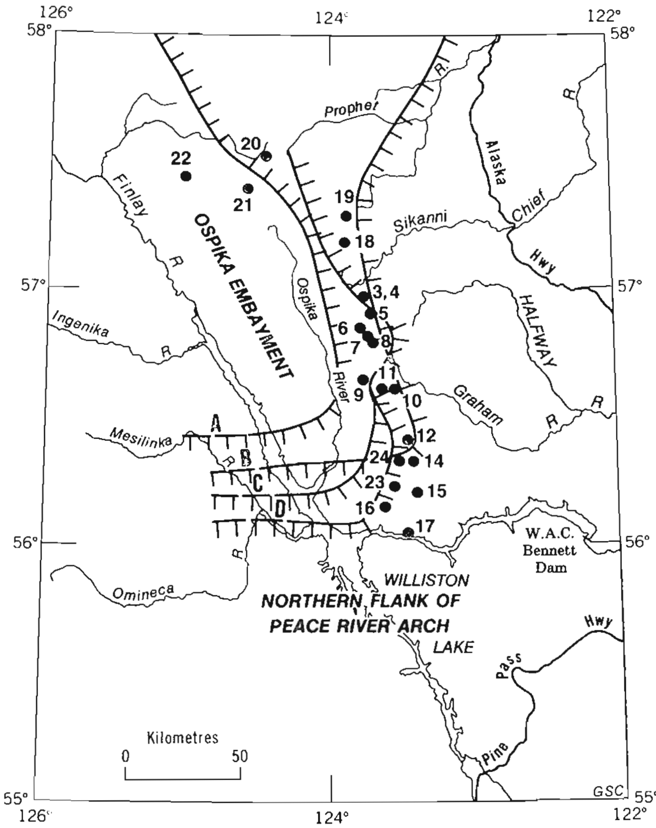
Depositional setting and essential stratigraphic characteristics of the Halfway River part of the miogeocline may be viewed in terms of three, gross, tectonostratigraphic packages, each of which records a final, regional, marine transgression of off-shelf facies. They are: Proterozoic through lower Paleozoic, mid-Ordovician through Upper Devonian, and Lower Mississippian through Upper Jurassic. The clastic wedge, the fourth tectonostratigraphic package, was deposited intermittently, from the Late Jurassic until early in the Tertiary. A description of each package and its stratigraphic link with adjacent parts of the eastern Cordillera follows.

Middle Proterozoic through Lower Ordovician

The oldest tectonostratigraphic package within the Halfway River map area embraces the Middle Proterozoic Muskwa assemblage, the Upper Proterozoic Windermere Supergroup, unnamed Lower and Middle Cambrian quartzite and dolomite, and the Upper Cambrian through Lower Ordovician Kechika Group (Fig. 3).

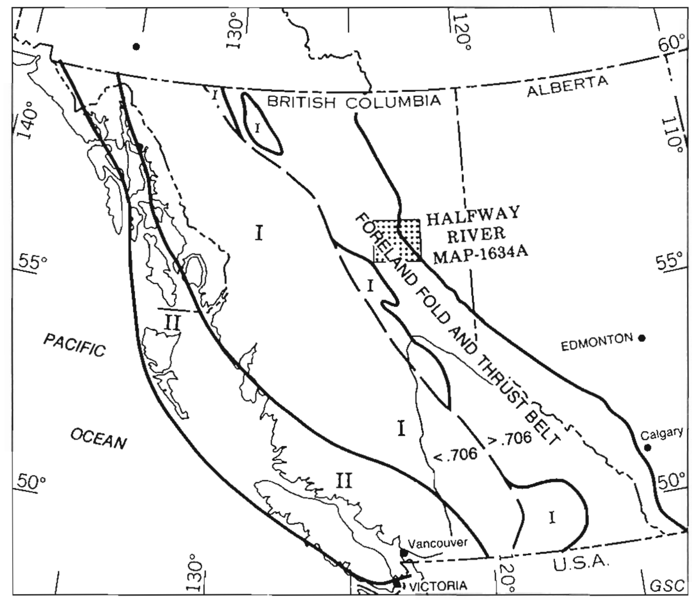


[1]This is the first published use of the term Ospika Embayment.



Mid Ordovician (SKOKI, from Cecile and Norford, 1979)............A
Mid-Late Devonian (DUNEDIN, from Taylor et al., 1975)............B
Late Silurian to Mid-Devonian (Muncho-McConnell and Stone).......C
Late Ordovician to Early Silurian (NONDA).......................D
Teeth on shelf side.................................| | | | |

Figure 4. *Summary plot of the lateral shifts through time (early and middle Paleozoic) of shelf to off-shelf facies transition. Palinspastic restoration of facies boundaries, now preserved in southern Halfway River map area, places them west of Williston Lake (northern Rocky Mountain Trench) in Mesilinka map area prior to Cretaceous folding and thrusting.*



SR^{87}/Sr^{86} strontium isotope ratio .... <.706 >.706

Distribution of allochthonous terranes that were accreted to the North American continent from mid-Jurassic to early Tertiary time................ II

Probable western limit of Precambrian continental crust that formed the pre-middle Jurassic North American margin (from Armstrong et al., 1977; Armstrong, 1979)..................

Figure 5. *The Canadian Cordillera (modified after Monger et al., 1982) showing the distribution of allochthonous terranes (I, II) that were accreted to the North American continent from mid-Jurassic to early Tertiary time. $^{87}Sr/^{86}Sr$ line (Armstrong et al., 1977, Armstrong, 1979) marks the probable western limit of Precambrian continental crust that formed the pre-mid-Jurassic North American margin.*

The Muskwa assemblage (Eisbacher, 1981) consists of epicratonic carbonate, shale, siltstone and sandstone deposited in a basin – possibly a re-entrant into the ancestral North American plate – that affected much of northeastern British Columbia. Basin shape and size are unknown, but it was much larger than the area of Muskwa strata now exposed in Tuchodi Lakes map area (94 K; Fig. 28). Seismic reflection data from the western margin of the Plains (physiographic) region in southeastern Fort Nelson map area (94 J), suggest more than 2000 m of pre-Middle Devonian strata, most of it Muskwa assemblage (Thompson, 1981). The 1000 m of Muskwa rocks inferred to underlie Halfway River map area (Fig. 3) are interpreted as stratigraphically thinned, flanking deposits near the southern limit of the depositional basin. This presumes that the basin was a re-entrant, similar in most aspects to the northern limit of the middle Proterozoic Belt - Purcell Basin in the Southern Rocky Mountains (McMechan and Price, 1982; Gabrielse, 1972).

Middle Proterozoic deposition lasted about 800 m.y. beginning approximately 1600 Ma and ending approximately 800 Ma. The portion of this time frame represented by the Muskwa assemblage is not known. Deposition in the Belt-Purcell Basin (southern Rocky Mountains) ended about 1300 Ma (McMechan and Price, 1982); in the Mackenzie and Wernecke mountains (Yukon Territory) deposition lasted until 800 Ma (Eisbacher, 1981).

Oldest rocks exposed within Halfway River map area belong to the Misinchinka Group, a thick succession of phyllite, gritty feldspathic sandstone, diamictite, conglomerate, limestone and dolostone of Late Proterozoic age that forms part of a thrust sheet within the southwestern corner of the Halfway River map area. The Misinchinka Group forms part of a distinctive lithostratigraphic interval called the Windermere Supergroup (Eisbacher, 1981), an assemblage that onlaps the western margins of middle Proterozoic deposits, and presumably, Precambrian shield along those portions of the North American plate margin where Middle Proterozoic strata were not deposited.

Unlike the mature epicratonic facies that typified Middle Proterozoic deposition, Upper Proterozoic strata reflect a depositional environment that included immediate source regions, substantial local relief, and rapid along-strike changes in basin morphology. Some conglomerate and diamictite contain clasts eroded from nearby scarps of middle Proterozoic strata (Lis and Price, 1976). Feldspar is a constituent of gritty sandstone successions, suggesting proximity to granitoid rocks, as do cobbles and boulders of coarse grained felsic intrusives (Monkman Pass map area). The stratigraphic record suggests that Windermere deposition reflects a protracted period of crustal instability, lasting approximately from 800 to 570 Ma, during which the craton margin underwent distension accompanied by block faulting.

Exposures of the Misinchinka Group in and adjacent to the Halfway River map area contain all of the main lithofacies elements typical of Windermere deposition. Slate and phyllite interbedded with diamict conglomerate and impure, feldspathic, gritty quartzite (Gabrielse, 1975) are overlain by thick, phyllitic slate containing a prominent carbonate marker.[1]

During latest Proterozoic time, the craton margin gradually regained its pre-Windermere stability; an epicratonic subsiding shelf was established that extended more than 50 km west of the Halfway River map area and resulted in the westward progradation of a thick blanket of shield-derived quartz sand and silt. This accumulation of mature quartz sandstones, siltstones and orthoquartzites continued until the Middle Cambrian, by which time a remarkably uniform assemblage of clastic rocks up to 3 km thick covered the eastern part of the Cordilleran orogen. In the Rocky Mountains, this succession is called the Gog Group, and it may be traced without interruption from Banff north to the Halfway River area, into Mesilinka (94 C) and Ware (94 F) map areas and possibly through Tuchodi Lakes (94 K) map area (Gabrielse, pers. comm., 1980). In the Mackenzie Mountains, a similar succession is called the Backbone Ranges Formation. West of the Rocky Mountain Trench, equivalent lithofacies of Eocambrian and Early Cambrian age are identified by a variety of names depending on location – Cranbrook and Hamill in the south, Yanks Peak in the Cariboo Mountains, and Atan in the north.

Only poorly exposed, thrust-faulted slivers of Gog quartzite are present in the southwest corner of the Halfway River area; however, west of the Ospika River in Mesilinka map area, more than 650 m of Lower Cambrian quartz sandstone, sandstone, and shale with minor dolostone and sandy dolostone beds, is exposed (Irish, 1970; Gabrielse, 1975).

By Middle Cambrian time, the supply of clastic material had dwindled. Gog deposition ceased, and the miogeocline became a typical, subsiding shelf comprising an eastern shelf-carbonate belt of clean, shallow water limestone and dolostone, and a western, off-shelf region of deeper water, shaly limestone and calcareous shale. Separating the two facies belts were reefs and shoals that acted as a huge, semipermeable barrier (called the Kicking Horse Rim in the southern Rockies) and protected the shelf carbonates from open marine conditions (Aitken, 1966, 1971). Except for major marine transgressions of deeper water detrital facies during the Late Cambrian and Early Ordovician, and again during the Late Devonian and Early Mississippian, this pattern of facies distribution within the miogeocline was the norm for most of Paleozoic time. The position of the barrier shifted through time, and the amount of detrital material deposited within the off-shelf region varied, but this basic tectonostratigraphic framework persisted.

As with the Lower Cambrian Gog Group, only thin, poorly exposed, folded thrust slices of unnamed Middle Cambrian dolostone are exposed in the southwestern corner of the Halfway River sheet. West of the Ospika River, in Fort Grahame map area, up to 350 m of dolostone, sandy dolostone and sandstone are exposed.

Elements of an off-shelf facies belt are present in the northwest of Ware map area, but the geometry and nature of the facies transition is not well displayed, and will require additional investigation (Gabrielse, pers. comm., 1980).

The variation in thickness of Lower and Middle Cambrian strata shown in Figure 3 is based on the assumption that these strata thicken uniformly westward from a zero depositional edge, a few kilometres east of the British Columbia – Alberta boundary (Pugh, 1973, 1975), to 1000 m at the western margin of the Halfway River map area (Gabrielse, 1975).

The 850+ m of Lower through Middle Cambrian strata in the Halfway River – Mesilinka map areas is significantly less than the thickness of equivalent strata within the Main Ranges of the Southern Rockies (see Cook et al., 1975 for a regional comparison of Cambrian isopachs). This suggests that the Peace River Arch remained a relatively positive feature during most of Cambrian time, thereby limiting the thickness of strata that accumulated across the Halfway and Mesilinka map areas. Pugh's work (1973, 1975) demonstrates that the arch intermittently emerged farther to the east.

Subsidence, accompanied by flooding of the outer margin of the carbonate shelf, began during the Late Cambrian. By Early Ordovician time, much of the shelf was blanketed by a thick succession of argillaceous, thin-bedded limestone, siltstone and shale of the Kechika Group and its southern correlatives, the upper Lynx and Chushina formations (central Rockies), and the McKay Group (southern Rockies). In the Halfway River map area, the Kechika Formation is at least 1000 m thick and consists of a monotonous succession of phyllitic noncalcareous shale, wavy banded limestone, and calcareous siltstone and shale. Mappable subdivisions were not recognized, but to the north in Trutch and Ware map areas, Cecile and Norford (1979) recognized five mappable units, all of which thin depositionally eastward from a total thickness of more than 1800 m in the west to 250 m at the mountain front. This rate of eastward depositional thinning suggests that little or no laterally equivalent platform sequence can be expected within the subsurface farther east.

Lower Ordovician through Middle Devonian

Toward the end of the Early Ordovician (late Canadian), shelf carbonates built westward, covering the Kechika Group with up to 1200 m of massive, light coloured dolostone and limestone (Fig. 6). The broad carbonate terrace remained stable until late in the Middle Ordovician (early Caradocian) and, on the basis of lithological similarity and time equivalence (B.S. Norford, pers. comm., 1975; Thompson, 1976; Cecile and Norford, 1979), is tentatively correlated with the Skoki Formation of the southern and central Rocky Mountains (Norford, 1969). The Skoki Formation extends north and west into Trutch (94 G), Mesilinka (94 C) and Ware (94 F) map areas (Gabrielse, 1975, 1981; Taylor, 1979) and south into Pine Pass (93 O) map area, where it links with Skoki traced from the southern Rockies. Synchronous with Skoki deposition, a conformable succession of argillaceous limestone, shale and slope breccias accumulated within the off-shelf belt (Cecile and Norford, 1979). This succession forms the basal part of Road River strata, a fine grained, clastic rock assemblage that ranges in age from Early Ordovician to Middle Devonian, and is recognized from Halfway River map area northwestward to and beyond the Selwyn Basin of the Yukon Territory (Gordey, 1981; Cecile, 1982).

[1]This carbonate marker may correlate with the Byng Formation (Taylor, 1983).

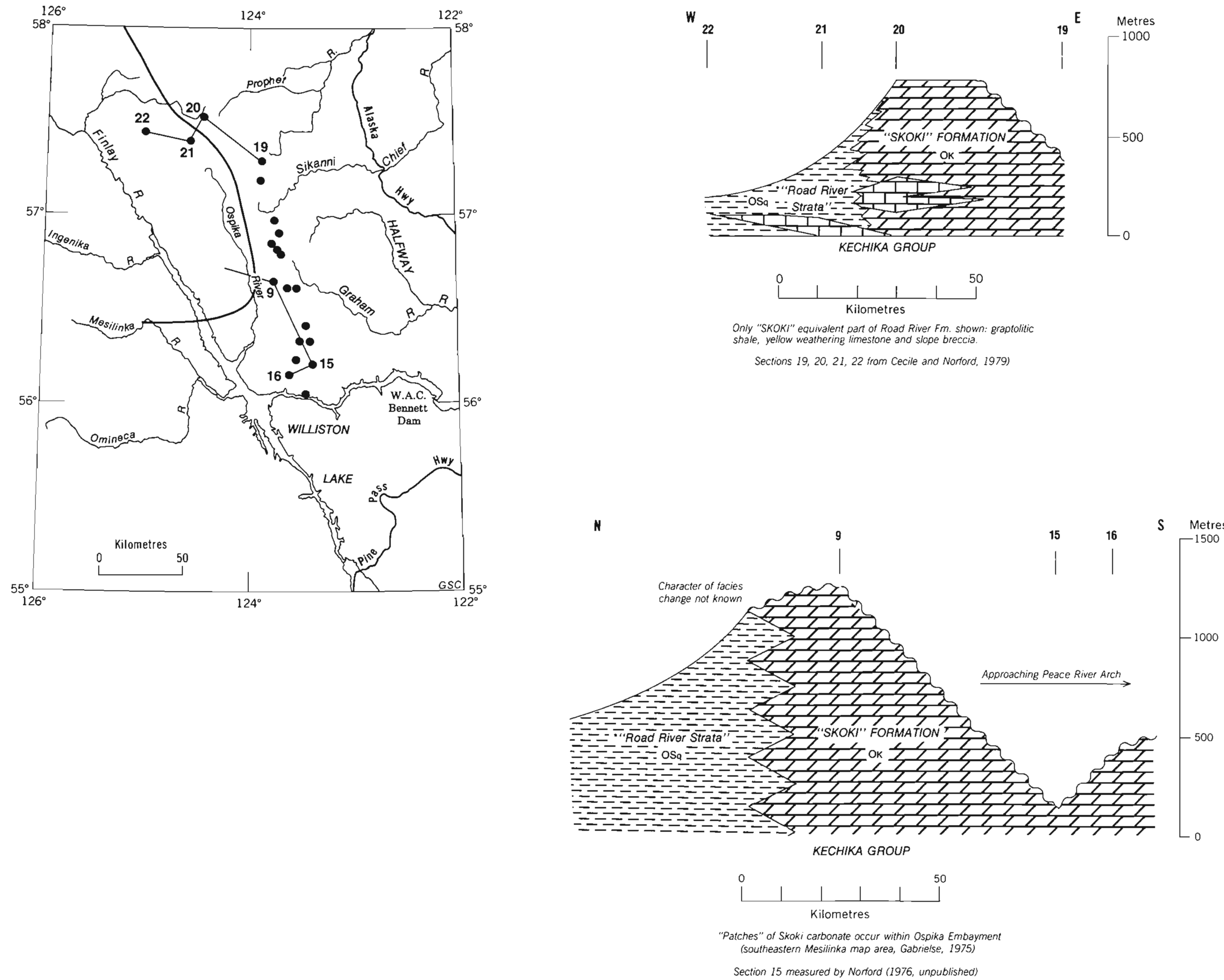


Figure 6. *The distribution of Middle Ordovician (Skoki and Skoki equivalent) shelf and off-shelf facies; stratigraphic cross-sections through the facies transition show the relative thicknesses of map units (refer to Appendix 2 for symbols legend).*

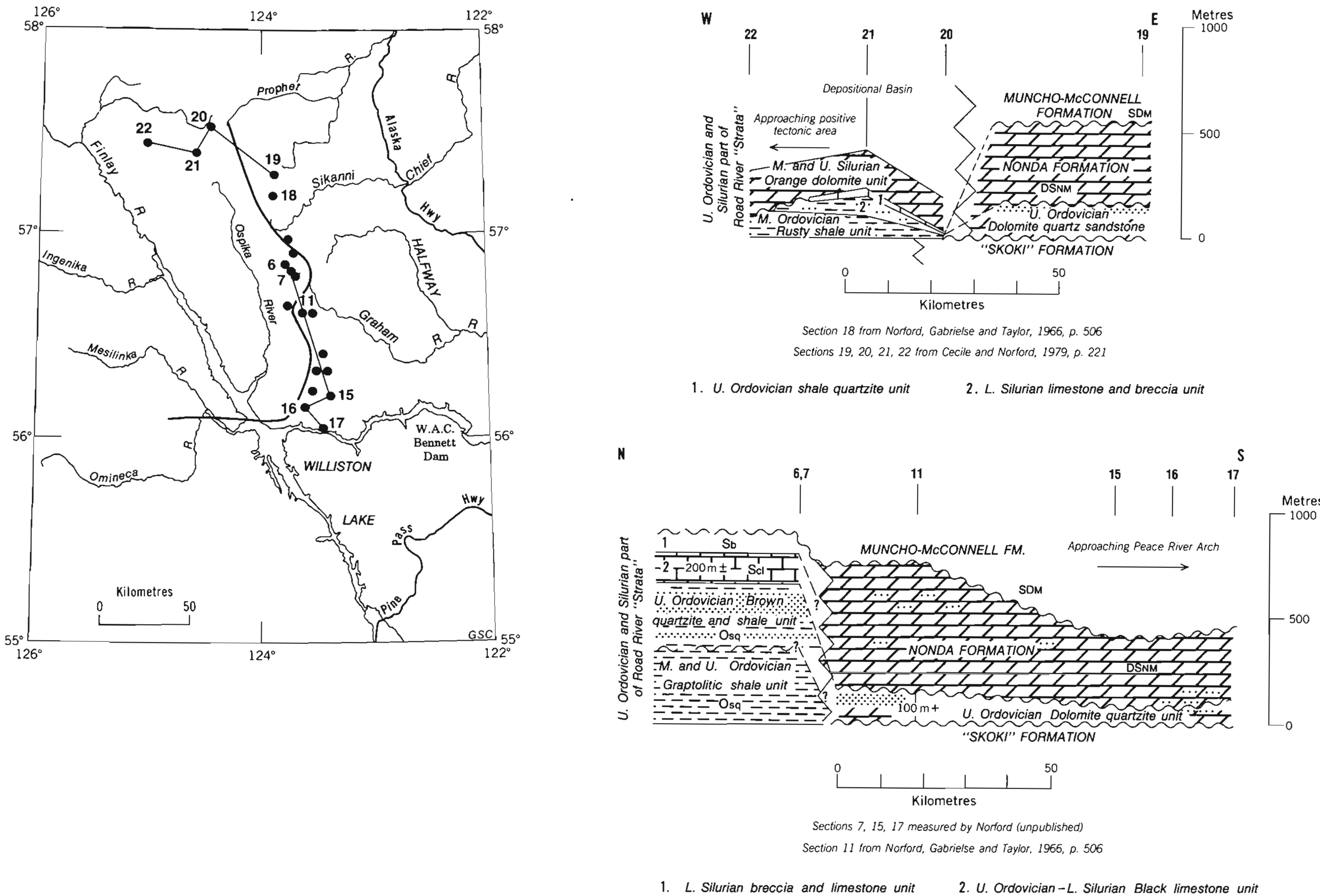


Figure 7. *The distribution of Upper Ordovician through mid-Silurian shelf and off-shelf facies; stratigraphic cross-sections through the facies transition show the relative thicknesses of stratigraphic units (refer to Appendix 2 for legend of symbols).*

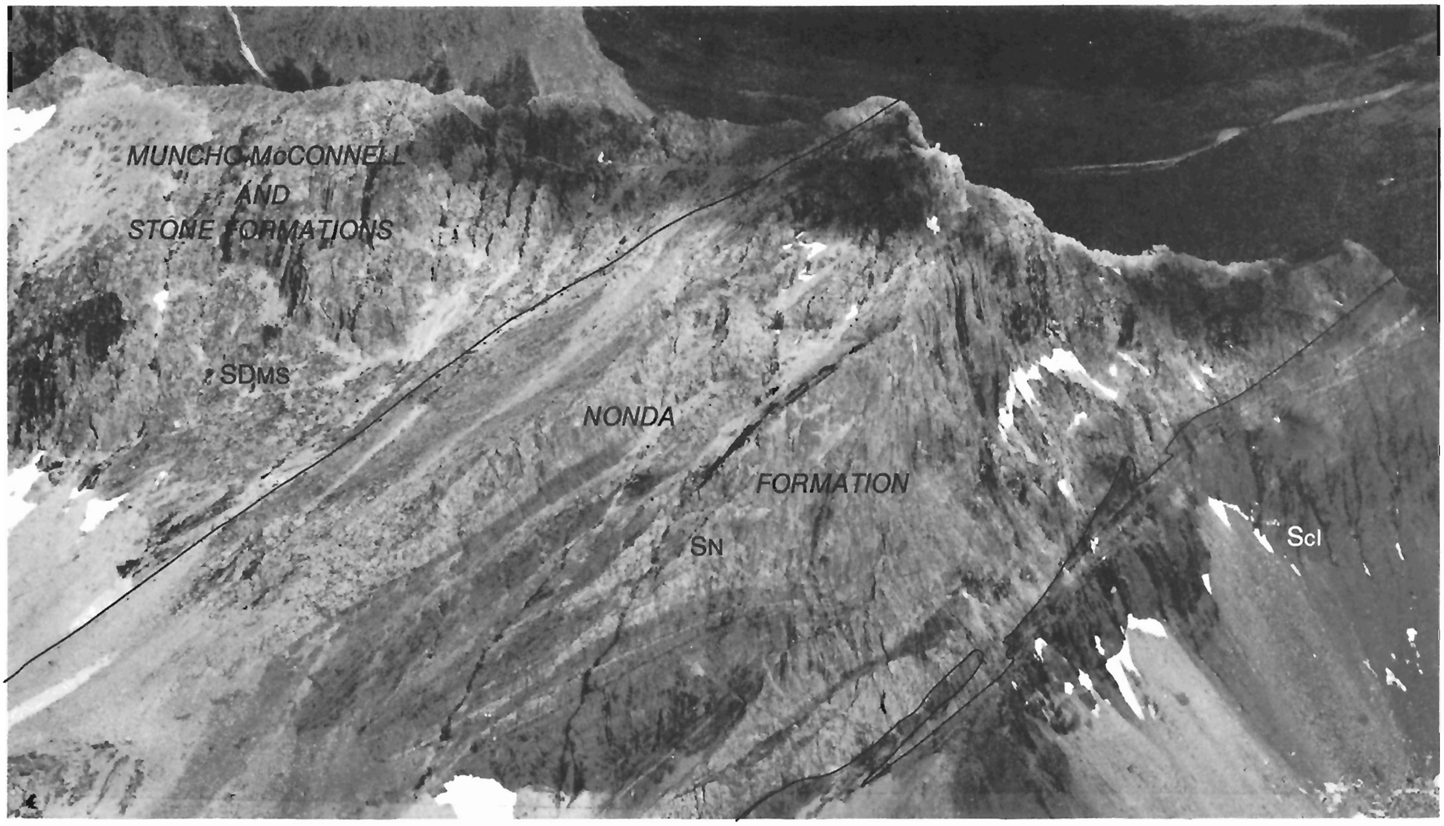


Figure 8. *View toward the northwest showing the facies transition between Silurian shelf (Nonda Formation,* SN *) and off-shelf facies. Tongues of carbonate debris are interlayered with off-shelf carbonaceous limestone along the foreslope of a Nonda Formation reef (5 km north of Mount Kenny; see also Norford, Gabrielse and Taylor, 1966). The younger Muncho–McConnell and Stone formations (* SDMS *; upper left of photo) built westward, across the shelf and off-shelf facies transition within the Nonda Formation. (GSC photo no. 819-20).*

The facies boundary between the Skoki Formation and the basal part of Road River strata outlines the Ospika Embayment, now represented by a southeastward-closing lobe of off-shelf facies rocks bounded on the south by the westward extension, perpendicular to the craton margin, of the Peace River Arch and its cover of shelf carbonate facies.

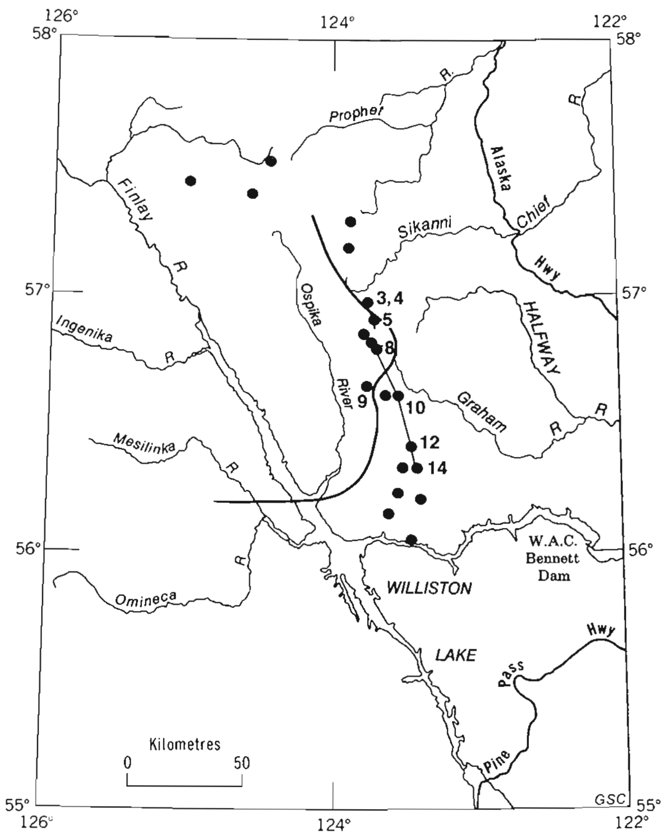
Skoki deposition marked the maximum extent of the carbonate shelf, and came to an end with a period of erosion (Rocklandian through Edenian) followed by incursion of off-shelf facies. During the Late Ordovician, the Ospika Embayment reached farther into the map area, pushing the limits of carbonate deposition to the south and east. This step-wise transgression across the carbonate shelf occurred several times during the evolution of the embayment. Each transgression was accompanied by the formation of a new carbonate terrace (Figs. 6, 7).

The Upper Ordovician part of the succession (Fig. 7) is unnamed, and comprises a lower limestone, shale, and dolostone, and an upper dolomitic sandstone and sandy dolostone. The lower part is a time equivalent of the Mount Wilson Formation, a quartzite unit found in the Southern Rockies, but the two units do not correspond lithologically. The upper unit, although time equivalent to the massive, light dolostone of the Beaverfoot Formation (Norford, 1969) found south of Pine Pass map area, does not bear a close lithological resemblance to it. The unnamed Upper Ordovician units are exposed only in the southwestern part of the Halfway River map area (and adjacent parts of Pine Pass), and the western part of the Trutch map area (Cecile and Norford, 1979). The Nonda Formation, a thick, Lower Silurian (Llandovery) dolostone, comprises the upper part of the carbonate shelf facies, and outcrops continuously from Pine Pass northward along the mountain front across Halfway River, Trutch, and Tuchodi Lakes map areas.

The relationship of one stratigraphic facies to another is normally apparent only after much of the regional mapping and stratigraphic work is completed – rarely is a facies transition well exposed. An exception is shown in Figure 8 where, on a single mountainside, the Nonda Formation can be observed changing laterally from shelf carbonate facies to the off-shelf facies of Road River strata.

Farther west, away from the reef front, the Upper Ordovician through Lower Silurian part of Road River strata comprises three units (Fig. 7): an Upper Ordovician, graptolitic, shale-quartzite unit that rests on Skoki carbonates and correlates with the unnamed Upper Ordovician carbonate units; a thin bedded, carbonaceous limestone unit of Late Ordovician through Early Silurian age that correlates, in part at least, with the Nonda Formation; and an upper breccia unit containing clasts of Nonda rock types.

The succeeding carbonate terrace (Fig. 9) is composed of a thick succession of light grey, clean, homogeneous dolostone and sandy dolostone of Middle Silurian to Middle Devonian age. It consists of two formations, the Muncho-McConnell and the overlying Stone. Together, these two formations form massive, steep cliffs near the mountain front. The platform succession extends from the northern part of Monkman Pass map area through Pine Pass, Halfway River, Trutch, and Tuchodi Lakes map areas and into Toad River map area. North of the Halfway River region, a thin



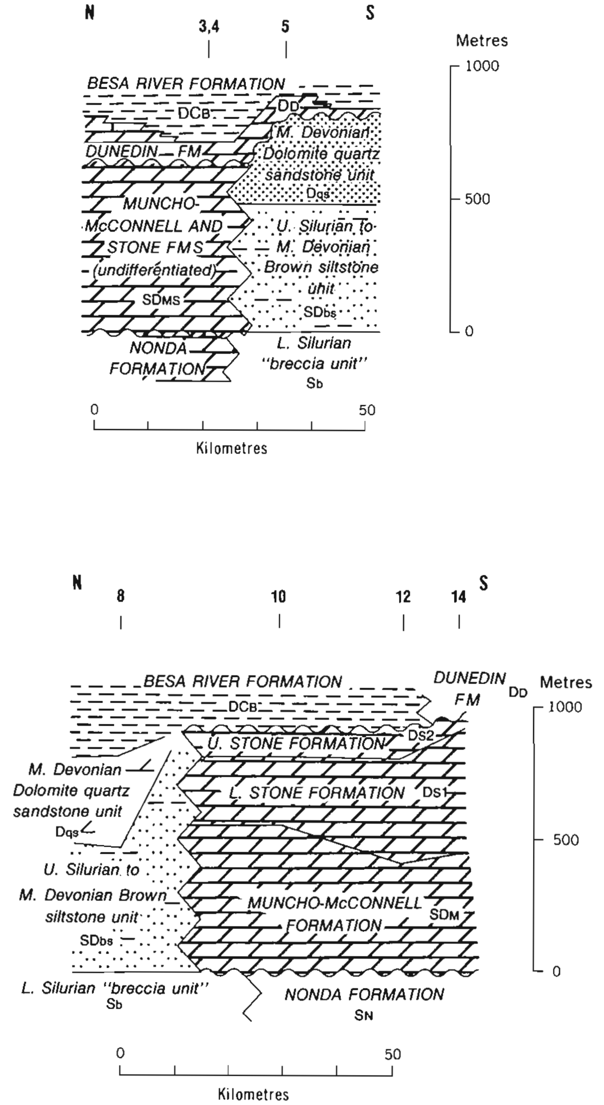


Figure 9. *The distribution of Upper Silurian through mid-Devonian shelf and off-shelf facies; stratigraphic sections through the facies transition illustrate the relative thicknesses of stratigraphic units (refer to Appendix 2 for legend of symbols).*

sandstone unit called the Wokkpash Formation separates the Muncho-McConnell from the Stone Formation (Taylor and Mackenzie, 1970).

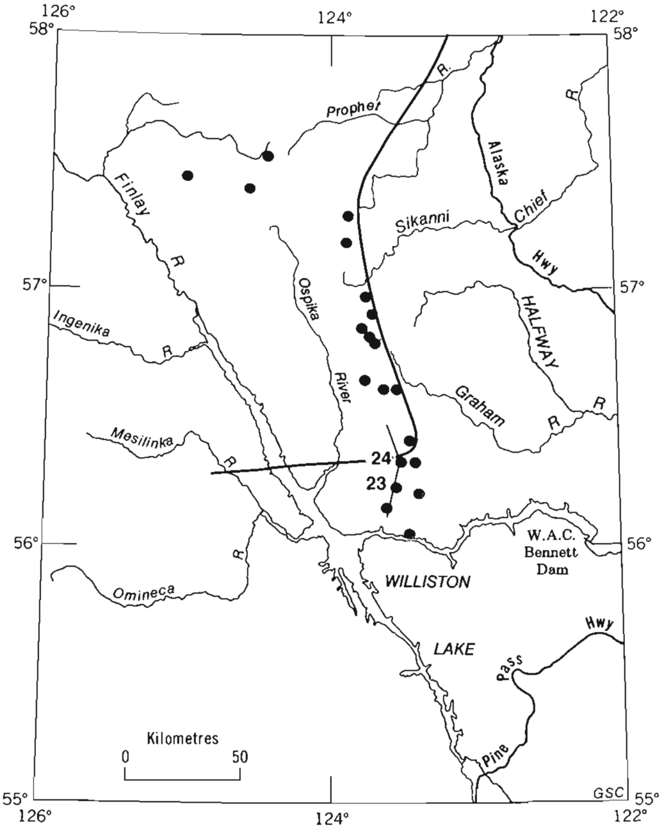
The Muncho-McConnell and Stone carbonate terrace built across the Nonda terrace edge, extending briefly the limits of carbonate shelf deposition within the Halfway River region (Fig. 9). To the west, the Road River off-shelf facies is called the brown siltstone unit; it consists of calcareous and noncalcareous siltstone, shale and thin carbonate debris flows. The brown siltstone unit is capped by a distinctive, dolomitic quartz sandstone unit. This upper part of Road River strata is exposed in a small structural salient called the Laurier Anticlinorium, in the northwestern part of the Halfway River map area.

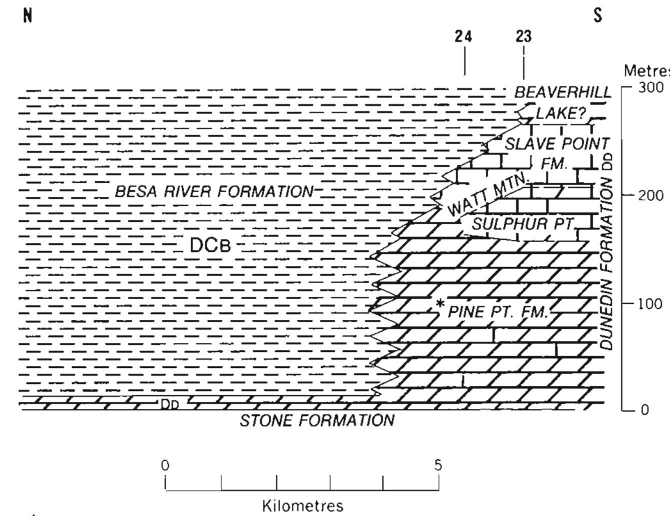
Davies (1966) suggested that the Road River brown siltstone unit lay beneath the Muncho-McConnell Formation – an interpretation based on graptolite collections of Early and Middle Silurian age. Coral collections from other parts of the brown siltstone unit range in age from Early Devonian to Middle Devonian (A.E.H. Pedder, pers. comm., 1977). Hence time equivalence, in part at least, between shelf carbonate and off-shelf detrital facies is demonstrated.

A brief period of erosion followed Stone deposition before the region underwent the initial phase of a protracted period of subsidence that lasted until Early Mississippian time, and during which off-shelf facies slowly transgressed the carbonate shelf terraces. In all but the southern portion of the Halfway River map area, the Stone Formation is overlain by a thin, dark, medium bedded dolostone and limestone called the Dunedin Formation. It is a transgressive shelf carbonate sequence (Morrow, 1978) that extends from Toad River map area southward across Tuchodi Lakes and Trutch map areas into the Halfway River map area, and westward into Mesilinka and Ware map areas.

The Dunedin Formation is overstepped by shales of the Besa River Formation, which transgressed the Dunedin shelf and pushed the carbonate shelf edge east of the present mountain front (Fig. 10).

The Peace River Arch continued to influence depositional patterns during the Middle and Late Devonian incursion of clastics across the carbonate shelf. A few kilometres north of Mount Burden, the carbonate shelf-edge can be traced westward to the limit of Devonian outcrop. Palinspastic restoration places these outcrops west of the Rocky Mountain Trench, indicating that the Arch projected across the Trench until latest Middle Devonian time.





*Subsurface terminology applied to Dunedin Formation by Macqueen and Taylor, 1974

Sections 23 and 24 from Macqueen and Taylor, 1974

Figure 10. The distribution of Middle and Upper Devonian shelf and off-shelf facies; a stratigraphic section through the facies transition illustrates the relative thicknesses of stratigraphic units (refer to Appendix 2 for legend of symbols).

Finally, regional subsidence of the shelf was accompanied by eastward flooding of the Besa River shale, completing the marine transgression that had begun with Dunedin sedimentation.

Upper Devonian through Lower Jurassic

During the latter part of the Late Devonian (Famennian) and into the Mississippian (Tournaisian), the Halfway River map area was blanketed by Besa River shale. Even the Peace River Arch, which had stood as a relative high for 200 million years, shrank eastward, and had less and less influence on facies distributions and patterns. By latest Devonian time, most of the western part of the platform and eastern part of the shelf was covered by a thin, black shale (the Exshaw Formation) and a succeeding dark, argillaceous limestone succession (the Banff Formation). Near the beginning of Late Tournaisian time, a shallow water carbonate platform succession was re-established across most of the interior platform regions and the southern part of the shelf (e.g., Rundle Group). By the early Viséan, this shallowing-upward, progradational-aggradational succession had established itself across the Foothills region of the Halfway River map area, where it is now called the Prophet Formation. The western limit of the Prophet carbonate shelf follows a north-south trend across Tuchodi Lakes, Trutch, Halfway and Pine Pass map areas (Bamber and Macqueen, 1979) with the thickest accumulation of carbonate occurring across the former site of the Peace River Arch (at this time an embayment). Argillaceous limestone and chert (containing sponge spicules) are important constituents of the Prophet Formation within Halfway River map area and support the interpretation that it is a distal, deeper water facies of the carbonate shelf (Bamber and Macqueen, *ibid.*). More proximal facies are encountered to the east in the subsurface and along strike south of Pine Pass, where shelf margin and restricted shelf carbonate facies predominate; here the carbonate succession is subdivided into four formations: the Pekisko, Shunda, Turner Valley, and Mount Head (Mamet et al., in press).

Deposition of the Prophet Formation ceased near the beginning of the late Viséan with the influx, from the east, of terrigenous clastics belonging to the Stoddart Group. This clastic succession blanketed the region north of Pine Pass, extending across the shelf and into the off-shelf facies belt where it interfingered with the Besa River Formation. In the subsurface east of the outcrop belt, the Stoddart Group comprises three formations – Golata, Kiskatinaw, and Taylor Flat – consisting respectively of shale, sandstone, and limestone rock units. These rock types are difficult to distinguish in surface exposures. The Stoddart Group is laterally equivalent to a thick sandstone succession within the southern Yukon Territory called the Mattson Formation. Between Pine Pass and Mount Robson map areas, strata of equivalent age have been removed by erosion; farther south, peritidal carbonates and terrigenous clastics of the Mount Head and Etherington formations (Rundle Group) range in age from late early Viséan through early Namurian.

Rocks of Late Carboniferous age are essentially absent, except for isolated occurrences in the southern Pine Pass and Monkman Pass map areas and farther south in the Banff area (Bamber and Macqueen, 1979). Nearly all of the carbonate shelf succession was affected by sub-Permian and sub-Moscovian disconformities, which truncate the Upper Carboniferous succession, and in places, cut deeply into the Lower Carboniferous.

The Permian of the northern Rockies is represented by a lower siltstone-shale-sandstone-dolomitic siltstone-chert succession called the Kindle Formation, and an upper, distinctive chert marker unit called the Fantasque Formation. The Kindle Formation was not mapped separately in the Halfway River area and, where present, has been included with the Stoddart Group. The Fantasque chert can be traced throughout most of the northern Rockies. In the

Halfway River region it consists of 15 m of black chert, and stands out as a distinctive, thin rib that is very useful for mapping details of the structure. South of Pine Pass map area, the equivalent Formation is called the Ranger Canyon, and the chert is lighter coloured, contains some dolostone as well as variable amounts of siltstone, and may be phosphatic (McGugan and Rapson, 1963; Bamber and Macqueen, 1979).

The Fantasque chert is overlain by a thin, phosphate pebble conglomerate that was deposited during an interval of erosion at the end of the Permian and/or beginning of the Triassic (Gibson, 1975).

Two main depositional phases are recorded in the Triassic lithofacies: an early, transgressive, clastic facies characterized by shale and siltstone, and deposited below wave base, followed by a regressive, shallow-water, clastic-evaporitic facies (Gibson, 1975; 1971). The lower clastic succession comprises the Grayling and Toad formations, which can be mapped southward from Tuchodi Lakes across Trutch and Halfway River map areas into the northern part of Pine Pass map area. South of Pine Pass, the same time-stratigraphic interval corresponds to all but the upper part (Llama member) of the Sulphur Mountain Formation. Open marine conditions, favouring deposition of detrital facies, persisted until late in the Middle Triassic (Ladinian) when a progradational, shallow-water, crosslaminated sandstone facies spread southwestward across the finer grained, more carbonaceous Grayling and Toad formations (Pelletier, 1964). North of Pine Pass, surface exposures of this sandstone succession are called the Liard Formation. South of Pine Pass the sandstone is too fine grained to be mapped separately and is included as the upper member (Llama) of the Sulphur Mountain Formation. To the east, in the subsurface, the correlative sandstone unit is called the Halfway Formation.

Following deposition of the Liard sandstone, a thick carbonate succession called the Ludington Formation was established near the beginning of the Carnian. To the east, the shelf consisted of shallower-water carbonates and evaporites of the Charlie Lake Formation, succeeded by slightly deeper water, bioclastic carbonates of the Baldonnel Formation. Circulation across this already shallow-water platform became even more restricted at the end of Baldonnel deposition, and the resulting euxinic conditions are reflected by a facies change to coquinoid banks of thin, fragmented, pelecypod shells that form the basal part of the Pardonet Formation. Sea level rose during Pardonet deposition, and this facies can be mapped westward across the facies boundary between the Ludington and Baldonnel formations.

The youngest Triassic Formation, the Bocock, occurs in parts of northern Pine Pass map area and records a last regressive phase of shallow-water carbonate deposition. The areal extent of the Bocock Formation was reduced by a period of pre-Jurassic erosion.

The Bocock Formation is the last record of shelf carbonate deposition across the eastern part of the miogeocline. Beginning in the Early Jurassic (Sinemurian) a shale-siltstone succession called the Fernie Formation transgressed the shelf. Its uppermost part contains numerous sandstone turbidite beds that grade upward into shoreface deposits (known as the "Passage Beds" in the southern Rockies). This transition records the inception of a fundamental change in the crustal dynamics affecting the craton margin. The passive Atlantic-type margin became the site of crustal convergence (Monger and Price, 1979) and consequent Alpine-type tectonism. Much of the shelf region was transformed into a linear foreland basin. The craton and eastern portion of the miogeocline became inland seas that received detritus largely from the emerging source areas to the west. Miogeoclinal deposition was at an end.

Upper Jurassic through lower Tertiary

Relationship of foreland basin deposition to orogenic evolution of the Eastern Cordillera

Sandstone beds within the upper part of the Fernie Formation are the first stratigraphic evidence within the eastern part of the Rocky Mountains that a fundamental change in the physical character of the craton margin was taking place. The sandstones are composed of detritus eroded from a mountainous terrain that began forming during the initial stages of a protracted and intermittently active orogeny lasting from the Late Jurassic until early in the Tertiary.

Much of the orogenic history of the Cordillera has been deciphered through careful study of the stratigraphic and sedimentological character of synorogenic clastic rocks deposited within and adjacent to the evolving mountain belt (Wheeler, 1961; Eisbacher et al., 1974; Rapson, 1965; Stott, 1975; McLean, 1976). Detritus eroded from emerging highland areas was deposited in troughs marginal to terranes and volcanic arcs, as clastic sequences in successor basins overlapping older terranes, and into a linear trough called the foreland basin, which formed along the eastern flank of the orogen.

Foreland basins are an essential component of an evolving mountain system such as the Cordillera, and a direct cause and effect relationship exists between the tectonic processes responsible for deformation within the supra-crustal (miogeoclinal) rocks and the development of a basin peripheral to the deforming belt in which the clastic wedge can accumulate and be preserved. During the progressive eastward spread of deformation across the miogeocline, off-shelf and shelf facies rocks are stripped from the underlying lithosphere basement and piled into a tectonically thickened succession of imbricated nappes and thrust sheets. The increased supracrustal load induces a downward flexure of the lithosphere which, because of its substantial thickness and viscoelastic properties, is depressed over a much broader area than that occupied by the superjacent load (in the same manner that an elastic beam responds to a localized load by bending over a much greater arc length than the area of the loaded surface). Consequently, detritus eroded from the tectonically thickening mountain belt is trapped within the adjacent peripheral part of the lithosphere downwarp – the foreland basin. The size and depth of the foreland basin is determined by the position and mass of the tectonic load, and as the load migrates with the eastward propagation of thrusts and folds, so does the foreland basin axis.

Isostatic response of the lithosphere to loading and unloading is essentially instantaneous within a geological time frame, making stratigraphic relations within the clastic wedge a sensitive indicator of the timing of tectonic pulses and, to a lesser extent, the severity of each pulse (Fig. 11). During Stage 1 (Fig. 11a, b), it is assumed that the rate of tectonic thickening of the supracrustal rocks exceeds the rate of denudation; mountains form at the same time as the lithosphere subsides (isostatically) beneath the increased load, increasing the size and depth of the foreland basin as well as the supply of synorogenic sediment available to fill it. During Stage 2 (Fig. 11c, d), deformation slows to an eventual stop, permitting denudation to outpace the rate of tectonic

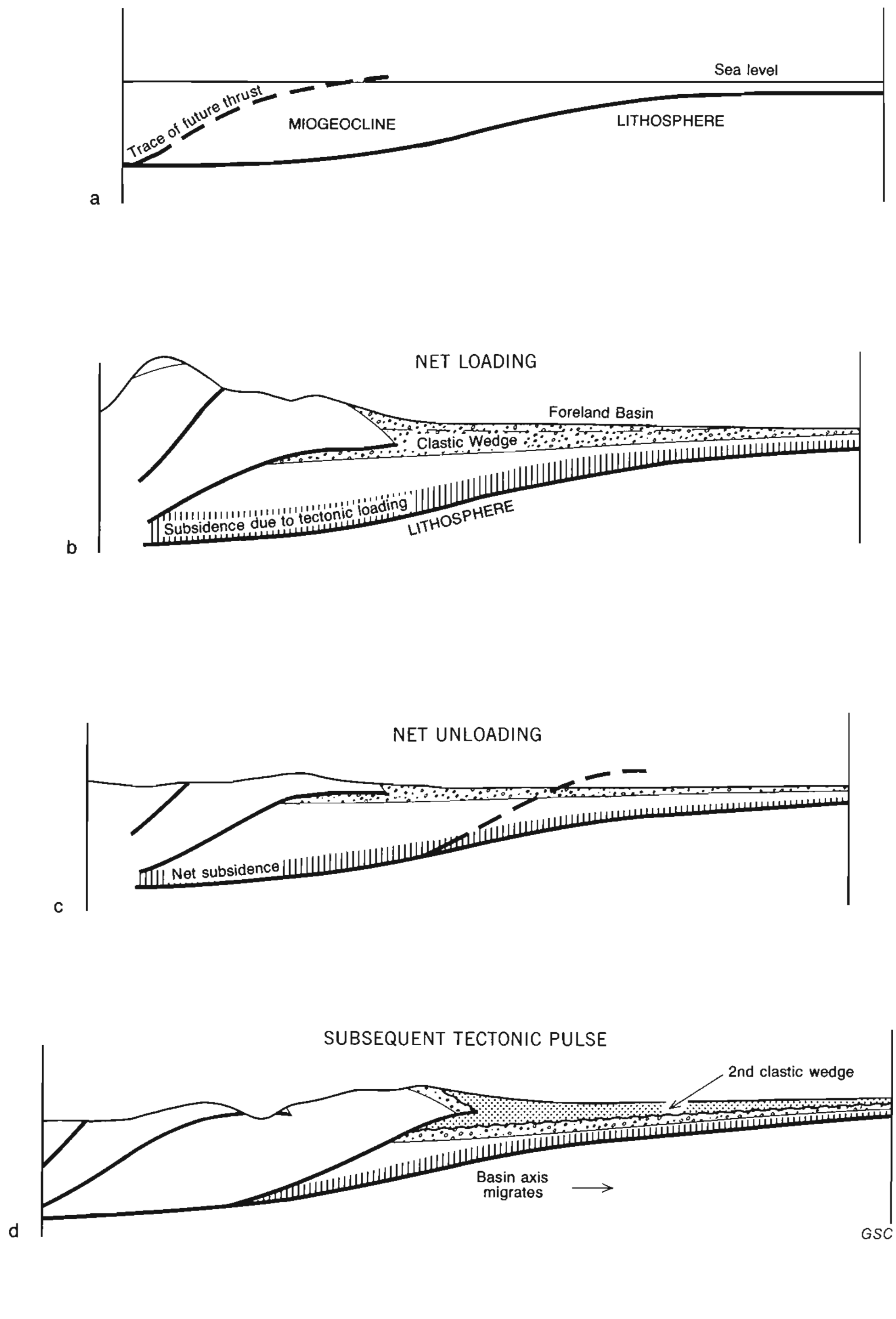


Figure 11. *Schematic diagram showing the relationships between tectonic transport, lithosphere subsidence and clastic wedge deposition: (a) shape of miogeocline at onset of deformation; (b) emplacement of thrust sheet causes net subsidence of the lithosphere and traps a clastic wedge adjacent to the thrust belt; (c) cessation of tectonism and continued erosion of the thrust belt causes net unloading and lithosphere rebound, part of the clastic wedge succession is eroded; (d) renewed tectonism and advance of the thrust belt to the right causes net subsidence of the lithosphere and deposition of a second clastic wedge.*

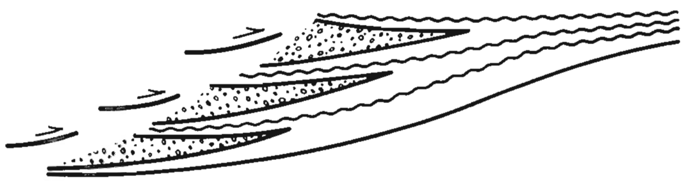


Figure 12. Idealized cross-section (modified from Royse et al., 1975) showing three, separate clastic wedges separated by unconformities. Each clastic wedge represents a tectonic pulse; each unconformity represents a period of tectonic quiescence. Migration of the tectonic front from left to right effects a similar migration of coarse proximal facies within each clastic wedge.

thickening; the result is a net decrease in the supracrustal load accompanied by rebound of the lithosphere (i.e. maintenance of isostatic equilibrium). Hence the size and depth of the foreland basin is reduced and the upper part of the clastic wedge is exposed to erosion. This cycle, repeated several times, results in a succession of unconformity-bounded, off-lapping clastic wedges, in which each wedge is related to vigorous tectonic activity and each unconformity to an interval of tectonic quiescence (Fig. 12).

The closely linked relationship between deformation, erosion and clastic wedge deposition for the southern part of the Canadian Rockies was described qualitatively by Price and Mountjoy (1970), and Price (1973). Using their approach, Beaumont (1981) has presented a quantitative model analysis in which he demonstrates the relationship between evolution of the foreland basin and its adjacent mountain belt through lithospheric flexure.

Within the southern Rockies, the clastic wedge contains three, distinct, off-lapping nonmarine and marine clastic successions, referred to here as the Kootenay, Blairmore, and Blackstone-Paskapoo wedges (in stratigraphic order). Each was deposited during a time interval of vigorous tectonism and mountain building within the adjacent part of the orogen, and each wedge was separated from the other by a period of tectonic quiescence during which the orogen and foreland basin was partly unloaded by erosion. Deposition of the Kootenay wedge was followed by a period of erosion and uplift that affected the entire foreland basin. The period of erosion and uplift that followed deposition of the overlying Blairmore wedge affected only the southern portion of the basin, and soon gave way to deposition of a thick, marine shale succession called the Alberta Group. Erosion and uplift have been essentially continuous since the last interval of mountain building, when the Brazeau-Paskapoo wedge was deposited.

The stratigraphic and time stratigraphic relationships within the clastic wedge sequence along the entire strike length of the Canadian Rockies are displayed in Figure 13. Significant changes in facies, thickness and distribution of the clastic wedge are apparent as the sequence is followed northward (Fig. 13a, b). The nonmarine Kootenay and Blairmore wedges change facies laterally into large deltaic complexes, north of which equivalent strata consist of marine shale, sandstone, and siltstone. The unconformity at the top of the Kootenay wedge truncates laterally equivalent strata in a northward direction across Halfway River map area. North of the unconformity, the entire time stratigraphic interval is removed. Most of the Alberta Group shales and all of the Brazeau-Paskapoo wedge are absent north of the Peace River (latitude 56°N). Thus in the north, evidence for three separate tectonic pulses is lacking because the oldest Kootenay equivalent strata and youngest Brazeau-Paskapoo equivalent strata are no longer present, and the Blairmore equivalent strata that remain are, for the most part, homogeneous shales.

At the latitude of the Halfway River map area, the clastic wedge consists of deltaic to marine sandstone, shale and siltstone transitional between the Kootenay and Blairmore wedges to the south, and marine shales to the north. The oldest succession, which is time equivalent with the Kootenay wedge, is named the Minnes Group. It is composed of the Monteith Formation, a basal quartz sandstone unit; the Beattie Peaks Formation, a marine shale; the Monach Formation, a second, less extensive sandstone unit; and an upper, coal-bearing, clastic unit called the Bickford Formation. Rapid truncation beneath the sub-Blairmore unconformity leaves little of the Minnes Group preserved north of the Halfway River map boundary.

The overlying, younger portion of the clastic wedge comprises three main stratigraphic parts: a coal-bearing, conglomerate-sandstone succession named the Bullhead Group; a marine shale, sandstone and coal bearing succession called the Fort St. John Group; and a thin but regionally extensive nonmarine conglomerate-sandstone succession named the Dunvegan Formation. The Bullhead and Fort St. John groups are lateral time equivalents of the Blairmore wedge, and were deposited as part of a major transgressive succession that progressively restricted the deposition of Blairmore strata farther south as the marine sediments stepped up and over successive delta complexes. The transgression was complete by the beginning of the Late Cretaceous, when marine shale was being deposited across the entire basin.

The uppermost shale within the Fort St. John Group is roughly equivalent to the lowermost shale of the Alberta Group (Fig. 13b). It is important to note that the major facies change between the Blairmore wedge and Alberta Group shales, which is interpreted as the end of the second major pulse of tectonic activity within the orogen, cannot be detected within the Fort St. John Group north of Peace River.

Deposition of the Fort St. John Group was terminated by a widespread marine regression during the earliest part of the Late Cretaceous. Conglomerate and sandstone of the Dunvegan Formation were deposited throughout the northern part of the foreland basin, but southward, these rock types grade into shales of the Alberta Group. The Dunvegan has not been cited as evidence of a major tectonic pulse within the orogen; however, paleocurrents from the northwest, and the presence of metamorphic clasts plus detrital mica suggest a period of emergence within the northern part of the Omineca and foreland belts (Eisbacher et al., 1974).

Rocks younger than the Dunvegan Formation are not preserved within the Halfway River map area or regions farther north, and it is not known how much, if any, Brazeau-Paskapoo time equivalent strata were deposited. Arguments will be presented in a later section that suggest that Late Cretaceous-early Tertiary mountain building was less vigorous and probably lasted for a shorter time than in the south. If correct, this would reduce the potential for deposition of a thick Late Cretaceous-early Tertiary clastic wedge.

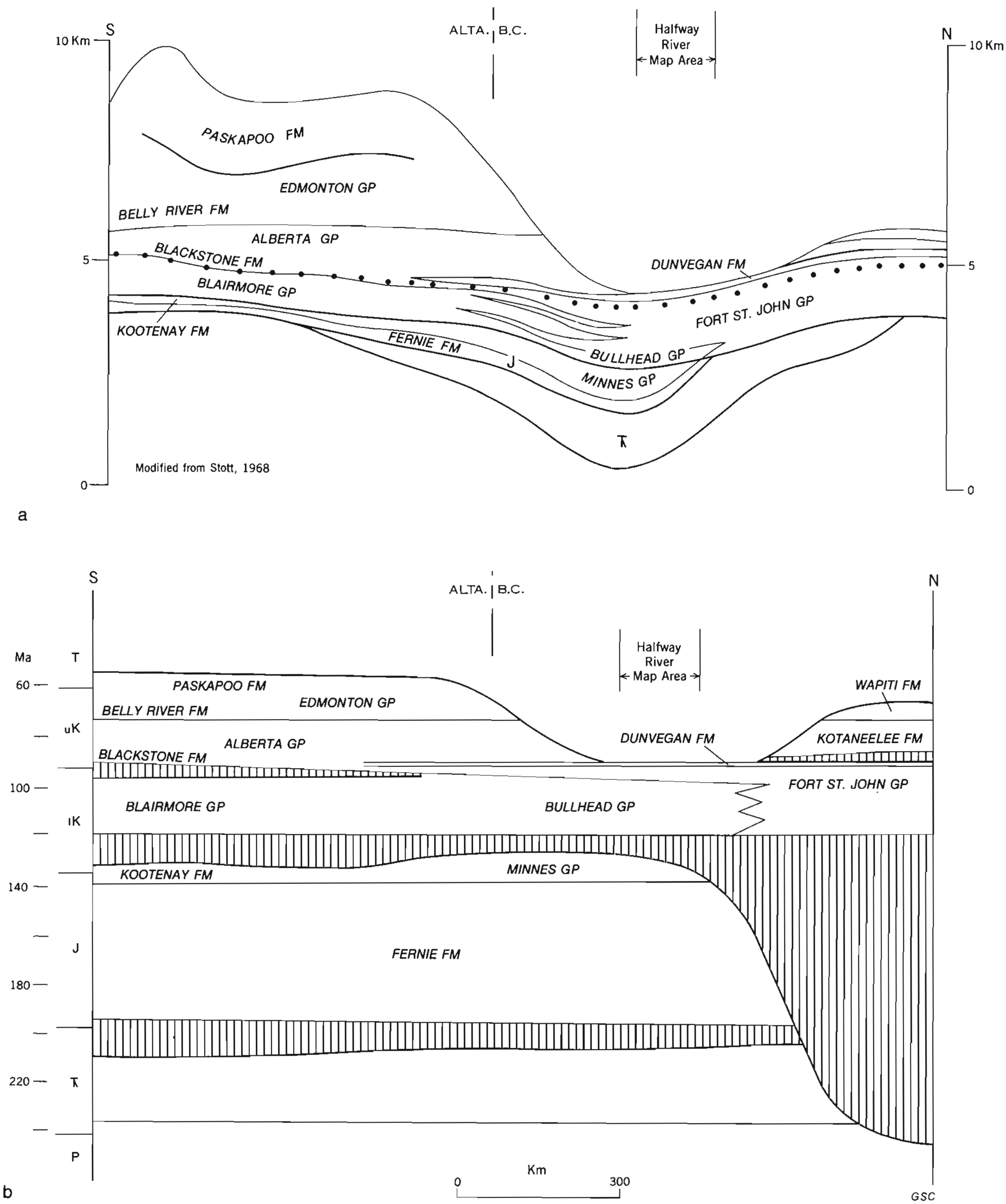


Figure 13. *Along-strike cross-sections through the clastic wedge succession of the Canadian Cordillera between lat. 49°00'N and lat. 60°00'N. (modified from Stott, 1968): (a) distribution of stratigraphic thicknesses within the clastic wedge succession; (b) time-stratigraphic distribution of units within the clastic wedge succession (vertical ruling denotes erosional period and/or nondeposition).*

STRATIGRAPHY

Introduction

Early stratigraphic studies in Halfway River map area focused, primarily, on Mesozoic strata exposed along the Peace River Canyon. Irish (1970) was the first to publish a systematic description of Halfway River stratigraphy. Since then, several parts of the stratigraphic column have received specialized attention: Mississippian through Permian strata (Bamber et al., 1968, 1980), Triassic strata (Tozer, 1967, 1982; Gibson, 1971, 1975), Jurassic and Cretaceous strata (Stott, 1967a, b, c), and portions of the Middle Devonian (Macqueen and Taylor, 1974). The stratigraphic part of this study focused on mid-Ordovician through mid-Devonian strata with the specific objective of establishing the areal distribution of shelf and off-shelf facies and of establishing how the facies correlate in time, one with the other. Consequently, the following description of Halfway River map units is very uneven. Kechika Group and older strata are given a cursory overview; the Besa River Formation and younger strata are summarized from the published descriptions of others. A summary of formations is provided in Table I, and the locations of all measured sections appear in Figure 14 and in Appendix 2.

Upper Proterozoic

Misinchinka Group
(Map units PM and PM2)

Description

The Misinchinka Group (Dawson, 1881) is the northern part of a continuously exposed belt of clastic, Upper Proterozoic rocks extending from Ware map area southeastward into the Main Ranges of the Southern Rocky Mountains. South of Pine Pass map area (93 O) the Misinchinka is called the Miette Group (Walcott, 1913). Both groups belong to the Windermere Supergroup: "... the predominately clastic succession between older Precambrian units (i.e. Hudsonian crystalline basement or Middle (?) Proterozoic carbonate-quartz successions) and the base of Lower Cambrian quartzites." (Eisbacher, 1981, p. 1).

The Misinchinka Group forms the Herchmer Thrust Sheet in southwestern Halfway River map area. Phyllitic slate, phyllite and fine grained argillaceous sandstone are the dominant rock types; they are penetratively cleaved and weather greenish grey and brown. Thin beds of black carbonaceous limestone may be present. Most Misinchinka strata exposed in Halfway River map area probably correlate with the "upper phyllitic slate subdivision" described by Gabrielse (1975) in the Fort Grahame map area. Lower subdivisions in the same region contain significant amounts of gritty feldspathic sandstone and/or conglomerate, rock types not typical of Halfway River exposures.

A resistant limestone marker unit ( ) approximately 150 m thick, extends from Williston Lake northwestward to the Ospika River. It consists of medium to thick bedded, variegated grey, fine crystalline limestone and silty limestone. This map unit persists southward into Pine Pass map area (McMechan, 1987; Taylor, 1983), and possibly northwestward into Fort Grahame map area, where a similar but laterally discontinuous carbonate unit is found (Gabrielse, 1975). It has not been established whether there are one or several carbonate markers within the Misinchinka Group. Taylor (1983) correlated the carbonate marker in Pine Pass map area with the Byng Formation (Slind and Perkins, 1966, p. 448), a 310 m thick sandy dolostone succession occurring 250 to 1000 m beneath the top of the Miette (Misinchinka) Group. However, until the internal stratigraphy of the Misinchinka Group is better known, and the relative stratigraphic position(s) of carbonate unit(s) better established, interregional correlations remain speculative.

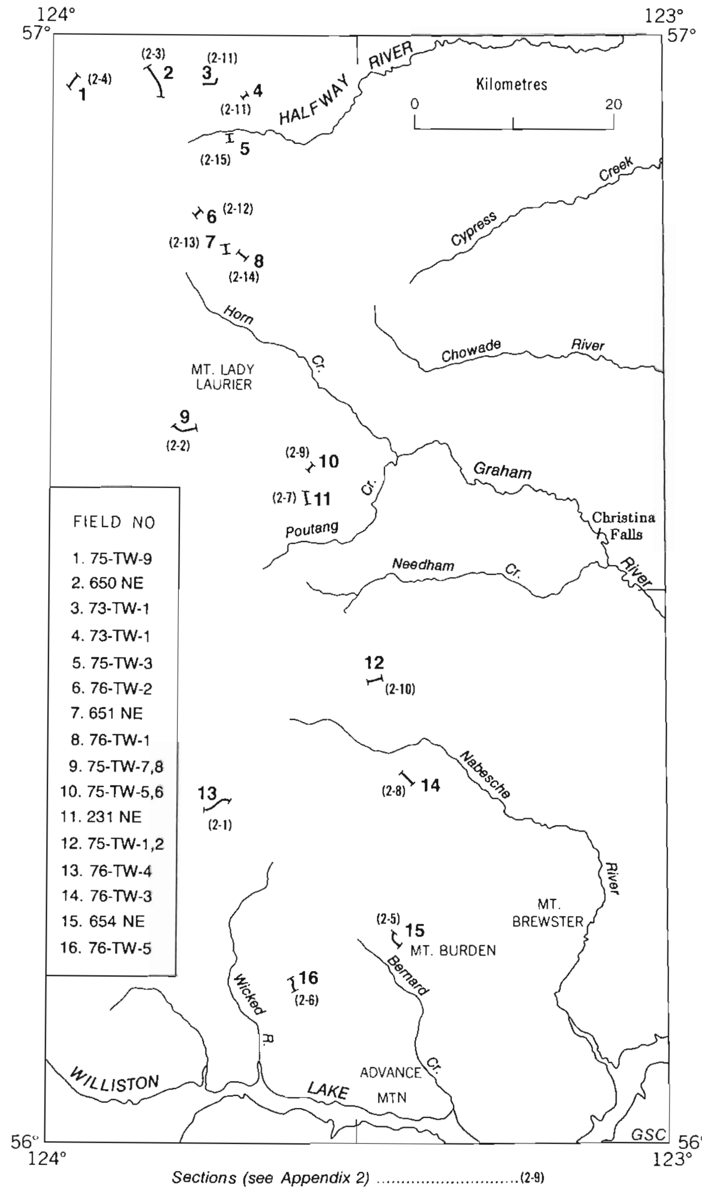


Figure 14. *Locations of all measured sections, numbered consecutively from north to south. Numbers in parentheses refer to appropriate appendix in text where stratigraphic section is described in detail.*

The large apparent thickness (> 2000 m) of Misinchinka Group phyllites in Halfway River map area is probably due to duplication across several unrecognized thrusts (McMechan, pers. comm., 1981). A more comprehensive description of the Misinchinka Group is provided by Irish (1970), Gabrielse (1975), Evenchick (1982) and McMechan (1987).

TABLE I

Table of formations, Halfway River map area

ERA	EPOCH	STAGE	GROUP	FORMATION	LITHOLOGY	THICKNESS (m)
CENOZOIC	PLEISTOCENE AND RECENT				Till; gravel; sand; silt	
MESOZOIC	UPPER CRETACEOUS	CENOMANIAN		*DUNVEGAN FORMATION*	Massive conglomerate; fine to coarse grained sandstone; carbonaceous shale	165
	LOWER CRETACEOUS	ALBIAN	*FORT ST. JOHN GROUP*	*SULLY FORMATION*	Dark grey marine shale with sideritic concretions; marker bed with fish remains	100±
			FORT ST. JOHN GROUP	*SIKANNI FORMATION*	Fine grained, crossbedded marine sandstone; silty mudstone	130±
			FORT ST. JOHN GROUP	*BUCKINGHORSE FORMATION*	Dark grey marine shale; siderite concretions; fine grained sandstone	1150±
		BARREMIAN APTIAN	*BULLHEAD GROUP*	*GETHING FORMATION*	Fine grained, cherty marine sandstone; rusty weathering shale; minor conglomerate and carbonaceous shale; coal seams	360-420
			BULLHEAD GROUP	*CADOMIN FORMATION*	Massive conglomerate and coarse grained sandstone; carbonaceous shale; minor coal	100±
				unconformity		
	UPPER JURASSIC AND LOWER CRETACEOUS	UPPER TITHONIAN VALANGINIAN	*MINNES GROUP*	*BICKFORD FM. MONACH FM. BEATTIE PEAKS FM. undivided*	Interbedded mudstone, siltstone, fine grained sandstone; thick bedded sandstone in middle; minor carbonaceous shale and minor coal	0-520
			MINNES GROUP	*MONTEITH FORMATION*	Massive quartzose sandstone; alternating units of sandstone and mudstone; minor conglomerate	200±
	JURASSIC	SINEMURIAN TITHONIAN		*FERNIE FORMATION*	Calcareous and phosphatic shale; rusty weathering shale; glauconitic siltstone; sideritic shale; interbedded sandstone, siltstone and shale	150-250
				unconformity		
	UPPER TRIASSIC	NORIAN		*PARDONET FORMATION*	Dark grey and brownish grey weathering carbonaceous and argillaceous limestone; silty limestone, calcareous and dolomitic siltstone; minor shale	30-100
		CARNIAN		*BALDONNEL FORMATION*	Light grey and brownish grey weathering carbonaceous limestone, dolostone; minor siltstone; very fine grained sandstone	70-100
				LUDINGTON FORMATION	Medium to light grey weathering, dolomitic to calcareous siltstone and silty bioclastic limestone	490-895
				CHARLIE LAKE FORMATION	Buff, yellow, grey and brown weathering dolomitic and calcareous sandstone, siltstone, sandy limestone and dolostone; minor intraformational breccia	300-440
	MIDDLE TRIASSIC			*LIARD FORMATION*	Buff, yellow and grey weathering dolomitic and calcareous sandstone and siltstone; silty and sandy dolostone and limestone	0-300
	LOWER TRIASSIC	GRIESBACHIAN LADINIAN		*TOAD FORMATION*	Dark grey, argillaceous, calcareous siltstone, silty limestone and silty shale; minor silty dolostone, calcareous sandstone	160-825
				GRAYLING FORMATION	Dark grey, dolomitic quartz siltstone and silty shale; minor calcareous siltstone, silty limestone, dolostone and very fine grained sandstone	35
				unconformity		
	UPPER AND LOWER PERMIAN	ARTINSKIAN WORDIAN		*FANTASQUE FORMATION*	Light to dark grey spicular chert, glauconitic in part; minor dark grey shale and siliceous mudstone	0-15
				unconformity		

PALEOZOIC	LOWER PERMIAN	ASSELIAN SAKMARIAN		*KINDLE FORMATION (included with Stoddart Group)*	Interbedded dark grey siltstone, calcareous sandstone, siliceous mudstone and chert; minor limestone	?
				unconformity		
	UPPER CARBONIFEROUS	LOWER NAMURIAN		*STODDART GROUP*	Grey sandstone, thick bedded; minor intercalated beds of grey-green shale and silty dolostone; dark grey-green shale at base	500+
	LOWER CARBONIFEROUS	VISÉAN		*PROPHET FORMATION*	Light grey spicular limestone, dolostone and chert; minor grey shale and siltstone	450-1830
	MIDDLE AND UPPER DEVONIAN	GIVETIAN TOURNAISIAN		*BESA RIVER FORMATION*	Dark grey shale, calcareous and siliceous; siltstone and silty dolostone; grey limestone	600 ±
		EIFELIAN GIVETIAN		*DUNEDIN FORMATION*	Dark grey, fine crystalline dolostone; stromatoporoid- and amphiporoid-rich limestone and dolostone; argillaceous limestone and dolostone; shale; debris flow breccia	50-250
				unconformity		
	LOWER AND MIDDLE DEVONIAN	PRAGIAN EIFELIAN		*STONE FORMATION*	Light grey and yellow weathering, fine crystalline dolostone, sandy dolostone and dolomitic quartz sandstone	350-600
	LOWER DEVONIAN AND UPPER SILURIAN	LUDLOVIAN? PRAGIAN?		*MUNCHO-McCONNELL FORMATION*	Light grey thick bedded to massive, fine crystalline dolostone; sandy dolostone	400-570
				unconformity		
	LOWER SILURIAN	UPPER LLANDOVERIAN		*NONDA FORMATION*	Medium grey, fine crystalline dolostone, silty-argillaceous dolostone, sandy dolostone, dolomitic sandstone, chert	340-660
	MIDDLE AND UPPER ORDOVICIAN	UPPER CARADOCIAN ASHGILLIAN		*Quartzite-dolomite unit*	Basal carbonaceous nodular limestone; middle dark grey microcrystalline dolostone; upper brown weathering dolomitic quartz sandstone and quartzite	60-90
				unconformity		
	LOWER(?) AND MIDDLE DEVONIAN	EIFELIAN	*ROAD RIVER "strata" (off-shelf facies)*	*Dolomitic quartz sandstone unit*	Light grey, dolomitic quartz sandstone, massive	0-360
	LOWER DEVONIAN AND SILURIAN	LUDLOVIAN EIFELIAN		*Brown siltstone unit*	Brown weathering calcareous and non-calcareous siltstone; silty crinoidal limestone; dark grey, argillaceous siltstone and shale	1000 ±
	LOWER SILURIAN	LLANDOVERIAN		*Breccia unit*	Medium grey, chaotic breccia; rectangular, angular, subrounded dolostone pebbles and cobbles in dolostone cement; bedded dark grey limestone with black chert nodules and lenses	50-100?
				Carbonaceous limestone unit	Dark grey argillaceous limestone; calcareous, silty shale	200 ±
				unconformity		
	MIDDLE AND UPPER ORDOVICIAN	CARADOCIAN ASHGILLIAN		*Graptolitic shale and quartzite unit*	Black shale, calcareous shale, argillaceous limestone; rusty weathering quartzite turbidites	700?
	MIDDLE ORDOVICIAN	LLANVIRNIAN CARADOCIAN		*SKOKI FORMATION*	Light and dark grey, fine crystalline dolostone; silty and sandy limestone; bioturbated beds; oncolitic beds	140-1300
	LOWER ORDOVICIAN UPPER CAMBRIAN	TREMPEALEAUAN ARENIGIAN		*KECHIKA GROUP*	Phyllitic, calcareous and noncalcareous silty shale; argillaceous limestone, calcareous quartz siltstone and sandstone	1500 ±
	MIDDLE CAMBRIAN			*Dolomite unit*	Light grey, fine to medium crystalline, thick bedded dolostone and sandy dolostone	200?
	LOWER CAMBRIAN			*Quartzite unit (Gog?)*	Quartzite, phyllitic sandstone, siltstone	650?
UPPER PROTEROZOIC				*MISINCHINKA GROUP*	Greenish grey weathering, phyllitic argillite, phyllitic slate and argillaceous sandstone; 60 m thick grey limestone and dolostone marker unit	?

GSC

Lower and Middle Cambrian

Introduction

Cambrian strata are exposed on the west side of the Ospika River between the outlets of Gavreau and Balden creeks, and in thin, folded imbricates beneath the Herchmer Thrust. A two-fold subdivision, comprising a lower, quartzite-siltstone-phyllitic argillite unit and an upper, massive, fine crystalline dolostone unit, is consistent with stratigraphic relations mapped to the west in Mesilinka map area (94 C; Gabrielse, 1975) and to the south in Pine Pass map area (93 O; Muller, 1961; Taylor, 1983; McMechan, 1987).

The lower quartzite subdivision correlates with the regionally persistent Lower Cambrian Gog Group. In the Monkman Pass map area (93 I) the Gog Group can be subdivided into three formations (Slind and Perkins, 1966): a lower quartzite (700-1700 m, McNaughton Formation); a middle carbonate (50-350 m, Mural Formation); and an upper quartzite (150-400 m, Mahto Formation). The succession thins northward to less than 700 m in the Pine Pass map area (Muller, 1961). In Halfway River and Mesilinka map areas, siltstone, shale and limestone members are more prevalent and total thicknesses vary between 400 and 1000 m.

The upper dolostone succession is unnamed, and represents a thinned, stratigraphically-condensed equivalent of the thick, Middle Cambrian carbonate succession in the southern Rocky Mountains (Aitken, 1966). Part of the succession probably contains stratigraphic equivalents of the Lynx Formation (McMechan, pers. comm., 1983). In the Pine Pass map area, the succession is approximately 500 m thick; in Halfway River and Mesilinka map areas, it is less than 400 m thick.

Quartzite unit (Gog Group)
(Map unit Єq)

Description

A succession of quartzite, phyllitic sandstone, siltstone, argillite and argillaceous limestone approximately 650 m thick overlies the Misinchinka Group. It consists of two parts: a lower, more recessive sequence dominated by fine grained quartz sandstone, siltstone, shale and argillaceous limestone; and an upper, cliff forming sequence of thin bedded to massive quartz sandstone and orthoquartzite. The two-fold subdivision is exposed west of Pattison Peak, where the lower part is very dark weathering and argillaceous. The lower contact with the Misinchinka Group was not observed. On the west slope of the Ospika River the upper contact with the overlying dolomite unit is placed beneath a mixed gradation of sandy dolostone, dolomitic quartz sandstone and quartzite. West of Pattison Peak, the upper contact is sharp, and is placed between dark, phyllitic siltstone and calcareous siltstone, and light, tan weathering dolostone.

Dolomite unit
(Map unit Єd)

Description

Light grey, fine to medium crystalline dolostone forms a distinctive marker along the footwall of the Herchmer Thrust. Occasional orange and tan weathering beds are prominent within an otherwise homogeneous, thick bedded to massive succession. Quartz sand and silt is ubiquitous. Undulating to wavy cryptalgal laminae suggest a peritidal depositional environment.

The basal contact is placed beneath a mixed gradation of quartzite and dolostone beds; the upper contact with the Kechika Group is abrupt. Thickness of the unit is estimated at 200 m within the tightly folded exposures west of Pattison Peak.

Upper Cambrian and Lower Ordovician

Kechika Group
(Map unit ЄOK)

Introduction

The Kechika Group comprises a cleaved succession of calcareous shale and argillaceous limestone that can be traced the length of the Canadian Cordilleran foreland belt. Parallel nomenclatural schemes developed independently for different regions complicate the simple stratigraphic picture. In the southern Rocky Mountains the name McKay Group is used (Evans, 1933; Norford, 1969); in McBride (Campbell, Mountjoy and Young, 1973), Monkman Pass (Taylor, 1979) and southern Pine Pass (Muller, 1961; Taylor, 1983) map areas, the term Chushina Formation is used (Mountjoy, 1962; Slind and Perkins, 1966, p. 456); and throughout the remainder of the northern Rockies, the terms Kechika Group (Gabrielse, 1963), Mount April Formation (Jackson, Steen and Sykes, 1965; Gabrielse, 1975), and Kechika Formation (Cecile and Norford, 1979) are used. Kechika Group is used in this report.

The Kechika Group was defined originally by Gabrielse (1963) as a distinctive and widely distributed succession of: "... Folded and cleaved ... thin-bedded shale, slate, calcareous phyllite, limestone and limestone conglomerate ..." (p. 32) of latest Cambrian through Middle Ordovician age, within McDame map area (104 P). Subsequently, in Ware map area (94 F), Jackson, Steen and Sykes (1965) recognized and named two stratigraphic sequences within the Kechika Group: the Mount April Formation, comprising argillaceous limestone of Lower Ordovician age, and the overlying Cloudmarker Formation, comprising noncalcareous, fine grained clastic rocks of Middle and Upper Ordovician age. It is now apparent that the Cloudmaker Formation should be included within overlying Road River strata – a succession of off-shelf graptolitic shale, siltstone and sandstone of mid-Ordovician through mid-Devonian age. Exclusion of the Cloudmaker Formation from the Kechika Group makes the Mount April Formation synonymous with the Kechika Group. In Mesilinka map area (94 C), Gabrielse (1975) followed convention and called Upper Cambrian and Lower Ordovician pelitic rocks Mount April Formation. In Trutch map area, Cecile and Norford (1979) described the same time-stratigraphic interval as the Kechika Formation; and Kechika, without a modifier, is used on the Ordovician correlation chart for Canada (Barnes et al., 1981).

Use of Kechika Group instead of Mount April Formation or Kechika Formation is suggested for two reasons:

1. Kechika appeared in the literature before Mount April, and is a more familiar and widely used term

2. Group rather than formation status permits future flexibility in defining mappable lithological subdivisions as formations.

Description

The Kechika Group (Fig. 15) comprises three main lithofacies: phyllitic, calcareous and noncalcareous silty shale; argillaceous limestone with interbedded calcareous silty shale; and calcareous quartz siltstone and sandstone. Compositional gradation between lithofacies (mixed and gradual) makes subdivision of the Kechika Group difficult. Phyllitic, silty, calcareous shale is the dominant lithotype; it is penetratively cleaved, indistinctly bedded, and thin beds of more calcareous shale containing limestone lenses and nodules are common. The argillaceous limestone comprises beds up to 3 m thick of medium grey, fine crystalline limestone separated by beds of calcareous shale. Cyclical alternation of the limestone and shale beds results in a ribbed weathering pattern.

The basal part of the Kechika Group consists of light to medium grey, platy limestone with argillaceous partings, in sharp contact with underlying Middle Cambrian dolostones. Neither breccias and arenaceous units, found along the basal contact in Trutch map area (94 G; Cecile and Norford, 1979), nor the putty-grey weathering argillaceous limestone unit (30 to 500 m thick) that sits 60 to 120 m above the basal contact in Trutch map area was observed in Halfway River map area.

Upper contact of the Kechika Group with overlying massive dolostones of the Skoki Formation is, in most places, gradational and conformable. An upward increase in the proportion of medium and thick bedded limestone is accompanied by an increasing proportion of lighter-coloured dolostone beds. Precise location of the contact is a matter of preference.

Thickness of the Kechika Group exceeds 1500 m near Mount Kenny (incomplete section). No complete section was found.

Age

Paleontological collections by Irish (1970), and Cecile and Norford (1979), suggest a latest Cambrian through Early Ordovician age. Upper age limit is within the latest Canadian stage (Appendix 1-1).

Mid-Ordovician through Devonian shelf carbonate units

Skoki Formation

(Map unit Os)

Introduction

The Skoki Formation (Walcott, 1928, p. 217, 218) is a thick, light grey weathering, dolostone succession that can be traced more or less continuously from the western Main Ranges of the southern Rocky Mountains (Norford, 1969) northward beyond the Halfway River map area (Thompson, 1976) into Trutch (94 G) and Ware (94 F) map areas (Cecile and Norford, 1979; Taylor, et al., 1979; Gabrielse, pers. comm., 1981). Regional correlation between the southern and northern Rockies is not fully established. In the northern part of the Halfway River map area, the Skoki

Figure 15. *Northward view of the Kechika Group, which forms the ridge complex on the west side of Mt. Kenny. (GSC photo no. 657-225).*

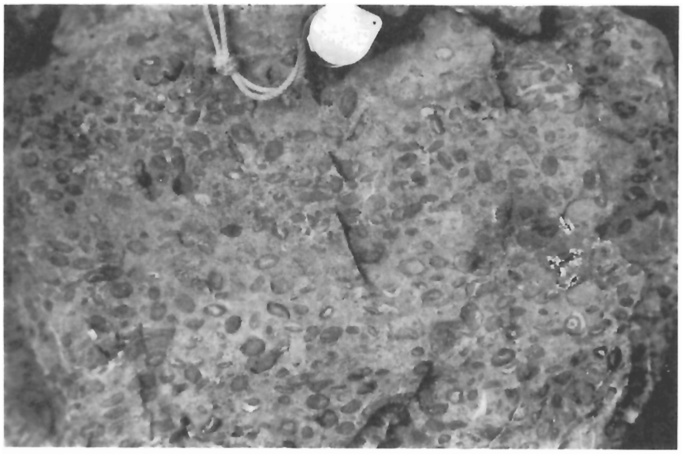
Formation is up to 1260 m thick and ranges in age within the Middle Ordovician from early Whiterock through to early Caradoc; in the south, the Skoki is only 240 m thick and is restricted in age to the early Whiterock. It may be that what is mapped as the upper part of the Skoki within the northern Rockies is correlative with a younger carbonate unit in the south called the Owen Creek Formation (B.S. Norford, pers. comm., 1981).

Within the Halfway River map area, thickness of the Skoki Formation is variable, and unlike the underlying Kechika Group, which changes little from map area to map area, the Skoki Formation has a definite western limit beyond which it changes facies to the black shales, cherts and siltstones of the lower part of the Road River (Fig. 6). To the north in Trutch map area (94 G), Cecile and Norford (1979) mapped a 20 km wide facies belt of slope breccias that are transitional between the Skoki Formation and the Road River Group. These breccias were not observed within the Halfway River map area.

Description

In the northern part of the Halfway River map area, near Robb Lake, the Skoki Formation consists of three distinctively coloured units – a middle light grey dolostone is bounded by dark grey, more massive dolostone units. This colour distinction does not persist south of Lady Laurier Lake, but the occurrence of a conspicuous, dark-green weathering volcanic unit near the middle of the formation (map unit Osv), from the headwaters of Balden Creek south to the headwaters of the Nabesche River, helps distinguish Skoki from other, younger carbonate units. South and west of Mount Burden, the volcanic member is absent. Oncolite-rich beds, and mottled-weathering bioturbated beds with well preserved burrows (Fig. 16) are other characteristics common to the Skoki Formation, but not other carbonate units.

Five kilometres north of Mount Kenny at Section 2 (Appendix 2-3), the Skoki Formation is at least 617 m thick, and comprises a basal 225 m of dark grey weathering, fine crystalline, medium to thick bedded, dark grey dolostone; a middle unit 134 m thick of light grey weathering, fine crystalline, medium to thick bedded, light grey dolostone; and an upper unit 258 m thick of dark grey weathering, siliceous, fine crystalline, medium to thick bedded, grey dolostone. Distinctive oncolite-rich units a few metres to a few tens of metres thick are interbedded with mottled weathering bioturbated beds. The oncolites suggest deposition in



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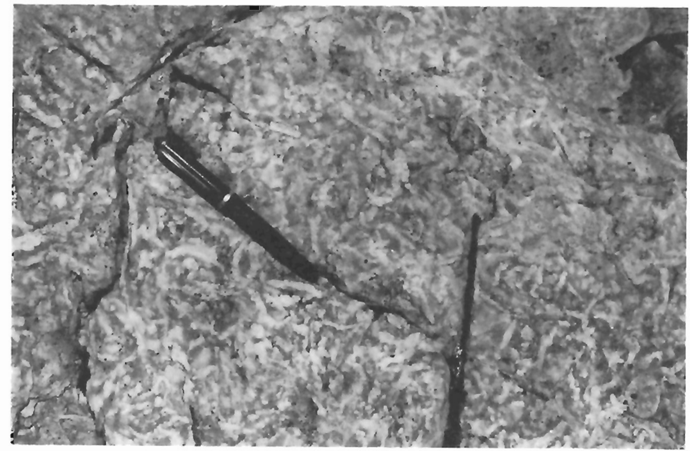

b

Figure 16 (a). *Oncolites typical of Skoki Formation.* ***(b)*** *Well preserved burrows typical of Skoki Formation. [GSC photo nos. 2212-4 (a), 943-14 (b)] .*

agitated water; the extensive burrowing suggests a more restricted, quiet water depositional environment. Patchy, light tan silicification is common in some beds. Basal contact with the underlying Kechika Group is covered; upper contact with overlying Road River strata shale and limestone is abrupt.

Seven kilometres west of Mount Kenny (Stratigraphic Section 1; Appendix 2-4), the Skoki Formation is a less resistant, limestone facies. It is 450 m thick (erosional top) and consists of thin bedded silty limestone, dolomitic siltstone, calcareous siltstone and limestone. Bedding character changes from planar, low in the succession, to undulating or wavy, high in the succession; bioturbation is common within the upper part. The basal contact is placed beneath a mixed gradation of wavy bedded, argillaceous, silvery grey limestone and medium to thick bedded, dark grey, silty limestone. The succession has a much larger fine-clastic component than sections farther east, and bedding character is consistent with a depositional environment beyond the main buildup of Skoki carbonates.

Five kilometres south of Mount Lady Laurier (Stratigraphic Section 9; Appendix 2-2) the Skoki Formation is 1262 m thick and consists of two units. The lower one is 588 m thick and is dominated by planar and wavy bedded silty limestone, thick siltstone beds, and occasional dolostone, limestone and calcareous shale interbeds. The upper unit is more massive, consisting of 647 m of dolostone, silty dolostone, dolomitic siltstone, and minor limestone and chert. Bioturbation is common, especially within the wavy bedded, silty limestones of the lower unit. Contact with the Kechika Group is conformable, and consists of a mixed gradation of silvery grey, bioturbated, thin bedded, argillaceous limestone and thin bedded, grey and tan weathering, silty limestone. The upper contact, although poorly exposed, is apparently conformable, and can be defined as a gradation from dolostone, silty dolostone and dolomitic siltstone, to the brown weathering, planar-laminated siltstone of Road River "strata". This is the thickest section of Skoki carbonates in the region and it contains both the limestone and dolomite facies described from sections farther north. Similar thicknesses and facies distributions persist southward to Wicked Lake (e.g. Section 13 located 3 km south of the headwaters of Gavreau Creek).

Two sections measured south of the Mount Burden Thrust, one at Mount Burden (654NE; Appendix 2-5) and the other 10 km to the southwest (Stratigraphic Section 16; Appendix 2-6) demonstrate a significant southward thinning of the Skoki Formation (Fig. 6). At Mount Burden it is only 144 m thick and consists of thin to thick bedded, light grey and olive grey weathering dolostone and siliceous dolostone. The basal contact is placed beneath a mixed gradation of calcareous dolostone, dolostone and dolomitic siltstone. The upper contact is abrupt and overlain by calcareous shale belonging to an unnamed Upper Ordovician quartzite-dolomite unit (map unit Oqd). Southwest of Mount Burden at Stratigraphic Section 16 (Appendix 2-6), the Skoki is 427 m thick and contains a greater diversity of rock types. The basal 105 m consists of limestone and argillaceous limestone that appears to be gradational with the underlying Kechika Group. The remainder of the section consists of medium bedded, sandy dolostone and dolostone. Oncolites and black chert nodules are abundant, and some beds contain crinoid ossicle debris. The presence of quartz sand is typical of the Skoki Formation south of the Burden Thrust and may reflect proximity to the northern boundary of the Peace River Arch.

Age

The Skoki Formation ranges in age from the latest part of the Late Ordovician (latest Canadian) into the Middle Ordovician (Whiterockian(?); see paleontological reports by Norford, Appendix 1-2).

Quartzite-dolomite unit
(Map unit Oqd)

Description

South and west of Mount Burden, a thin (less than 100 m) Middle and Upper(?) Ordovician quartzite-dolomite unit forms a slight, recessive notch between the Skoki Formation and the Lower Silurian Nonda Formation. This unit is unnamed and has not been previously described. It occupies the same stratigraphic position as the Beaverfoot Formation, a thick (>500 m) cliff forming dolostone unit that outcrops in Monkman Pass (93 I) map area and the western Main Ranges of the Rocky Mountains (Norford, 1969), but the lower carbonate part of the unit is older than the Beaverfoot. North of Mount Burden, the unnamed quartzite-dolomite unit is represented by the graptolitic shale and quartzite unit (map unit Osq) within the lower part of Road River strata (Fig. 7).

Two stratigraphic sections of the quartzite-dolomite unit were measured. At Mount Burden (Stratigraphic Section 15; Appendix 2-5) it is 86 m thick, and comprises three sub-units: a basal, carbonaceous, nodular limestone, a middle, dark grey weathering, microcrystalline dolostone with black chert nodules, and an upper, brown weathering quartzite and dolomitic quartz sandstone with minor shale interbeds. Ten kilometres to the southwest (Stratigraphic Section 16; Appendix 2-6) the unit is 62 m thick and comprises three members: a basal, thin bedded, sandy dolostone with chert nodules; a middle, black, fissile shale; and an upper, thick bedded, sandy dolostone with red weathering argillaceous partings and an abundance of brachiopod shell fragments.

The basal contact with the underlying Skoki Formation is sharp and unconformable (Norford, pers. comm., 1976); the upper contact is placed beneath a gradational change to sandy dolostone beds that form the basal part of the overlying Nonda Formation.

Age

The unnamed quartzite-dolomite unit is of Middle and possibly Late Ordovician age. According to Norford, (Appendix 1-3), the basal limestone member is late Caradocian. A disconformity may exist between it and the overlying member(s), which are middle Ashgill. A disconformity separates the upper member(s) from the overlying Nonda Formation.

Nonda Formation
(Map unit SN)

Description

The Nonda Formation (Norford et al., 1966) is a 200 to 660 m thick dolostone succession of Early Silurian (Llandoverian) age that can be traced from northern Pine Pass map area northward to the British Columbia – Yukon border. It is a distinctive, medium grey weathering, medium bedded to massive dolostone in which quartz silt and sand are ubiquitous. Chert nodules and silicified fossils are common throughout, in contrast with overlying Silurian and Devonian carbonate units in which fossils are rare. In the northern part of the outcrop belt, the basal part of the Nonda consists of a distinctive quartz sandstone unit; in the Halfway River map area, the basal beds are composed of dark grey weathering, thin bedded, carbonaceous dolostone and silty dolostone.

At its type section (Norford, et al., 1966) near Toad River bridge on the Alaska Highway, the Nonda is 320 m thick. It thins southward to 210 m at Gathto Creek, and then thickens toward Halfway River, reaching a maximum of 660 m at Guilbault Creek (Stratigraphic Section 11; Appendix 2-7). From there to the Peace River, the Nonda thins to 340 m at Advance Mountain. The changes in stratigraphic thickness along strike are controlled by unconformities at the base and top – the thicker the section, the younger the fossil assemblages that are found in its uppermost units. On Advance Mountain, relief of 30 cm was observed along the basal contact (B.S. Norford, pers. comm., 1980). This contact is abrupt, and separates the thick bedded, light grey dolostone of the unnamed Upper Ordovician quartzite-dolomite map unit from the thin to medium bedded, dark grey dolostones of the Nonda Formation.

Most of the exposed Nonda Formation within the Halfway River map area forms the core of the Bernard Anticline. At the base of the Guilbault Creek section (Stratigraphic Section 11; Appendix 2-7), the Nonda overlies more than 100 m (exposed) of dark grey, thin to medium bedded, carbonaceous limestone and dolostone. This unit was previously described as "Kechika like" (Norford et al., 1966), but subsequent study shows these rocks contain fossils typical of basal Nonda beds. Accordingly, this basal unit is interpreted as a more argillaceous facies of the Nonda (Norford, pers. comm., 1976) and is correlated (on the basis of lithological similarity and stratigraphic position) with the Silurian carbonaceous limestone unit (map unit Scl).

The Nonda Formation, is, in part, correlative with shale, siltstone and slope breccia units within Road River strata (Fig. 7). The lateral transition, exposed near Mount Kenny, is abrupt (Fig. 8).

Age

The Nonda Formation is Early Silurian (late Llandovery) (Appendix 1-4; Norford et al., 1966).

Muncho-McConnell and Stone formations

A thick succession of light grey weathering, resistant dolostone and sandy dolostone forms most of the high peaks adjacent to the mountain front, from Peace River northward to the 60th parallel. North of Trutch map area, three formations can be distinguished (Taylor and Mackenzie, 1970): Muncho-McConnell, Wokkpash, and Stone. The Muncho-McConnell and Stone formations are thick, cliff forming dolostones, separated by the thin, more recessive dolomitic quartz sandstone of the Wokkpash Formation. As this succession is traced southward toward Halfway River map area, several lithostratigraphic changes occur. First, the Wokkpash Formation becomes indistinguishable from dolomitic quartz sandstones at the base of the Stone Formation (it is not clear whether this is due to a facies change, or beveling beneath a disconformity at the base of the Stone Formation). Second, the Muncho-McConnell Formation becomes more massive and homogeneous in character; and third, the quartz sand content of the Stone Formation increases, making dolomitic quartz sandstone an important lithological constituent.

The overall character of the Muncho-McConnell and Stone formations suggests that they represent low energy, virtually completely dolomitized, high salinity lagoonal, intertidal and supratidal carbonates (Macqueen and Thompson, 1978). The quartz sand may have come from an emergent Peace River Arch, perhaps as wind blown dunes that became dispersed by intertidal currents.

The Muncho-McConnell and Stone formations change facies laterally into the brown siltstone and dolomitic quartz-sandstone map units (Fig. 9).

Muncho-McConnell Formation
(Map unit SDM)

Description

The Muncho-McConnell Formation (Fig. 17) is a massive, cliff forming unit that can be distinguished from the underlying Nonda Formation by its lighter colour and more massive character, and from the overlying Stone Formation

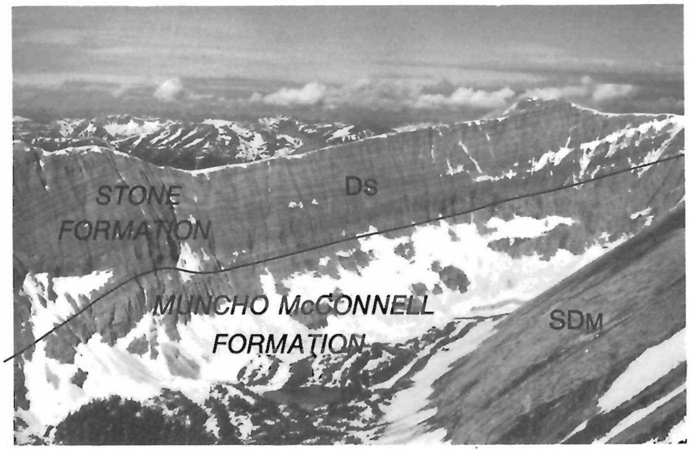

Figure 17. *Southward view of the Muncho–McConnell and Stone formations, eastern flank of Bernard Anticline, south of the headwaters of the Nabesche River. (GSC photo no. 819-37).*

by a lack of colour banding and its slightly more resistant character. Three sections (Appendices 2-8, 2-9, 2-10) were measured along the strike of the Bernard Anticline. Thickness varies from 564 m at Lapierre Creek (Stratigraphic Section 10; Appendix 2-6) to 410 m south of Needham Creek (Stratigraphic Section 12; Appendix 2-10).

The Muncho-McConnell Formation is difficult to describe because of its homogeneity. There are no distinctive lithological members that can be traced between sections, and little variation in lithology within or between individual sections. Sand and silt content does vary, generally increasing upward. Brecciation and concentrated veining is sporadic, vague stromatolite-like forms are discernable in some sections, and bedding character may be planar, wavy or nodular. A faint mottled texture indicative of bioturbation is common. Sandy beds comprise well rounded and sorted, medium to fine grained quartz sand in a fine crystalline dolomite matrix. Some of the sandstone beds are crosslaminated, with low angle forms predominating; no effort was made to determine current direction. Lower and upper contacts are abrupt. Thick bedded to massive, light-grey, fine crystalline silty and sandy dolostone overlies, with apparent conformity, darker grey, wavy to nodular and thin bedded, fossiliferous, chert-bearing, dolomitic siltstone and dolostone of the Nonda Formation. The upper contact is marked in most places by an abrupt change from thick bedded, light grey dolostone and sandy dolostone to yellowish grey weathering, medium bedded, dolomitic quartz sandstone and sandy dolostone of the basal Stone Formation.

Age

No fossils were found. A Late Silurian age for part of the Muncho-McConnell Formation is suggested on the basis of a single occurrence of the brachiopod *Kirkidium* reported by F. Manns (pers. comm., 1976) from the Robb Lake area[1] (map unit SDMS). Taylor and Mackenzie (1970) reported an Early Devonian age on the basis of fish remains found near the base; this age assignment has since been modified to Late Silurian–Pridolian (Thorsteinsson, pers. comm., 1984). Late Silurian and Early Devonian fossils were collected from the laterally equivalent brown siltstone unit.

[1]The stratigraphic level of this collection was not reported.

Stone Formation
(Map unit DS)

Description

The Stone Formation may be subdivided into two informal members; a thick; lower one consisting of striped light grey and brownish grey dolostone and sandy dolostone, and a thin, upper member consisting of massive, light grey to white weathering dolostone and sandy dolostone (Fig. 18).

The thickness of the lower member varies between 254 and 463 m along the eastern flank of the Bernard Anticline (Stratigraphic Sections 10, 5; Appendices 2-6, 2-8). Its distinctive striping is produced by gradual alternation of quartz-rich, slightly ferruginous, dolomitic sandstone with nonferruginous dolostone and sandy dolostone. It is less massive than the underlying Muncho-McConnell Formation but similarly homogeneous. Bedding is normally planar, bioturbation is rare, and sedimentary features such as crossbedding and ripples are rare.

The upper member of the Stone Formation is between 100 and 150 m thick (Stratigraphic Sections 10, 12; Appendices 2-9, 2-10) and consists of light grey dolomitic siltstone, dolomitic sandstone, dolostone and sandy dolostone. It is medium to thick, planar bedded; the dolomite is fine to medium crystalline, and quartz sand grains are well rounded and sorted. The basal contact is gradational, and the upper contact with the overlying Dunedin Formation is abrupt.

Age

No fossils were collected from the Stone Formation. An Early to Middle(?) Devonian age is suggested on the basis of stratigraphic position.

Facies changes affecting the Muncho-McConnell and Stone formations north of Lapierre Creek
(Map unit SDMS)

North of Lapierre Creek, the Muncho-McConnell and Stone formations cannot be subdivided for mapping purposes because the distinctive colour striping of the Stone Formation is lacking. Northeast of Robb Lake, the combined succession is 655 m thick (Stratigraphic Sections 3, 4; Appendix 2-10) and consists almost entirely of thick bedded to massive, fine crystalline dolostone, with dolomitic quartz-sandstone units at the base and near the top. The basal contact is marked by 2 m of breccia comprising angular pebble to boulder sized dolostone and silty dolostone fragments in a buff weathering, quartzite matrix. This breccia appears to be sedimentary in origin. The upper contact is marked by an abrupt change from highly fractured, dolomitic quartz sandstone to dark grey, thin bedded dolostone of the Dunedin Formation.

Dunedin Formation
(Map unit DD)

Introduction

Middle Devonian strata undergo rapid lateral changes in thickness and facies. At the northern extreme of the Halfway River map area, a thin (0-60 m), unconformity

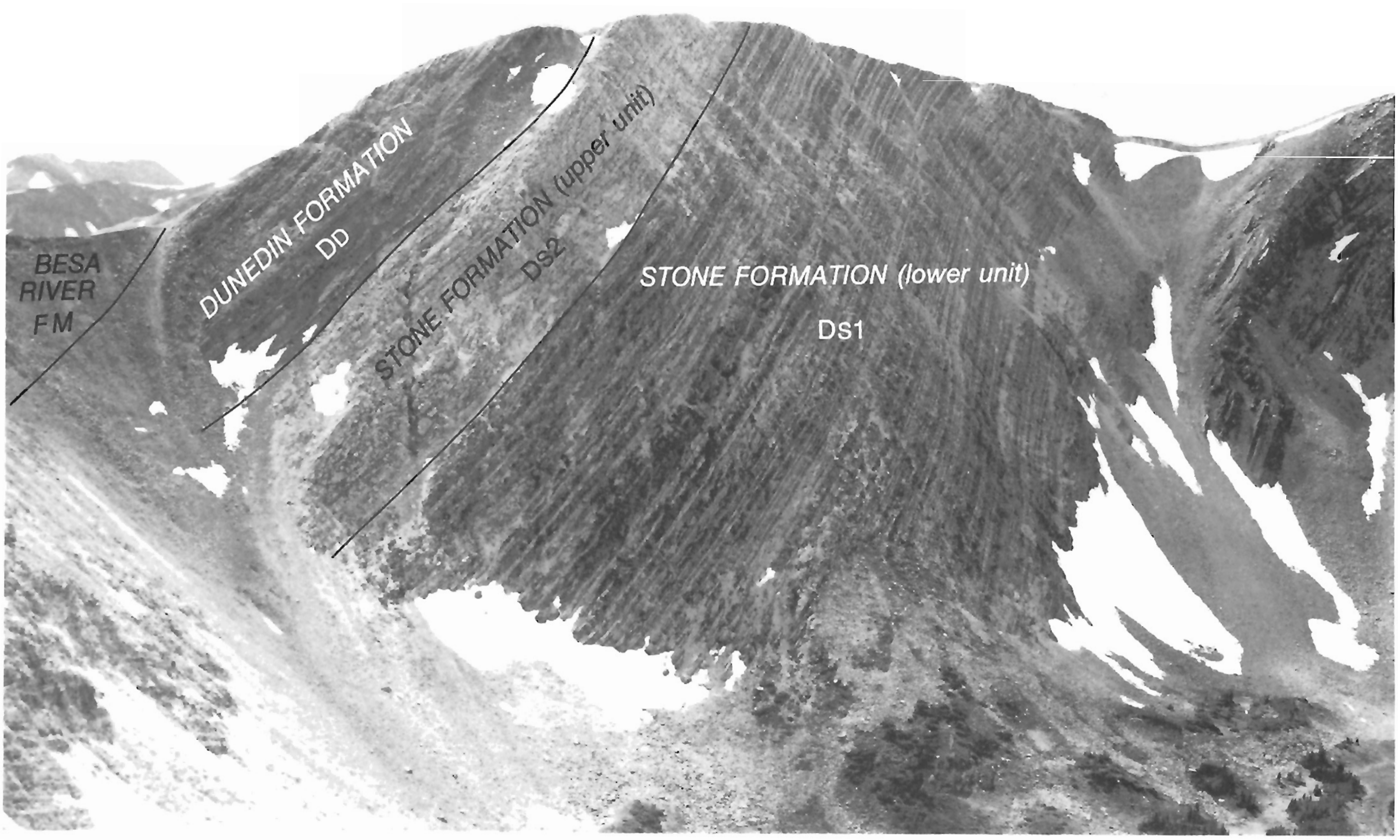

Figure 18. Southward view of the lower Stone (DS1), the upper Stone (DS2) and the Dunedin (DD) formations, eastern flank of Bernard Anticline, 5 km south of Poutang Creek. (GSC Photo no. 819-34).

bounded succession of dark grey, fine-crystalline dolostone occurs beneath the Sidenius Thrust and along the flanks of the Robb Anticline. Farther south, along the eastern flank of the Bernard Anticline, this interval contains debris flows, and beds rich with amphiporids and other stromatoporoids. Near the Bernard Fault, there is a significant increase in stratigraphic thickness to a massive stromatoporoid-rich reef complex overlain by argillaceous limestone (Macqueen and Taylor, 1974; Fig. 10). The barrier complex extends northward, in the subsurface, across Trutch and into Fort Nelson (94 J) map areas, and thence northeastward across northern British Columbia and Alberta to Great Slave Lake. It has two names: Shekilie Barrier (Hriskevich, 1970; applied mainly to the Alberta and Territories part); and Slave Point-Elk Point facies front (Griffin, 1967; applied mainly to the subsurface British Columbia part).

All Middle Devonian (post-Stone Formation) carbonate facies exposed in Halfway River map area are included within the Dunedin Formation (Taylor and Mackenzie, 1970). This departs from earlier convention (Macqueen and Taylor, 1974; Taylor and Bamber, 1970), which restricted the name Dunedin Formation to the relatively thin, dark grey dolostone succession beyond the barrier complex. Macqueen and Taylor (1974) correlated the barrier complex with the Pine Point Formation, and correlated the argillaceous carbonate units overlying the barrier complex with the Sulphur Point, Watt Mountain and Slave Point formations (Fig. 19). These correlations were tentative, based on long range extrapolation of subsurface data across the Foothills subprovince (where no data exists) to the Rocky Mountains.

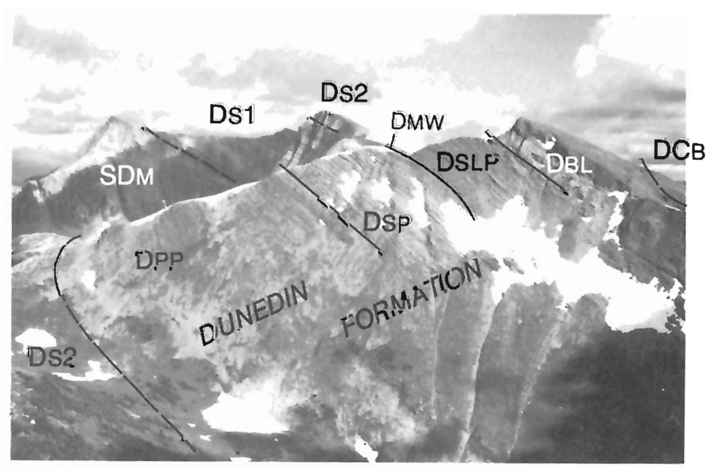

Figure 19. Southward view of the barrier facies of the Dunedin Formation south of Bernard Fault along the plunge-out of the Bernard Anticline (SDM, Muncho–McConnell Formation; DS1, lower Stone Formation; DS2, upper Stone Formation; DPP, Pine Point Formation; DSP, Sulphur Point Formation; DSLP, Slave Point Formation; DWM, Watt Mountain Formation; DBL, Beaverhill Lake Formation; DCB, Besa River Formation). DPP, DSP, DSLP, DWM, DBL are subsurface stratigraphic equivalents of Dunedin Formation (Macqueen and Taylor, 1974). (GSC photo no. 819-59).

Description

East of Mount Kenny, the Dunedin Formation forms a continuous, dark grey marker beneath the Sidenius Thrust. Above the thrust, around the margin of the Robb Anticline, the marker is less obvious and becomes discontinuous. Near the axis of the Robb Anticline (Stratigraphic Sections 3, 4; Appendix 2-11) the Dunedin consists of 61 m of medium to thick bedded, fine crystalline dolostone, wackestone and grainstone in which stromatoporoids and crinoid ossicle fragments are common. Upper and lower contacts are sharp. The lower contact is at the base of a 3 m dolomitic quartz sandstone; the upper contact occurs at the abrupt change from thin bedded, oolitic dolostone to black and brown weathering, noncalcareous shale of the overlying Besa River Formation. Relief of up to 1 m along the lower contact was reported by Taylor (1979).

The Dunedin Formation was not mapped within the Laurier Anticlinorium (between Halfway River and Lady Laurier Lake). A 20 m thick calcareous shale unit occurs at the base of the Besa River Formation. This shale unit might be a Dunedin correlative.

East of Lady Laurier Lake at the north plunging termination of the Bernard Anticline, the Dunedin Formation was not mapped separately, but it is present in the form of dark grey, stromatoporoid-bearing, medium to fine crystalline dolostone occurring on the eastern flank of the anticline and within the adjacent, narrow, anticlinal fold core.

Northeast of Lapierre Creek, the Dunedin Formation is mappable, and can be traced southward along the eastern flank of the Bernard Anticline. It is accompanied by a progressive southward change in lithofacies to wavy to nodular bedded, silty dolostone, stromatoporoid-rich limestone, thin-laminated calcareous shale, and diamict conglomerate comprising pebble and cobble sized dolostone clasts in a quartz-dolomite matrix. These lithofacies are typical of forereef slope deposits and signal proximity to the barrier complex encountered farther south. South of Needham Creek (Stratigraphic Section 8; Appendix 2-10), the Dunedin Formation is 82 m thick, but from here southward and westward, thickness and complexity of the carbonate succession increase significantly.

West and south of the Bernard Fault, where the barrier complex is best developed (Fig. 19), the Dunedin Formation comprises three main lithofacies (Macqueen and Taylor, 1974). The basal unit consists of a stromatoporoid-rich reef complex overlain by two, bedded, argillaceous carbonate units that are separated by fine and medium grained sandstone. The argillaceous carbonate units interfinger with shale tongues belonging to the laterally equivalent Besa River Formation. In order of superposition, Macqueen and Taylor (*ibid.*) correlated these lithofacies with the Pine Point (barrier complex), Sulphur Point (argillaceous fine crystalline dolostone and limestone), Watt Mountain (sandstone) and Slave Point (argillaceous carbonate) formations. The transition from carbonate to shale occurs along an east-west trend north of the Bernard Fault.

South of the Bernard Anticline, within the Burden Thrust sheet, the upper three lithofacies are absent. The lowest unit is thinner (only 52 m at Mount Burden; Macqueen and Taylor, 1974) and consists of dark grey, argillaceous limestone and calcareous shale in which debris flow breccias are common; it is overlain by Besa River shale. The apparent anomalous position of the deeper water foreslope facies in the Burden Thrust sheet, due south of the barrier facies within the Burden Anticline, can be interpreted in two ways: as a consequence of later (Mesozoic) west to east movement on the Burden Thrust sheet; or as a result of the inversion of the Peace River Arch resulting in a depression.

Age

The Dunedin Formation ranges from the Eifelian to the Givetian. At Stratigraphic Sections 3 and 4, ostracodes collected 10 m above the base (Appendix 1-5) give an Eifelian age. South of the Peace River adjacent to Mount Selwyn (GSC loc. C-56105) brachiopods collected from a basal, recessive, calcareous shale unit are Eifelian and/or Givetian. Both Eifelian and Givetian fossils were collected from Middle Devonian carbonate strata by Irish (1970, p. 35, 36). There is no way of assessing the diachroneity of the Dunedin Formation without further systematic biostratigraphic studies.

Mid-Ordovician through mid-Devonian off-shelf clastic units: Road River strata
(Map unit SDMS)

Introduction

Road River strata are composed of a thick succession of off-shelf clastic rocks that were deposited adjacent to and contemporaneously with the quartzite-dolomite unit, and the Nonda, Muncho-McConnell, and Stone formations. The off-shelf facies make up the Laurier Anticlinorium between the Halfway River and Lady Laurier Lake, and form parts of more westerly thrust sheets between Balden Creek and the outlet of Wicked River. Five separate and mappable lithofacies are recognized within the Laurier Anticlinorium: a Middle and Upper Ordovician graptolitic shale and quartzite succession (map unit Osq); a Silurian carbonaceous limestone succession (map unit Scl); a Silurian breccia unit (Nonda Formation equivalent; map unit Sb); an Upper Silurian through Lower Devonian brown siltstone succession (map unit SDbs), and a mid(?)-Devonian, dolomitic, quartz sandstone succession (map unit Dqs). Farther west and south, not all of the subdivisions can be distinguished.

The term Road River is applied in an informal sense to include all off-shelf rocks of mid-Ordovician through mid-Devonian age.

Graptolitic shale and quartzite unit
(Map unit Osq)

Description

The graptolitic shale and quartzite unit can be traced from Lady Laurier Lake northward around folds within the Laurier Anticlinorium to the Halfway River. It is approximately 655 m thick (combination of Stratigraphic Sections 6, 7; Appendices 2-12, 2-13) and comprises two lithofacies: 1. a black, calcareous shale, shale and argillaceous limestone succession overlain by 2. a rusty weathering, quartzite turbidite (Fig. 20). The lower, argillaceous succession is typically planar laminated to thin bedded; the shales are fissile and often graptolitic, and contact with the underlying Skoki Formation is abrupt (unconformable?). The quartzite succession comprises turbidites 3 to 40 m thick separated by one to several metres of black shale. The quartzites are massive, fine to very fine grained and homogeneous. Basal contacts may be fluted. Contact with the overlying carbonaceous limestone unit was not observed.

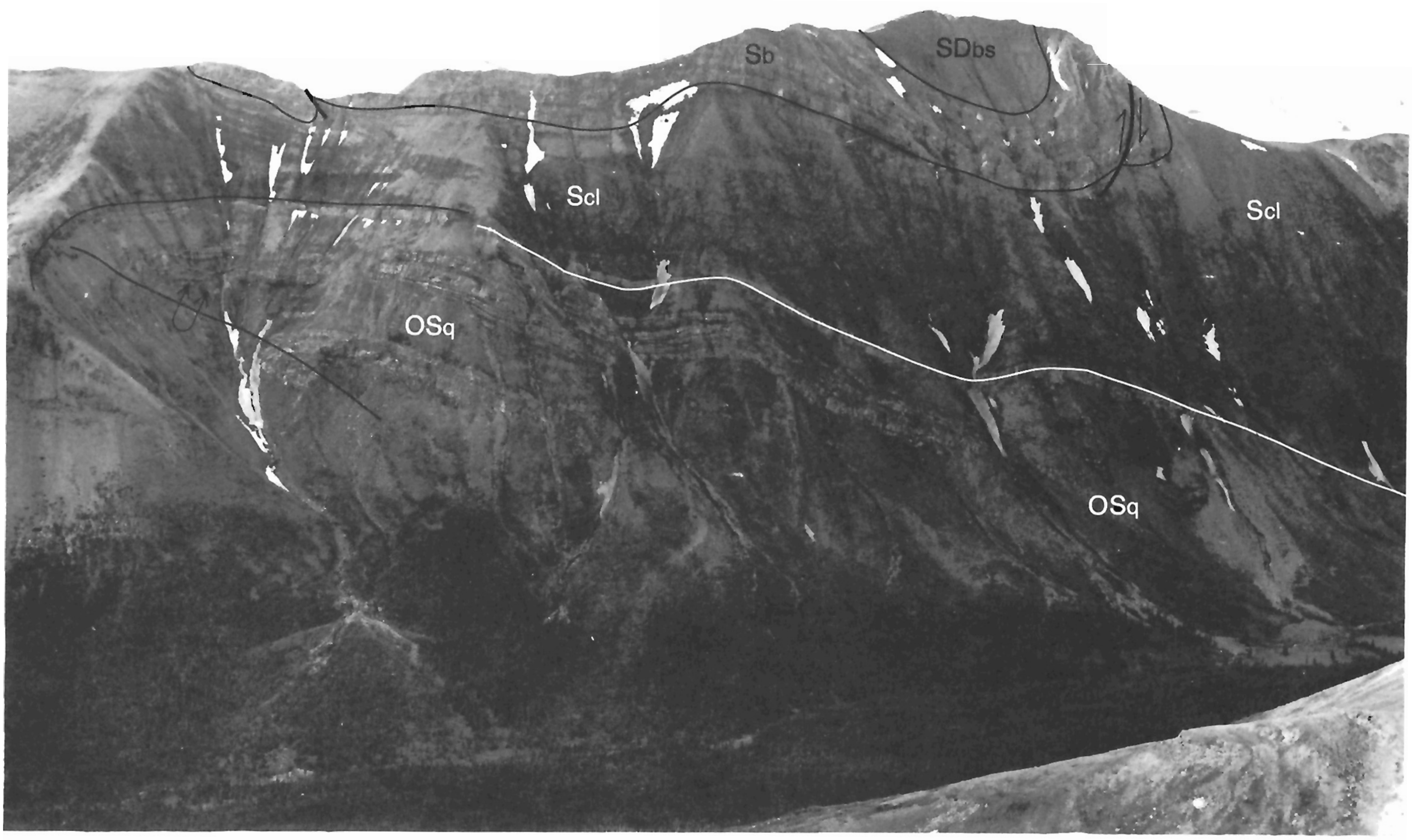

Figure 20. Southward view of the graptolitic shale and quartzite unit (OSq), the overlying carbonaceous limestone unit (Scl), the breccia unit (Sb), and the brown siltstone unit (SDbs), located on the eastern flank of Laurier Anticlinorium, 8 km south of Halfway River. (GSC photo no. 819-10).

Age

The graptolitic shale-quartzite unit ranges from late Middle Ordovician through Late Ordovician, late Caradocian through Ashgillian (see Appendix 1-7).

Carbonaceous limestone unit
(Map unit Scl)

Description

The carbonaceous limestone unit (Fig. 20) comprises dark grey to black weathering, flaggy, argillaceous limestone, silty shale and shale that is mapped from the headwaters of Aley Creek northward to the Halfway River. Its thickness is estimated at 200 m (no stratigraphic sections were measured). The argillaceous limestone is thin to medium planar bedded. Bedding of the noncalcareous silty shale is thin and lenticular, occasional beds of dolostone and quartzite intervene. Near the top are numerous calcite-filled veins, often with accessory pyrite. The uppermost beds consist of thin bedded, flaggy dolostones with pinkish brown weathering siltstone partings and laminae. Contact with the overlying breccia unit is abrupt.

In the hanging wall succession of the Sidenius Thrust (5 km north of Mount Kenny), dark weathering limestones interfinger with carbonate breccia tongues that extend down the foreslope of a Nonda reef (Fig. 8; Thompson, 1976; Norford et al., 1966). On this basis, the carbonaceous limestone unit is considered a lateral off-shelf facies equivalent of the Nonda Formation.

Age

A Lower Silurian age is postulated for two reasons: North of Mt. Kenny (Fig. 8) carbonaceous limestones underlie and intertongue with debris flow breccias derived from Nonda reefs; and at the base of the Nonda Formation at Guilbault Creek (Stratigraphic Section 11; Appendix 2-7; Appendix 1-4, collections C-64489, C-60930) are platy carbonates of Llandovery age that are lithologically similar to the carbonaceous limestone unit. A single fossil collection from the carbonaceous limestone unit north of Lady Laurier Lake (Appendix 1-8) indicates a probable Silurian age.

Breccia unit
(Map unit Sb)

Description

A thin, cliff forming, dolomite breccia unit can be mapped from Lady Laurier Lake northward to beyond Mt. Kenny. It consists of angular and tabular dolostone and sandy dolostone pebbles and cobbles in a coarse, dolomite cement, interbedded with chert-bearing, fine crystalline dolomite (Fig. 21). The breccias are monomictic; the fragments are derived from the Nonda Formation and include randomly oriented corals. This unit is interpreted as a debris flow (or succession of debris flows) deposited on the foreslope of a Nonda reef (see Figure 8).

***Figure 21.** Breccia unit (Sb) containing randomly oriented rectangular blocks of Nonda carbonate. (GSC photo no. 943-2).*

Age

A Lower Silurian Nonda-equivalent age is indicated on the basis of clast provenance and the lateral continuity of Nonda into breccia exposed northeast of Mount Kenny (Fig. 8).

Brown siltstone unit
(Map unit SDbs)

Description

Poor exposure, complex folding and rapid lateral changes in stratigraphic character make the brown siltstone unit difficult to subdivide. A minimum thickness of 457 m was measured across the upper limb of a recumbent fold (south slope of Calnan Creek, Stratigraphic Section 8; Appendix 2-14). This section is typical of the unit exposed along the mountain front between Horn Creek and the Halfway River. It is recessive, forms smooth ridges, and weathers a distinctive light rusty brown to light brownish grey. The main lithological component is laminated, thin bedded, calcareous and noncalcareous siltstone; dark grey calcareous shale, and ribs of laminated, lenticular-bedded light grey dolostone also occur. South of the headwaters of the Graham River, the unit consists of a lower, more competent, orange-brown weathering noncalcareous siltstone succession and an overlying, brownish grey weathering, calcareous siltstone and silty limestone.

On the lower, western slopes of Horn Creek, a more westerly facies of the brown siltstone unit is repeated across a small thrust. A well exposed but structurally complicated section is easily accessible along the east-west ridge immediately northeast of Lady Laurier Lake. stratigraphic thickness is probably closer to 600 m. Contact with the underlying Silurian breccia unit is not exposed, but the basal 10± m consists of fine grained, grey, blocky sandstone. This is overlain by planar laminated to thin bedded, noncalcareous, grey and brown weathering siltstone, containing occasional, thin, lenticular beds of fine grained sandstone and crinoidal grey limestone. Brecciated siltstone beds composed of "rip-up" clasts are common. The upper 500± m comprise siltstone interbedded with fine crystalline crinoidal limestone, a thin interval of interbedded dolostone and fossiliferous black chert, a fine grained sandstone-siltstone succession, and an upper, poorly exposed unit of crinoidal limestone interbedded with brown weathering, calcareous, crinoidal siltstone. Cut and fill structures occur within siltstone and sandstone units, especially within the upper portion of the section. Bedding is planar in the lower half of the section but may be wavy to lenticular with low profile crosslaminae in the upper half.

Farther north, in Trutch map area, a prominent, orange weathering, calcareous siltstone marker unit (Cecile and Norford, 1979) can be mapped separately. It is probably equivalent to most of the brown siltstone unit.

Age

The brown siltstone unit ranges from Silurian (post-Nonda) through Early Devonian (Appendix 1-9). Graptolites collected at the base are latest Llandovery to Wenlock; corals collected in the map unit span most of the Early Devonian.

Dolomitic quartz sandstone unit
(Map unit Dqs)

Description

A physically prominent and lithologically distinctive dolomitic quartz sandstone succession (Fig. 29) can be mapped from the Halfway River southward along the mountain front to the ridges directly south of the headwaters of the Graham River. It weathers light grey to white and consists of a homogeneous admixture of medium and fine grained quartz (90%) in a dolomite cement (10%). The quartz grains are typically well rounded and spherical with frosted exteriors. A maximum thickness of 360 m occurs on the south slope of the Halfway River (Stratigraphic Section 5, Appendix 2-15). Farther south, near Horn Creek, it is only 20 m thick. The massive, unsorted, homogeneous character of the unit suggests it was deposited as a succession of bank margin grain flows (Thompson, 1978).

Age

A Middle Devonian age is suggested on the basis of stratigraphic position beneath the Besa River Formation and above the brown siltstone unit. No fossils were found.

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[1]This section was measured but is not reported in detail because of local, unresolved structural complexities.

Upper Paleozoic

Besa River Formation[1]

(Map unit DCB)

Description

The Besa River Formation (Kidd, 1962, p. 97; Bamber et al., 1968) consists of black weathering, calcareous and noncalcareous shale, silty carbonaceous limestone, and brown weathering, argillaceous siltstone and mudstone. The shales are pyritic, contain discontinuous beds of siliceous nodules, and may be siliceous. Only one marker unit was recognized, a thin (approximately 10 m) limestone (map unit DCBl) which was mapped from the headwaters of the Nabesche River northward to near the Graham River. This marker unit also outcrops east of the Robb Anticline.

The Besa River Formation forms a narrow, topographically low, continuous outcrop belt along the western margin of the Foothills structural subprovince. It is intricately folded, faulted and cleaved. At its type section just north of the Muskwa River in Trutch Map area, the Besa River Formation is 672 m thick. To the northwest Bamber et al. (1968) measured 305 to 397 m. In Halfway River map area, its thickness is estimated at 600 m.

Age

The Besa River Formation ranges in age from Middle Devonian through Viséan or early Namurian.

Prophet Formation[2]

(Map unit CP)

Description

The Prophet Formation (Sutherland, 1958) consists of cherty, calcareous and dolomitic spiculite and spicule lime-packstone and minor grainstone that accumulated on the slope between the shelf and off-shelf facies. Thicknesses vary. Sutherland (1958, p. 23) measured 452 m along the Halfway River (123°30'W; 56°59'N) and Bamber et al. (1968) measured 1830 m along the Nabesche River. Near the western limit of the Foothills subprovince, thick tongues (<50 m) of Prophet carbonate extend laterally into the siltstone and shale of the Besa River Formation.

The Prophet Formation has been subdivided into three members (Sutherland, *ibid.*; Bamber et al., *ibid.*; Fig. 40). Member A consists of 165 m of orange-brown and dark grey, slightly recessive spiculite, radiolarite and lime-wackestone, shale, and minor siltstone. Member B comprises 165 to 195 m of chert (60 to 70% by volume), spicular limestone and dolostone that form irregular (often patchy) beds and are conspicuous as dark grey, resistant cliffs. Member C consists of 49 to 65 m of limestone (predominant) and chert, and is slightly more recessive than Member B.

For mapping purposes, the Prophet Formation was not subdivided, except along the Halfway River, where a more massive, high, grey unit (CP2) could be distinguished from a darker, more recessive weathering lower unit (CP1). It is presumed that CP2 corresponds to Member C and CP1 to Members A and B, despite differences in weathering character from those of published descriptions.

Age

The base of the Prophet Formation is diachronous, becoming younger westward. In the Halfway River map area, the Prophet Formation ranges in age from late Tournaisian to early Viséan.

Stoddart Group[2]

(Map unit CS)

Description

The Stoddart Group (Halbertsma, 1959) consists of the terrigenous, clastic sediments that blanketed the Prophet Formation and the Besa River Formation. Irish (1970) recognized two subdivisions within the Halfway River map area. The lower part of the Stoddart Group (Golata Formation) consists of partly calcareous shale with minor limestone, dolostone and sandstone; the upper part (Kiskatanawa and Taylor Flat formations) comprises a progradational succession of sandstone, shale, limestone and dolostone. These subdivisions were not mapped separately.

Exposure is confined to a continuous, narrow belt along the western margin of the Foothills subprovince, and in the cores of two anticlines farther east. Thickness is controlled by a sub-Permian unconformity. Near the Nabesche River, the Group is more than 500 m thick, but is thinned erosionally to 0 m in the northeastern part of the map area. Similarly, erosional thinning occurs southward into Pine Pass map area.

Age

The base of the Stoddart Group is early late Viséan (Early Carboniferous). The top is as young as the Mississippian–Pennsylvanian boundary (early Namurian).

Kindle Formation[1]

(Map unit PK ; included with Stoddart Group as map unit CPS)

Description

The Kindle Formation (Laudon and Chronic, 1949; Sutherland, 1958) comprises a succession of Permian siltstone, sandstone, shale and limestone 98 to 130 m thick that outcrops within the Foothills subprovince of Trutch, Tuchodi Lakes and Toad River map areas. It has been reported within the Halfway River map area (Bamber et al., 1968, p. 10; McGugan, 1967, Mount Greene beds), but could not be distinguished or mapped separately from the Stoddart Group.

Age

The Kindle Formation is Early Permian (Asselian–Sakmarian) in age. It rests unconformably on the Stoddart Group and Prophet Formation, and is overlain unconformably by the Permian Fantasque Formation.

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[1]Summarized from Bamber et al., 1968.
[2]Summarized from Bamber et al., 1968; Bamber et al., 1980.

Fantasque Formation[1]
(Map unit PF)

Description

The Fantasque Formation (Harker, 1961) consists of about 15 m of medium bedded, black spiculite chert with thin interbeds of siliceous, grey siltstone. It forms a prominent weathering rib and is a critical marker unit within otherwise incompetent and often poorly exposed shales and siltstones of the Stoddart Group and Triassic Grayling and Toad formations.

Age

The Fantasque Formation is Permian (Artinskian-Wordian). It rests unconformably on the Kindle Formation and underlying Carboniferous unit, and is overlain unconformably by the Triassic Grayling Formation.

Triassic[2]

Introduction

Triassic strata (Fig. 38) consist of a lower sequence of open marine shales and siltstones subdivided into the Grayling and Toad formations, and an upper sandstone and carbonate sequence subdivided into the Liard, Charlie Lake, Baldonnel and Pardonet formations. The upper sequence changes facies westward into the thick bedded limestone of the Ludington Formation. Lateral variations in cumulative thicknesses (as well as formation thicknesses) between closely spaced stratigraphic sections, suggests deposition onto a morphologically complex shelf.

For mapping purposes, some of the formations are combined: Grayling with Toad, Liard with Charlie Lake, and Baldonnel with Pardonet.

Most anticlines within the central part of the Foothills subprovince have a core of Triassic strata. In cross-section, some folds have a cusp shape, comprising a deeply eroded central axis of Grayling and Toad shales flanked by cliffs of more resistant sandstone and carbonate of Liard and younger formations.

Grayling and Toad formations
(Map unit TRTG)

Description

The Grayling Formation (Kindle, 1944, p. 6) consists of approximately 35 m of shaly to flaggy weathering dolomitic siltstone, silty shale, and minor calcareous siltstone, silty limestone, dolostone and very fine grained sandstone. It is recessive weathering, poorly exposed, and cannot be mapped separately from the Toad Formation within Halfway River map area.

The overlying Toad Formation (McLearn, 1946) consists of dark grey, shaly to flaggy weathering, calcareous siltstone, silty limestone, silty shale and minor silty dolostone and calcareous sandstone. The more resistant, carbonate rich beds protrude as prominent ribs. Within Halfway River map area, the Toad Formation thins westward from more than 825 m (Section 6, Gibson, 1971) in the central Foothills, to less than 160 m (Section 14, Gibson, 1975) along the eastern margin of the Rocky Mountains subprovince.

The contact of the Grayling Formation with the underlying Fantasque Formation is sharp. The contact of the Toad Formation with the overlying Ludington and Liard formations is gradational to abrupt, in places marked by a conglomerate up to 2 m thick.

Age

The Grayling Formation ranges in Early Triassic age from Griesbachian to Nammalian. The Toad Formation ranges from Early Triassic (Nammalian) to Middle Triassic (Ladinian).

Liard Formation
(Map unit TRL)

Description

The Liard Formation (Kindle, 1946) comprises dolomitic to calcareous sandstone and siltstone, with minor dolostone. Its resistant character and light buff colour contrast with the underlying Toad Formation.

The basal contact is sharp. The upper contact is gradational and difficult to place; for mapping purposes it is located at the first recessive break within the Liard – Charlie Lake succession. This does not always coincide with the stratigraphic criterion defined by Gibson (1975), which is a change from dolomitic sandstone and siltstone to sandy and silty dolostone.

Thickness of the Liard Formation varies from zero in the northwestern part of the Foothills subprovince to about 300 m along the Peace River. It is not clear how much of this northwestward thinning is due to lateral facies variations, and how much was caused by erosion beneath a pre-Ludington Formation unconformity.

Age

The Liard Formation ranges in age from the Middle to Late Triassic (Ladinian into the Carnian).

Charlie Lake Formation
(Map unit TRC)

Description

The Charlie Lake Formation (Clark, 1957; Hunt and Ratcliffe, 1959) consists of dolomitic and calcareous siltstone and sandstone, siltstone, sandy limestone and dolostone that weathers buff to light yellowish brown; anhydrite is common in eastern exposures. It is a shallow, subtidal to supratidal succession that was deposited behind (east of) a broad, carbonate shelf, the deposits of which are called the Ludington Formation. The Charlie Lake Formation is

[1]Summarized from Bamber et al., 1968.
[2]Summarized from Gibson, 1971, 1975; Pelletier, 1964.

thickest – up to 440 m – along a northwest axis that parallels the Graham River. To the north and south, thicknesses decrease gradually to about 300 m at the northern and southern borders of the map area. Lower and upper contacts are gradational. The upper contact was mapped on the basis of relief; the overlying Baldonnel Formation limestone and dolostone is more resistant.

Age

The Charlie Lake Formation is Late Triassic, probably restricted to the Carnian. This assignment is based on stratigraphic position. Fossil collections are few and not diagnostic.

Ludington Formation
(Map unit T̄Lu)

Description

The Ludington Formation (Gibson, 1971, p. 17) consists of light grey weathering dolomitic to calcareous siltstone, sandstone and bioclastic limestone that forms a westward extension of the Liard, Charlie Lake and Baldonnel formations. White weathering pelecypod bioherms are obvious (Fig. 22) within an otherwise thick bedded to massive, homogeneous, fine grained succession. Thickness varies between 490 and 985 m (approximately) with isopach trends parallel to those for the Charlie Lake Formation. Lateral facies transition with the Charlie Lake Formation is gradational. Even in areas of continuous outcrop, placement of a map boundary is, at best, a qualitative judgement. The criteria used were: homogeneity – the Ludington is compositionally more homogeneous and finer grained than the Charlie Lake; and, composition – the Ludington Formation contains more carbonaceous, silty limestone, having a distinctive pinkish weathering hue, and also contains bioherms (not present in the Charlie Lake).

Contact with the underlying Toad Formation is abrupt, marked by up to 3 m of diamictite comprising tabular siltstone clasts in a shale matrix. The diamictite is interpreted as being the result of a debris flow on the foreslope of the Ludington depositional edge. Upper contact with the overlying Baldonnel Formation is generally abrupt; when no Baldonnel is present, the contact is gradational into brown weathering Pardonet limestone.

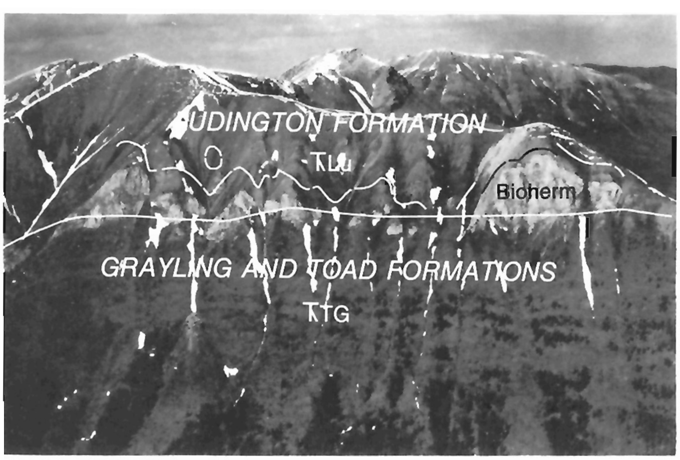

Figure 22. *Eastward view of Ludington Formation (T̄Lu) (Mt. Laurier area) showing light grey coquina bioherms overlying siltstones of the Grayling and Toad formations (T̄TG) (GSC photo no. 2212-5).*

Age

The Ludington Formation ranges in age from Middle to Late Triassic (Ladinian to late Carnian).

Baldonnel Formation
(Map unit T̄B)

Description

The Baldonnel Formation (Hunt and Ratcliffe, 1959) is a cliff forming, massively bedded, light grey weathering limestone and dolostone succession between 70 and 100 m thick. Constituent rock types are: carbonaceous wackestone, crinoid-pelecypod-brachiopod grainstone and packstone, medium to coarse crystalline dolostone, and lenses and nodules of dark grey chert. Vuggy porosity and a strong petroliferous odour when fractured are intrinsic properties. Baldonnel carbonates may be interpreted as representing a southwest to northeast, transgressive, peritidal succession.

Contact with the overlying Pardonet Formation is abrupt in most places; 3 m thick mixed gradations of coarse crystalline packstone, and medium bedded, carbonaceous wackestone immediately overlie the contact in some areas.

Age

The Baldonnel Formation is Late Triassic (Carnian).

Pardonet Formation
(Map unit T̄P)

Description

The Pardonet Formation (McLearn, 1946, 1960) comprises thin to medium bedded carbonaceous and argillaceous limestone, calcareous siltstone and minor shale that weather dark grey and brownish grey. The Pardonet is generally fossiliferous and contains wavy beds and laminae of pelecypod packstone. It is more recessive weathering, darker coloured and more evenly bedded than the Baldonnel Formation. Thicknesses increase westward from about 30 m along the eastern margin of the Foothills to 100 m at the western margin.

Contact with the overlying dark shale and siltstone of the Fernie Formation is abrupt and may be marked by a thin (3 cm) phosphatic pebble conglomerate. Near Black Bear Ridge, the contact is gradational, with no stratigraphic evidence for a break in sedimentation. However, absence of upper Norian faunal zones demonstrates the presence of a disconformity (Tozer, 1982).

Age

The Pardonet Formation is Late Triassic, Norian (see Tozer, 1967, 1982).

Jurassic[1]

Fernie Formation
(Map unit JF)

Description

The Fernie Formation (Leach, 1903, 1912; Stott, 1967b, c) forms a convenient 150 to 250 m recessive marker separating massive Triassic carbonate strata from the cliff forming sandstones of the Lower Cretaceous Monteith Formation. The Fernie Formation can be subdivided lithologically into six members:

1. A basal calcareous siltstone, limestone and shale unit similar to the Nordegg Member (Spivak 1949; 1954, p. 222) of southern Alberta

2. Black, calcareous fissile (papery) shale

3. Rusty weathering marine shale with sandstone interbeds

4. Glauconitic siltstone

5. Sideritic shale

6. Thinly interbedded sandstone and mudstone.

These members were not mapped separately.

Contact with the overlying Monteith Formation is gradational and is placed at the base of the first thick sandstone unit.

Age

The Fernie Formation ranges in age from Early Jurassic (probably Sinemurian), to Late Jurassic (Oxfordian and possibly into the Kimmeridgian).

Upper Jurassic and Lower Cretaceous clastic wedge succession[2]

Minnes Group
(Map unit JKMn)

Description

The Minnes Group (Ziegler and Pocock, 1960; Stott, 1967b) comprises up to 840 m of marine and nonmarine deltaic sandstone, siltstone and shale. In the Halfway River map area, two subdivisions were mapped by Stott (1969). The lower one is a resistant quartzitic sandstone about 200 m thick called the Monteith Formation (map unit JKM; Mathews, 1947). The upper subdivision (map unit KBp) consists of a recessive, shale-siltstone-sandstone succession called the Beattie Peaks Formation; an overlying, more resistant, well sorted sandstone succession that may correlate with the Monach Formation (Mathews, *ibid.*); and an upper coal-bearing, recessive unit comprising interbedded sandstone and shale, called the Bickford Formation (Stott, 1981). Thickness of the Beattie Peaks, Monach and Bickford formations is 520 m near the Peace River. From the Peace northward to the Halfway River it is beveled to zero metres beneath a pre-Bullhead Group unconformity.

Stott (1967b) interpreted the Minnes Group as a deltaic sequence comprising transgressive-regressive cycles deposited during the pulsating advance of an inland seaway from the north. Paleocurrent evidence suggests northward transport along the axis of the inland sea (Eisbacher et al., 1974).

Detailed stratigraphic descriptions of the Minnes Group, and the interpretation of facies relationships and paleogeography are contained in Stott (1967b; 1967c; 1969; 1973).

Age

The Minnes Group ranges in age from Late Jurassic (Tithonian) to Early Cretaceous (mid to late Valanginian).

Bullhead Group[1]
(Map unit KCG)

Description

The Bullhead Group consists of conglomerate overlain by sandstone. The conglomerate is called the Cadomin Formation (Mackay, 1929, p. 9B; 1930, p. 1310), and the sandstone is called the Gething Formation (McLearn, 1923, p. 4B). In Halfway River map area, Stott (1969) mapped the Bullhead Group without division except around the Peace River Canyon (i.e. near the present location of the Bennett Dam) and immediately to the north within the drainage basin of Dunlevy River. Beyond here, the Cadomin does not persist as a mappable unit and is probably not developed in most areas.

The Cadomin Formation (map unit KCd) consists of resistant weathering, thick to massively bedded, conglomeratic sandstone about 100 m thick. Most of the clasts are chert and quartzite pebbles; the matrix consists of medium to coarse grained, well indurated sand. Lower contact with the Minnes Group appears to be conformable. Contact with the overlying Gething Formation is gradational.

The Gething Formation (map unit KG) consists of marine and nonmarine, coal-bearing, interbedded mudstone and sandstone along the Peace River; northward it becomes a thinly interbedded succession of sandstone and shale (nonmarine?) with minor coal interbeds. Between Peace and Halfway rivers, the Gething Formation varies in thickness between 360 and 420 m. Where exposed directly on the Minnes Group (north of the last Cadomin exposures), the contact is gradational. The upper contact with the overlying Fort St. John Group is marked by a 1 to 6 m thick, glauconitic, silty sandstone (Bluesky Formation).

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[1]Summarized from Stott, 1967.
[2]Summarized from Stott, 1967c, 1973.

Age

The Bullhead Group ranges in the Early Cretaceous from mid-Barremian through to the earliest part of the Albian.

Fort St. John Group
(Map unit KFJ)

Description

The Fort St. John Group consists of marine shale and sandstone that underlies most of the eastern third of the Halfway River map area. It is composed of three parts: a lower shale and siltstone succession called the Buckinghorse Formation (map unit KBh); a middle sandstone succession called the Sikanni Formation (map unit KSk); and an upper shale succession called the Sully Formation (map unit KSu).

The Buckinghorse Formation (Hage, 1944) is about 1150 m thick and consists almost entirely of rusty weathering marine shale containing sideritic concretions. Contact with the overlying Sikanni Formation is gradational.

The Sikanni Formation (Hage 1944; redefined by Stott, 1960) is about 130 m thick and consists of three sandstone units separated by thin, silty, shale intervals. The sandstones are fine grained and finely laminated; the shales are silty, black weathering, and contain sideritic concretions. Contact with the overlying Sully Formation is abrupt.

The Sully Formation consists of about 100 m of silty, dark grey shale with occasional sideritic concretions. A prominent sandstone member that contains abundant fish scale remains correlates with the widely distributed "Fish Scales" marker of the adjacent subsurface.

The Fort St. John Group was deposited during a sequence of marine transgressions associated with southward expansion of an inland seaway extending from the Arctic Ocean.

Age

The Fort St. John Group ranges in age from the Early Cretaceous (earliest Albian) into the Late Cretaceous (earliest Cenomanian).

Dunvegan Formation
(Map unit KD)

Description

The Dunvegan Formation (Stott, 1982; Dawson, 1881) consists of sandstone, shale, and conglomerate and is about 164 m thick. The conglomerate is composed of chert and quartzite pebbles in a medium to coarse grained sand matrix; the sandstone is medium to coarse grained, composed mainly of chert; and the shale is silty and carbonaceous.

The Dunvegan Formation thickens northward. Sediment transport directions are from the north (Eisbacher et al., 1974). This succession represents a major advance of a deltaic, nearshore depositional environment across the deeper water marine environment extant during Fort St. John deposition.

Age

The Dunvegan Formation is thought to be Late Cretaceous (Cenomanian) in age.

STRUCTURE

Introduction

This section describes the folds and faults that effected lateral telescoping of the shelf and off-shelf assemblages within the Halfway River area. A model is presented illustrating how thrusts that cut mid-Paleozoic and older carbonate strata are mechanically linked to folds of late Paleozoic and Mesozoic strata.

Details of structural style vary along the Canadian Rocky Mountains. This variation is a response to differences in character of the shelf and off-shelf assemblages, to the duration and severity of deformation, and to the presence and type of pre-existing basement features that influenced later structural patterns. Moving north from the southern Canadian Rocky Mountains toward Halfway River map area, the belt narrows, the amount of shortening decreases, there is a greater diversity of structural styles, the effects of Late Cretaceous deformation are less pronounced, and the influence of pre-existing basement features on regional structural trends is more obvious.

The southern Canadian Rocky Mountains consist of four linear belts or subprovinces, each characterized by distinctive topography, stratigraphy and structure, and each separated by major thrusts or thrust systems (North and Henderson, 1954). The Foothills subprovince is topographically low, consisting of closely spaced thrusts within Cretaceous shales and sandstones. Structural complexity and relief increase gradually westward to a thrust boundary, where a thick sequence of Middle Cambrian carbonate rocks rests on folds of Cretaceous shale (Figs. 23, 24). The Front Ranges subprovince is composed of closely spaced, linear, thrust-bound mountain ranges of lower and middle Paleozoic carbonate rocks separated by valleys of recessive-weathering upper Paleozoic and lower Mesozoic clastic rocks (Fig. 25). The Main Ranges comprise broadly folded, shallow dipping thrust sheets of Lower Cambrian and older quartzite, grit and phyllite eroded into high castellated peaks. The Western Ranges consist of tightly folded and thrust faulted lower Paleozoic clastic rocks.

Only two structural subdivisions are present north of the Peace River: the Foothills structural subprovince and the Rocky Mountains structural subprovince. Each contains high mountains, there is considerable overlap in the ages of exposed rocks, and they are not separated by a large thrust. The Foothills is a fold subprovince of mainly Mesozoic age rocks comprising north trending anticlinoria separated by broad, flat synclines (Fig. 26). The eastern margin is abrupt, defined by the steep, east dipping limbs of large-amplitude en echelon anticlines (Fig. 27); 500 to 800 m of relief is average. The Rocky Mountains is a fold and thrust subprovince composed of mainly mid-Paleozoic and older rocks. It comprises an en echelon series of structural culminations; structural styles change rapidly along and across strike in response to changes in regional facies. Consequently, the northern Rocky Mountains subprovince lacks the internal organization and linear continuity of mountain ranges seen farther south; 1000 m of relief is average (Fig. 59a). The boundary between the two subprovinces is normally stratigraphic, defined by the steep, east dipping limb of a large anticline composed of lower and middle Paleozoic carbonate rocks (Figs. 27, 50b).

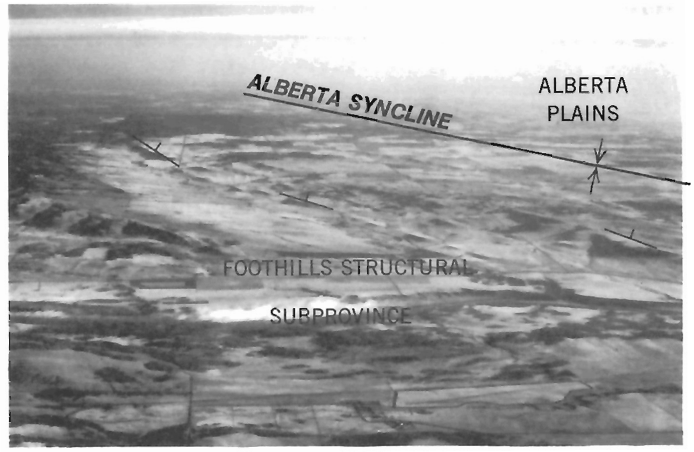


Figure 23. *Northward view along the western limb of the Alberta Syncline (lat. 51°15'N), which forms the eastern margin of the southern Foothills structural subprovince. Deformation is slight, increasing in complexity and physiographic expression toward the west. (GSC photo no. 2212-6).*

Figure 24. *Northward view along the mountain front at Mount Yamnuska (Mt. Y); the McConnell Thrust (Mc T) forms a horizontal line at a break in slope (Trans-Canada Highway, about lat. 51°15'N). Middle Cambrian Eldon (ЄE) Formation carbonate strata are thrust over recessive weathering siltstone and shale of the Upper Cretaceous Brazeau Formation. The abrupt change in physiographic expression between the Foothills subprovince on the east and the Front Ranges structural subprovince (FRSS) on the west is typical of the Southern Rockies. (GSC photo no. 2112-7).*

The mechanics of deformation within the northern and southern Rocky Mountains is the same despite differences in structural style between them. It will be argued that structural culminations and anticlinoria in the north are detached on flat décollement zones, one within the Besa River Formation (Upper Devonian through Lower Mississippian) and the other within the Kechika Group (Upper Cambrian and Lower Ordovician). Lack of a large, mappable

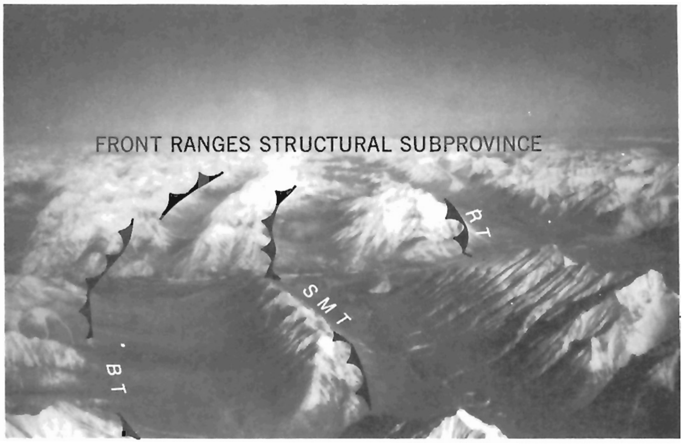


Figure 25. *Northward view along the strike of the Southern Rocky Mountain Front Ranges subprovince (town of Banff (B), Alberta, in centre middleground) showing the well organized rectilinear arrangement of thrust-bound mountain ranges comprising Paleozoic carbonate strata separated by recessive weathering clastic strata of upper Paleozoic and Triassic age (RT = Rundle Thrust; SMT = Sulphur Mountain Thrust; BT = Bourgeau Thrust). (GSC photo no. 2112-8).*

thrust along the eastern edge of the Rocky Mountains subprovince does not necessarily imply that the mountain front anticline(s) are not detached at depth, only that displacement on the detachment(s) is absorbed within higher stratigraphic levels.

Analysis of restored cross-section interpretations suggests that the significant divergence of regional structural trends north of the Peace River is an inherited feature produced by transport of miogeoclinal rocks up and across pre-existing basement structures.

The amount of shortening across the northern Rocky Mountains is considerably less than the shortening across the southern Rocky Mountains – 50 km versus 200 km. Strain affected only part of the miogeocline and none of the platform assemblage in the north but passed through the entire miogeocline and more than 50 km farther east into the platform assemblage in the south. This northward decrease in lateral shortening is also expressed within the clastic wedge assemblage. A thick, Upper Cretaceous through lower Tertiary succession occurs in the south but is essentially absent in the north, probably because there was never adequate relief in, or extent to, the deforming regions that were the source of synorogenic sediment.

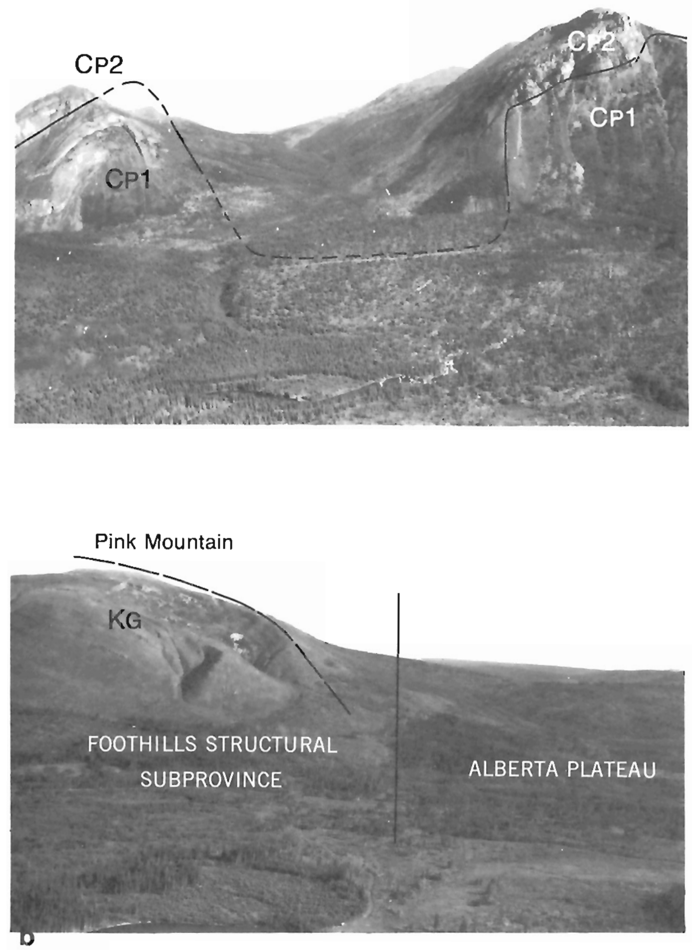


Figure 26. *(a) Northward view across the Halfway River showing anticlines of Mississippian Prophet Formation carbonate strata forming high linear ridges (500-800 m local relief) typical of Foothills terrain within the Foothills structural subprovince (compare with Figs. 23, 24). (b) Northward view across the Halfway River at Pink Mountain (lat. 57°00'N) showing the abrupt physiographic change from the Alberta Plateau on the east to the Foothills structural subprovince on the west. The change is defined by the steep dipping limb of an anticline composed of Mesozoic strata (Gething Formation [KG]; compare with Fig. 23). (GSC photo nos. 657-2 (a), 657-60 (b)).*

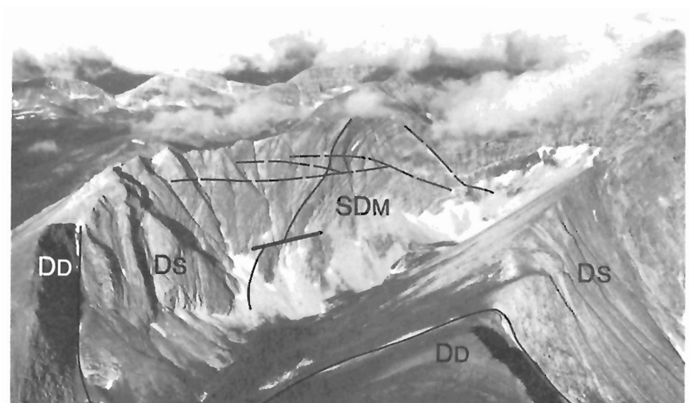


Figure 27. *Southward view along the eastern flank of the Bernard Anticline illustrating typical mountain front geometrical relationships: the eastern limb of the anticline dips steeply beneath the western Foothills subprovince; contacts are stratigraphic, there is no large mountain front thrust (compare with Fig. 24). Dashed lines parallel kinks within Muncho–McConnell (SDM) and Stone (DS) formations; subhorizontal extension fault offsets contact (DD , Dunedin Formation). (GSC photo no. 819-161).*

Physical character of the rocks

Structural styles within the Halfway River map area reflect the contrasting physical character of three, gross lithostratigraphic sequences (Fig. 28). In the east, the Foothills subprovince is characterized by a dominantly clastic, multilayered sequence of Upper Devonian through Lower Cretaceous shale, siltstone, sandstone, limestone and dolostone. Adjacent to the Foothills, and forming a narrow spine of rugged peaks along the eastern margin of the Rocky Mountains subprovince, is a second lithostratigraphic sequence comprising massive dolostone and limestone of Middle Ordovician through Middle Devonian age (shelf carbonate facies). A third assemblage dominated by shale, siltstone and argillaceous limestone of Late Proterozoic through Early Ordovician age characterizes the western part of the Rocky Mountains subprovince off-shelf facies.

Layering is the dominant physical inhomogeneity within the rock mass, and it is well developed at both mesoscopic and macroscopic scales. Within the finer grained clastic successions, layering occurs at the millimetre or centimetre scale, and reflects slight changes in grain size and/or grain composition. Abrupt compositional breaks are common at the scale of metres (up to tens of metres), with the more competent beds or rock units forming resistant ribs that outline details of the structural geometry. Structural disharmony is commonly evident between competent and incompetent units (Fig. 29) and suggests that significant strain has occurred along the planes that separate them. The more massive carbonate sequences, which dominate the eastern part of the Rocky Mountains subprovince, appear well layered when observed at a distance. However, close inspection reveals that most layering is gradational and consists of colour variations associated with slight changes in clay content and/or quartz sand content of the limestone and dolostone. The carbonate sequence is far more homogeneous compositionally than the clastic sequences, and fewer bedding surfaces were mechanically active during deformation.

Minute details of the depositional and diagenetic history of individual rock layers are preserved within the carbonate and competent clastic units, demonstrating that they have not undergone penetrative strain (flow) or significant, thermally induced mineralogical reconstitution. Strain was localized, for the most part, along discrete surfaces such as bedding planes and/or other, more widely spaced nonpenetrative surfaces. Bedding plane slippage throughout the succession was the dominant form of strain within the folded Foothills and western Rocky Mountains clastic sequences, whereas strain was concentrated along relatively few planar fault and fracture surfaces within the thrust-faulted carbonate rocks of the eastern Rockies sequence. This generalization does not apply to the dominantly shale successions, such as the Devonian and Carboniferous Besa River Formation, or parts of the Ordovician Kechika Group, in which cleavage is commonly well developed and layering is transposed (Fig. 30).

Although layering acted as an important mechanical strain absorber within the rock sequences, other parameters, such as thickness and relative competency of adjacent layers, the proportion of different rock types, and degree of anisotropy within the layered sequences, determined the geometry and style of the structures that evolved. The importance of layer parallel shear is a direct consequence of the parallelism of layering with the major compressive stress direction. As structures evolve, layering is sometimes oblique to the major stress axes, however, forcing other planes of weakness to become the sites of displacement (Price, 1965).

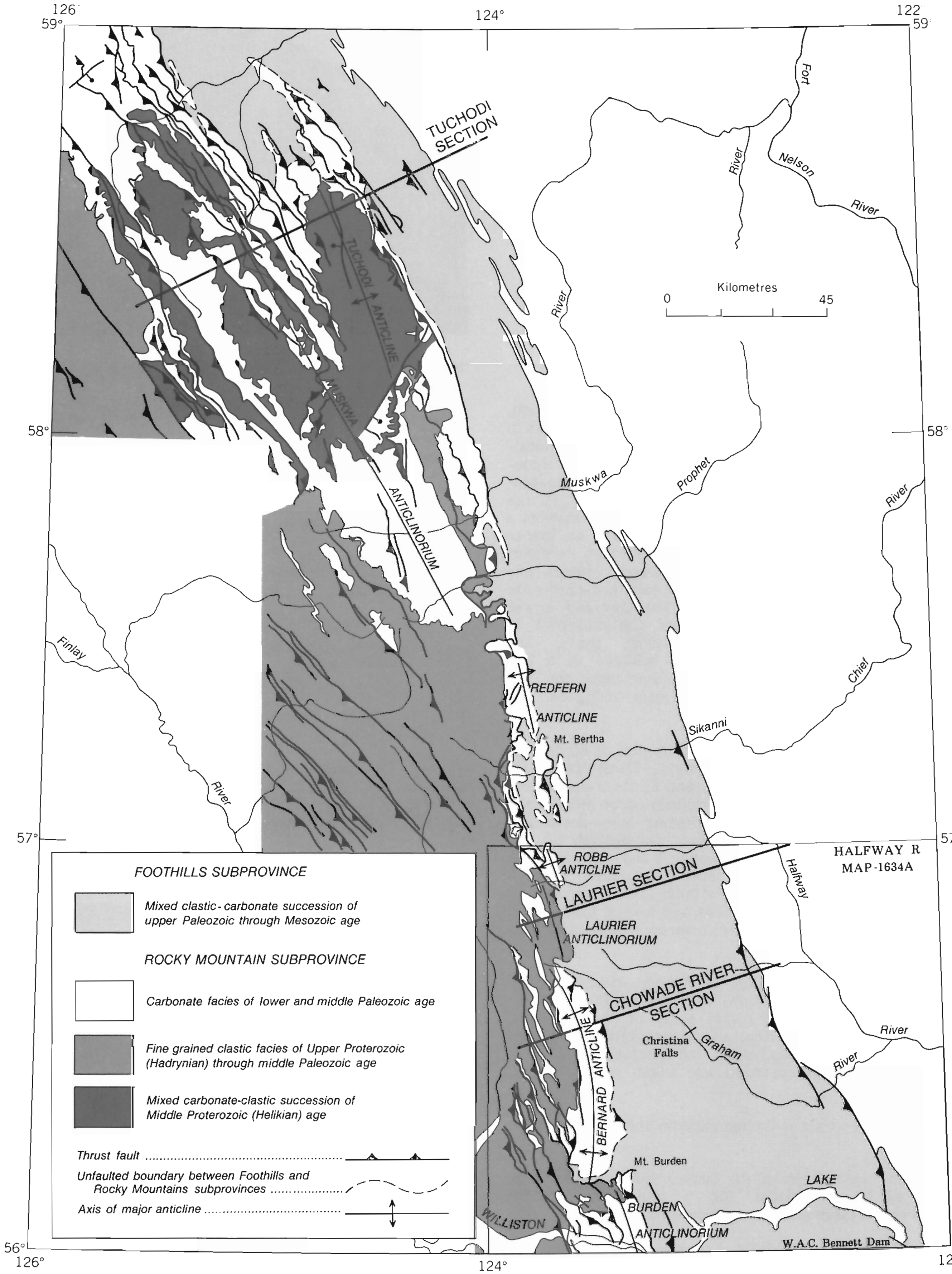
Folds and folding

Introduction

Folding has played an essential role in the structural development of the Halfway River map area. The distribution of map units within the Foothills structural subprovince is controlled by large, macroscopic folds. Much of the supracrustal shortening across the tectonic strike is accommodated by folds, and folds functioned as disharmonic strain absorbers that compensated in large part for the displacement across thrust faults at deeper structural levels. The purpose of this section is to describe how folds contribute to the gross tectonic fabric of the Halfway River region by assessing fold geometry, and the manner in which that geometry influences the preparation and interpretation of structure cross-sections. This is achieved using photographs of well exposed fold profiles as a database for the description and geometric classification of individual folds.

There are important variations in style and structural importance of folds along the strike of the Rocky Mountains. The predominance of folds in the north, especially within the Foothills structural subprovince, represents a significant departure from the geometry and style of Foothills and Front Ranges structures farther south as depicted within the Bow Valley and Waterton regions (Bally et al., 1966; Price and Mountjoy, 1970). In the south, folds are of secondary importance to thrusts; in the north, the reverse is true – the Foothills subprovince is a large-amplitude fold terrane cut by few thrusts. The difference in structural style between south and north is also reflected by the topography of each region. In the south, the Foothills have a low, rolling topography distinct from the adjacent high peaks of sculptured carbonate rocks that form the Front Ranges structural subprovince to the west (Fig. 24). In contrast, relief of 1000 m is common within the mountainous Foothills subprovince of the northern Rockies (Fig. 26) and there is little or no change in topographic expression with the adjacent Rocky Mountains structural subprovince. Pursuing this regional comparison a step further, it is noteworthy that the change from a thrust- to a fold-dominated Foothills structural style brings with it a change in the nature of the eastern limit of the Foothills subprovince as it is traced from south to north. In the south, the transition from Foothills to undeformed Plains on the east is gradual, with the eastern limit of deformation marked by the gentle, east-dipping limb of a broad syncline produced by the convergence of west- and east-dipping thrusts that form an upward closing, triangular shaped structural high called 'the triangle zone' (Price, 1981; Gordey and Frey, 1975; Fig. 23). In the north, the eastern margin of the Foothills

Figure 28. *Simplified tectonic map of part of the Northern Canadian Rocky Mountains (Halfway River (94 B), Trutch (94 G), Ware east half (94 F) and Tuchodi Lakes (94 K) map areas) showing the location of generalized structural cross-section interpretations comparing the geometry of structural development along the belt (Figs. 73, 74; from Thompson, 1981). The boundary between the Foothills and Rocky Mountains structural subprovinces is defined, usually, by the east-dipping limb of a large anticline of Paleozoic carbonate strata bounded on the east by recessive weathering shale of the Besa River Formation. There is a marked divergence in structural trend between the north-northeast trending eastern margin of the Rocky Mountains and Foothills structural belts, and northwest trending structures within the central and western portion of the Rocky Mountains subprovince (see Fig. 75 for explanation of structural divergence).*


126°
59°
124°
122°
59°
58°
58°
57°
57°
56°
126°
124°
56°
122°
TUCHODI SECTION
TUCHODI ANTICLINE
MUSKWA
ANTICLINORIUM
Kilometres
0
45
Fort
River
Nelson
River
River
Muskwa
Prophet
River
Finlay
River
Chief
Sikanni
REDFERN
ANTICLINE
Mt. Bertha
ROBB ANTICLINE
LAURIER SECTION
LAURIER ANTICLINORIUM
HALFWAY R
MAP-1634A
Halfway
CHOWADE RIVER SECTION
Christina Falls
Graham
River
River
BERNARD ANTICLINE
Mt. Burden
LAKE
BURDEN
ANTICLINORIUM
WILLISTON
W.A.C. Bennett Dam
FOOTHILLS SUBPROVINCE
Mixed clastic-carbonate succession of upper Paleozoic through Mesozoic age
ROCKY MOUNTAIN SUBPROVINCE
Carbonate facies of lower and middle Paleozoic age
Fine grained clastic facies of Upper Proterozoic (Hadrynian) through middle Paleozoic age
Mixed carbonate-clastic succession of Middle Proterozoic (Helikian) age
Thrust fault
Unfaulted boundary between Foothills and Rocky Mountains subprovinces
Axis of major anticline

subprovince is defined by an abrupt change in fold amplitude, from a large amplitude, high relief fold on the west with a steep, east dipping planar limb, to a broad, open fold on the east that is best described as a gentle warp (Fig. 26b).

There are other dissimilarities between the southern and northern Rockies. Southern Rockies folds have been described as concentric or parallel in form (Price, 1967; Dahlstrom, 1960; an exception is the study by Norris, 1971) with bounding surfaces that approximate concentric arcs. In the northern Rockies, chevron, kink and box forms predominate; limbs are planar and steep dipping, and hinge zones are narrow. The details and importance of this geometry are described in succeeding sections.

Foothills structural subprovince

Summary

At the scale of regional cross-section interpretation, the Foothills structural subprovince may be viewed as a folded multilayer consisting of five major structural-stratigraphic units: three incompetent (low viscosity) shale-dominated successions, separated by two, more competent (higher viscosity) carbonate-sandstone successions. Large folds are outlined by the Carboniferous Prophet Formation, a carbonate succession 300 to 600 m thick and, at higher stratigraphic levels, by the Triassic Liard through Pardonet formations and the Jurassic and Cretaceous Minnes and Bullhead groups, all sandstone-siltstone-carbonate successions. The alternate units of Devonian and Lower Carboniferous Besa River shale, Upper Carboniferous to Triassic Stoddart through Toad-Grayling shale, and Cretaceous Fort St. John Group shale, behaved as a less viscous confining medium forced to accommodate the fold styles imposed by the more rigid units (refer to Figure 3 for summary of stratigraphic relationships).

Chevron or box-style folds with limb dips of 60 degrees or more characterize the structural style. Thrust faults are of minor importance (in most areas) and faults transverse to the structural grain are absent. Anticlines occur individually and as anticlinorial complexes; separating them are broad, simple, flat-bottomed synclines. Fold wavelength (at the cross-section scale) varies between 4 and 10 km; fold amplitude is also variable; the largest anticlines reach a maximum of 1.8 km. The Foothills fold complex is interpreted as having been detached and having ridden over less deformed Devonian carbonate rocks on a cushion of Besa River shale.

In the following discussion, an attempt will be made to describe the essential structural features of the Foothills subprovince, as well as to relate some of the empirical observations to the theoretical and experimental work that has been done on the mechanics and mechanisms of fold development. Several questions are worth noting in this regard:

1. How did the straight-limbed chevron and box-style folds evolve?

2. Can the wavelength of the largest folds be related to theoretical predictions for a multilayer of the dimensions described for the Foothills?

3. Did the folds form serially or simultaneously?

4. Why is there such a sudden change in fold amplitude along the eastern margin of the Foothills subprovince?

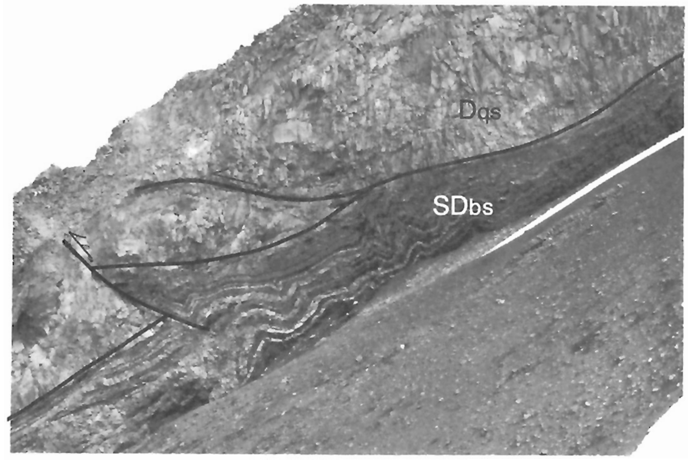


Figure 29. *Massively bedded dolomitic quartz sandstone (map unit Dqs) overlying well bedded siltstone and dolomitic quartzite (SDbs) illustrating the effect of competency contrast on style of deformation. The quartzite is detached from the siltstone and is shortened across discrete contraction faults; the siltstone is shortened by folds. Structural disharmony between competent and incompetent strata occurs at all scales of observation and has had a significant effect on the surface distribution of map units. (GSC photo no. 819-43).*

5. Is this fold belt typical of other external fold belts in the world?

6. How does the geometry of Foothills structures influence the preparation of subsurface cross-section interpretations?

7. How do Foothills folds relate mechanically to the structural evolution of the Rocky Mountains subprovince?

Answers to these and other questions are not necessarily obvious, but parallels exist between experimental and field observations. The first step is to adequately describe and classify the fold structures.

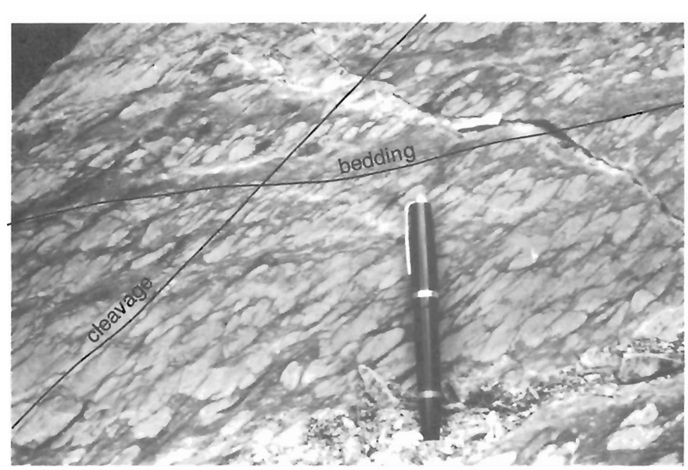


Figure 30. *Cleaved Kechika Group calcareous siltstone and limestone with original bedding (subhorizontal) partly transposed into the plane of the cleavage. (GSC photo no. 2112-1).*

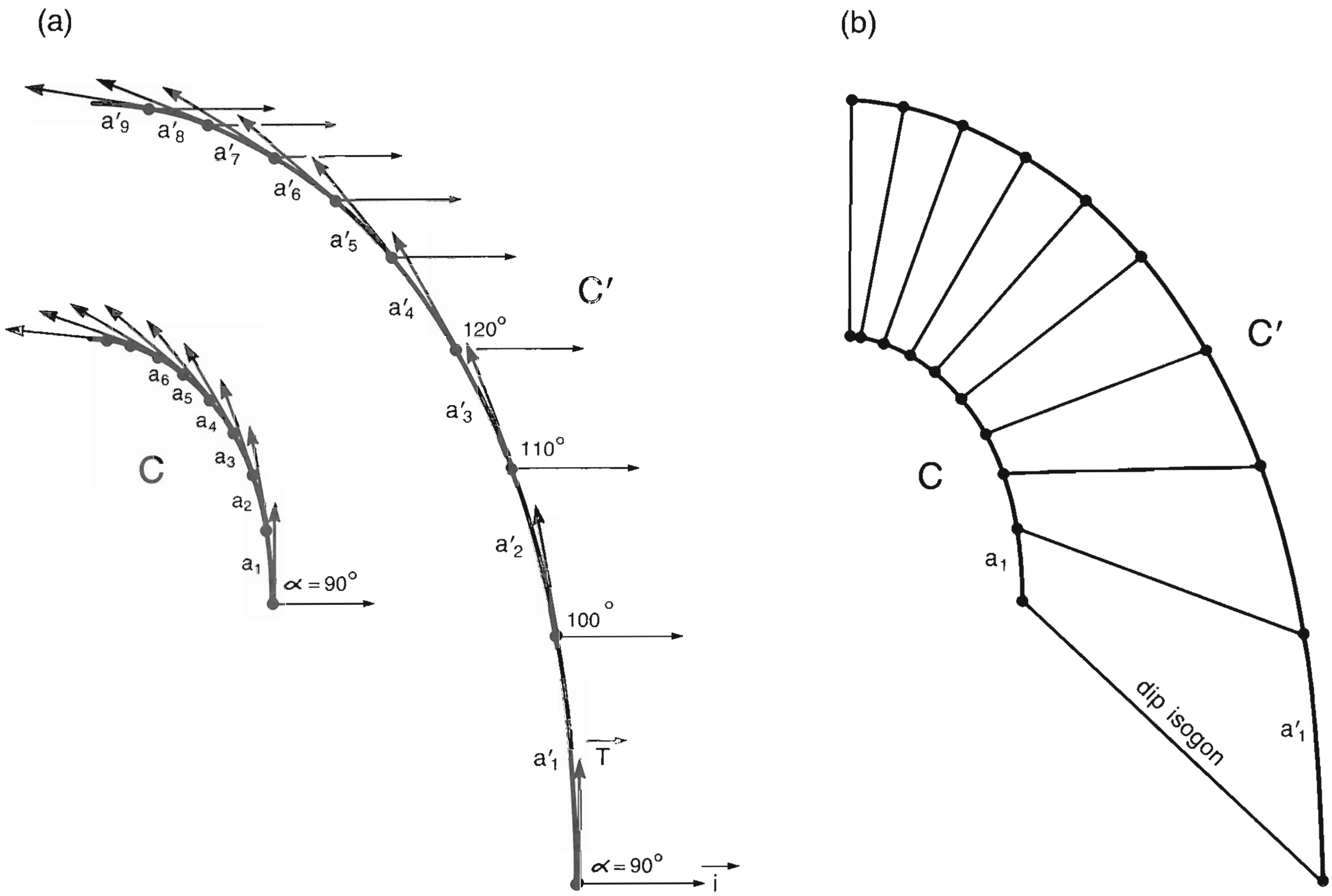


Figure 31. *Illustration of how relative curvatures of adjacent arcs can be assessed. (a)* ***i*** *is a reference vector,* ***T*** *is a tangent vector, α is the angle subtended by* ***i*** *and* ***T***, ***a*** *is the arc length required for α to change 10°.* ***C*** *is said to turn more quickly than* ***C'*** *because the length of* ***a*** *is less than that for the corresponding* ***a'***. *Therefore, for adjacent points on* ***C*** *and* ***C'*** *(i.e. for points having equal inclinations of* ***T***), ***C*** *has greater curvature than* ***C'***. *(b) Lines joining points of equal inclination on adjacent curves are called dip isogons. Dip isogons converge toward the curve with the greatest curvature. The rate of convergence is a measure of the magnitude of the difference in curvatures.*

The use of dip isogons and orthogonal thickness to describe the geometric properties of fold profiles

The object of this section is to describe a method of fold description and classification that does not depend upon a database of systematically gathered mesoscopic (outcrop scale) fabric measurements. The physical size of the Halfway River region precluded detailed investigation of individual fold structures, yet internally consistent and realistic cross-section interpretations require a firm knowledge of the basic geometric properties of major fold structures within the region. The approach described here[1] is taken from an excellent discussion of fold classification by Ramsay (1967) with minor modification to suit the needs of this project.

The well exposed, mountainous topography of the Halfway River region makes it practical to use photographs of structural features for the purpose of structural analysis. Many large folds are transected by deeply eroded valleys that expose details of fold profile geometry – photographs of these profiles form the analytical database.

The shape of an individual folded layer is normally described in terms of the profile shape of the inner and outer bounding surfaces. If the bounding surfaces have exactly the same shape, the fold is called 'similar'. If the surfaces are parts of concentric arcs (that is, layer thickness around the fold is constant), the fold is termed 'parallel' or 'concentric'. These two fold forms are specific geometric forms within an infinite spectrum of possibilities that are a function of the relative curvatures of the bounding surfaces. Fold description and classification depends, in large part, on an assessment of the relative curvatures (in two dimensions) of bounding surfaces.

Curvature is a measure of how quickly an arc changes inclination; it is independent of the magnitude of the change. In Figure 31, tangent vectors $\vec{T}$ are drawn for each 10 degree

[1]Only the essential elements of the classification scheme are described here, the reader is referred to Ramsay (1967) for a comprehensive treatment of the subject.

change in inclination of curves C and C'. That is, the angle measured between $\vec{T}$ and a reference vector $\vec{I}$, having a fixed inclination, changes by 10 degree increments. The arc lengths a and a' between each successive position of $\vec{T}$, give a measure of how quickly $\vec{T}$ changes position. In this case, curve C is turning more quickly, and has greater curvature, than C' because the length of a is everywhere smaller than the corresponding a'. It is also important to note that the curvatures of C and C' are increasing in the up direction; that is, a and a' decrease systematically (in the same direction) along each curve. If tangent vectors could be drawn at infinitely small intervals to each curve, the curvature of each infinitely small interval could be recorded by measuring successive changes in α. In fact, if the curves C and C' could be expressed in terms of vector valued functions, one could calculate a value of curvature for every point in each curve. Thus, when comparing the curvatures of C and C' in Figure 31, it is mathematically more precise to say that for points of equal inclination on C and C', C has a greater curvature.

The geometric analysis of fold profiles requires that the relative curvatures of adjacent (or bounding) arcs be determined. Fortunately, this does not necessitate the calculation of absolute values of curvature for the bounding arcs. Relative curvatures can be assessed using two related and simple graphical techniques. One is the construction of dip isogons, and the second is the measurement of changes in orthogonal thickness t.

Dip isogons are lines drawn between bounding arcs that connect points of equal inclination. This is accomplished by drawing a pair of tangents of equal slope and then joining the tangent points with a straight line. This has been done for curves C and C' in Figure 31b by joining the tangent points constructed in Figure 31a. In this case, the dip isogons form an array of "radiating spokes". Because the arc lengths along C and C' between adjacent isogons represent the distance required for each arc to turn through 10 degrees, the isogons must necessarily converge toward the curve with the greatest curvature – in this case C. The rate of isogon convergence is a measure of the disparity in curvatures between the arcs. If the isogons are parallel, then the curvatures are equal. In Figure 32, a cursory inspection reveals that curves A and A' change relative mean curvatures along their lengths: along segment (1) the dip isogons converge toward curve A, therefore it has the greatest mean curvature; along segment (2) the isogons are essentially parallel, which means that A and A' have equal curvatures; and along segment (3) the isogons converge toward A' because it now has the greatest mean curvature. In summary, two generalizations can be made: 1. dip isogons converge toward the arc with the greatest mean curvature, and 2. the arc lengths or segments subtended by adjacent dip isogons decrease in the direction of increasing curvature.

The ease with which dip isogons may be constructed makes their use a very rapid and practical method for comparing variations in the relative curvatures of adjacent arcs. When this technique is applied to folds, it is convenient to describe the geometry of the isogons with reference to the profile trace of the axial plane. Therefore dip isogons are said to converge or diverge toward the axial trace.

Dip isogons provide the framework for subdividing fold profiles into three basic classes (Fig. 33). Class 1 folds have convergent isogons (the mean curvature of the inner arc is greater than that of the outer arc); Class 2 folds have parallel isogons (a unique fold class in which the mean curvatures of outer and inner arcs are equal); and Class 3 folds have divergent isogons (the curvature of the outer arc is greater than that of the inner arc).

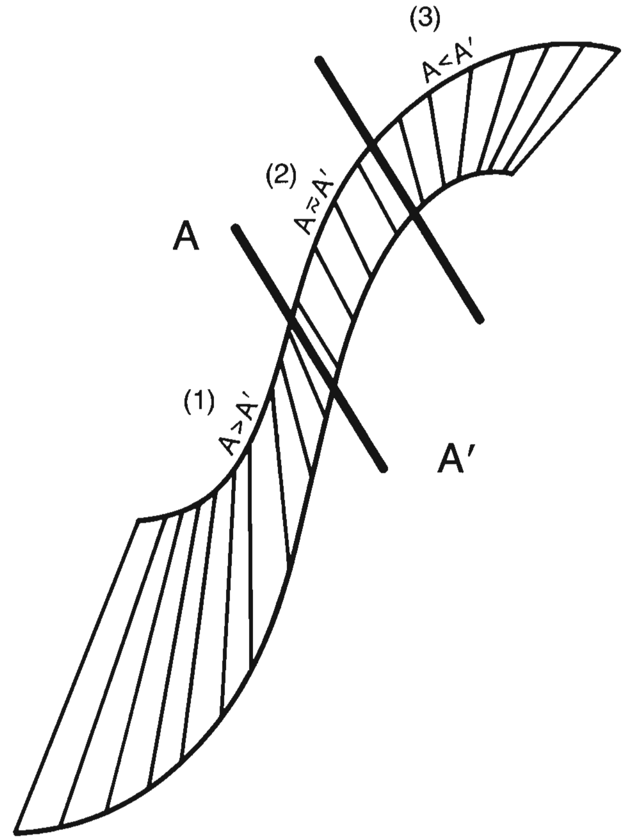


*Figure 32. Illustration of how dip isogons can be used to determine the relative changes in curvature between adjacent arcs. **A** has greater relative curvature than **A'** in segment (1), the curvatures are essentially equal in segment (2), and **A'** has greater relative curvature than **A** in segment (3).*

In the foreland part of most alpine orogens, thick, competent, clastic and carbonate sequences are generally folded into Class 1 folds. In some instances, units form concentric or parallel folds in which bed thickness is constant around the fold. This is a unique type of Class 1 fold and provides a reasonable basis for subdividing Class 1 folds into three types (Fig. 33): Class 1A, in which isogons are strongly convergent; Class 1B, which are parallel folds and require the bounding arcs to have curvatures that permit original bed thickness to be maintained; and Class 1C, in which dip isogons are weakly convergent. Therefore Class 1B folds are unique, and separate Class 1A folds, in which limbs thicken away from the hinge zone, from Class 1C folds, in which limbs thin away from the hinge zone.

The advantage of the dip isogon method of fold classification is that it allows rapid analysis of a complex fold containing many deformed surfaces. Commonly, what appears as a simple fold is actually a composite of several classes, all illustrated in Figure 34. A drawback to this approach is that it does not provide data that is easily measured or graphed for comparative purposes. To compensate for this, the orthogonal thickness 't' can be measured at the same time the dip isogon is constructed.

Orthogonal thickness t is the perpendicular distance measured between adjacent tangents having the same inclination (Fig. 34). It records changes in relative curvature by determining systematically how limb thickness changes

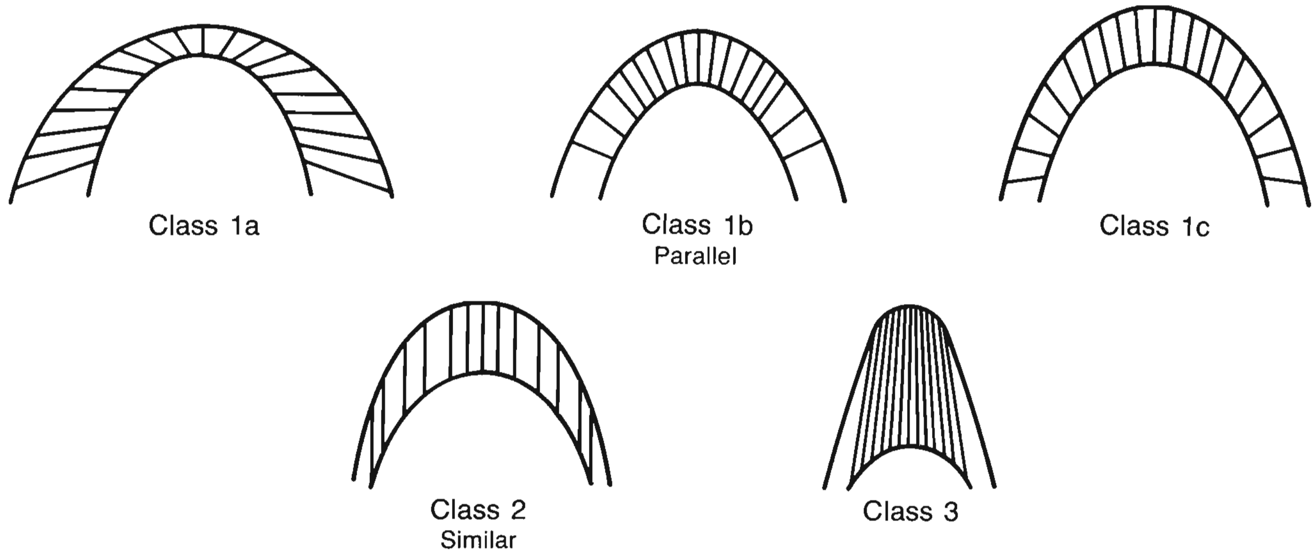


Figure 33. *Classification of folds based on the convergence or divergence of dip isogons. In Class 1 folds, isogons converge toward the inner fold arc, in Class 2 folds (similar) the isogons are parallel, and in Class 3 folds the dip isogons diverge toward the inner fold arc (from Ramsay, 1967).*

around a folded layer. Figure 34 demonstrates how t can be measured when a dip isogon is being constructed. Rather than plot values of t directly, it is more convenient to plot the ratio t/t_o ($=t'$) where t_o is the orthogonal thickness measured at the fold hinge. In this way, results from several folds may be plotted onto the same diagram for comparative purposes. The composite fold in Figure 34 consists of two Class 1A folds and a Class 3 fold. The overall geometry is that of a Class 1B (parallel) fold.

Because Class 1B and Class 2 folds are unique fold forms, they plot as lines on a classification diagram in which t/t_o is plotted against inclination α of the tangent pair. Figure 35 is a summary classification diagram in which the field for each fold class is plotted as well as the t/t_o values for the composite fold in Figure 34. In the following section, several folds from the Halfway River area are analyzed using dip isogons and the measurement of orthogonal thickness.

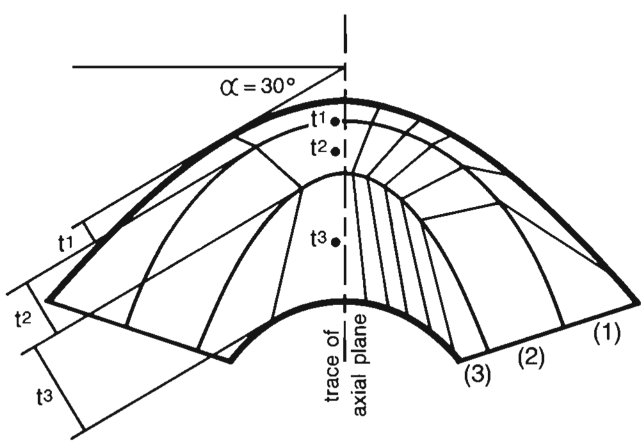


Figure 34. *Illustration of the geometrical relationship between the dip isogon and the orthogonal thickness of a layer. When a dip isogon is constructed across a layer, the orthogonal thickness* ***t'*** *can be measured by extending tangent lines and measuring the perpendicular distance between them.* ***α*** *is the angle measured between the tangent lines and the trace of the axial plane of the fold.* t_o' *is the orthogonal thickness measured along the trace of the axial plane.*

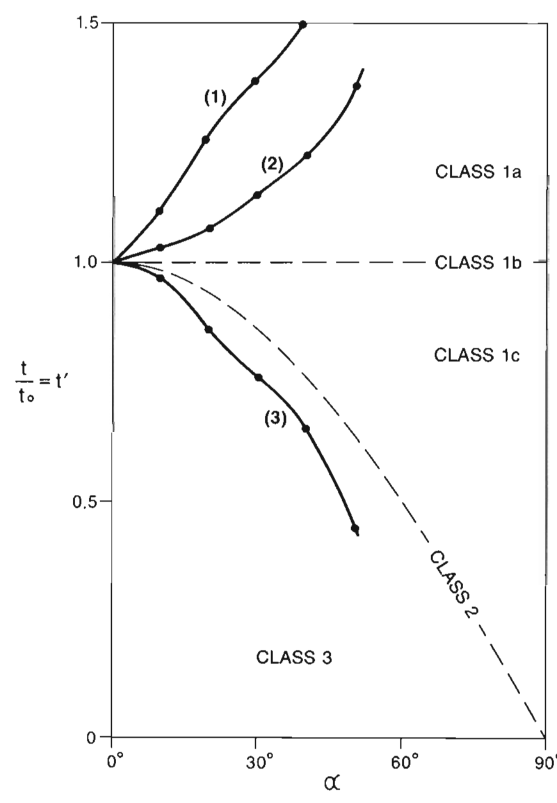


Figure 35. *Plot of* ***t'*** *versus* ***α*** *for each of the layers shown in Figure 34.*

The geometry of Foothills folds

Individual map scale folds within the Foothills and Rocky Mountains of the Halfway River map area are composites of Class 1B, 1C, and Class 3 folds which, when considered together, approximate the overall geometric properties of a Class 2 (similar) fold. Figure 36 is a summary plot of t/t_0 versus α for nine macroscopic folds. It shows that the variation in fold form is confined, for the most part, to a narrow interval that contains and parallels the plot of a Class 2 fold form. Unlike terranes in which parallel (Class 1B) folds predominate, Class 2 fold geometry accommodates the "room problem" that occurs within the hinge zones of developing parallel folds, allowing Class 2 folds to extend, in theory, to any depth. For this reason it is not possible to calculate (with confidence) the depth of folding within the Halfway River area using the decreasing radius of curvature method outlined by De Sitter (1964).

The geometry of Foothills folds is remarkably consistent in orientation, attitude and form, regardless of stratigraphic position or location within the subprovince. Consequently, surface geometry has been used as a guide in the preparation of subsurface cross-sections. In contrast to the Foothills, the form and attitude of folds within the Rocky Mountains subprovince is more varied and is controlled to a greater degree by rock type. Nevertheless, most folds approximate a Class 2 form.

Procedure. Clear plastic film was laid over enlarged black and white photographic prints of fold profiles. Some of the more distinctive layers were traced onto the film and dip isogons constructed across them. On a second overlay of tracing paper, values of t were measured across selected layers (usually a cliff forming one and a recessive one) and t/t_0 calculated.

Results. Results obtained from the analysis of nine large folds are presented below. Fold style, amplitude, and topographic expression do not change significantly with stratigraphic level or position across the tectonic strike of the Foothills subprovince. Folds in Triassic strata adjacent to the Peace River (Fig. 40) exhibit the same geometric properties as folds at the same stratigraphic level exposed along the Halfway River (Figs. 38, 39). These in turn compare equally well with folds in rocks of Cretaceous (Fig. 37), Permian (Fig. 44) and Carboniferous (Figs. 40-43) age, and individual folds are consistent in form along strike (Figs. 41, 42).

Folds of the size and form shown in Figure 37 typify large parts of the eastern margin of the Foothills subprovince. Box-style and chevron forms may occur individually or as parts of larger, anticlinorial closures. The intervening synclines form broad, flat valley bottoms. In Figure 37, the dip isogons converge downward across layers (1) and (2) to layer (3) where the isogons become nearly parallel and the fold form changes from Class 1B to Class 2 (approximate). This inward change from a box-style to a chevron form is common; orthogonal bed thickness is nearly maintained (Class 1B) around the box-style form because the total bending is divided between two hinges. Where only one hinge is present within the fold core, a space problem develops and orthogonal thickness must change locally to accommodate it, resulting in a progressive inward change from Class 1B toward Class 3, as illustrated by the orthogonal thickness plot that shows the right hand side of the hinge (layer 2) assuming a Class 3 geometry.

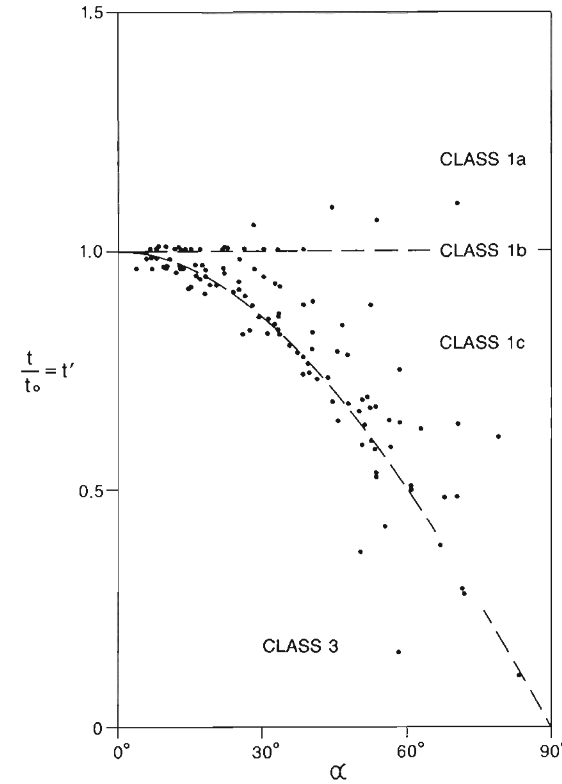


Figure 36. *Summary plot of* t' *versus* α *for Foothills folds within the Halfway River map area.*

The results from Figure 38 illustrate a pattern that is consistent within box-style folds throughout the area. The competent, cliff forming units approximate Class 2 (similar) folds. Another feature, common to many large folds (Fig. 38) is curvature of the axial trace.

In places, Triassic strata have been folded into a succession of closed, chevron-shaped folds. Folded layers (1) and (2) in Figure 39 are part of the Liard and Charlie Lake formations (the competent units of Figure 38). The steep dip of the fold limbs requires that the hinge zone approximates a Class 2 form in order to accommodate the space problem caused by such severe curvatures. The lack of discrete fault ruptures within the hinge area suggests that thickening is accommodated by flowage into the hinge of material within thin, less competent layers.

The anticline shown in Figure 40 is the first large exposure of Mississippian Prophet carbonate rocks along the Halfway River transect. Despite the massive character of the carbonates, the inner layers approximate a Class 2 form and the outer layers a nearly parallel Class 1B form. The axial trace is curvilinear, the limbs planar and the hinge area narrow and unfaulted. Its geometry compares well with folds in Triassic and Cretaceous strata, in spite of the difference in stratigraphic level, lithological character and position within the Foothills subprovince. Consequently, the Class 2 and composite Class 2 - Class 1B forms serve as a reasonable model that may be applied to the interpretation of fold form within Mississippian and Triassic strata within the subsurface of the Foothills subprovince.

The Laurier Anticline is a large, box-style anticline in the western Foothills, formed of Carboniferous Prophet Formation carbonates with a core of Besa River shale. Figure 41 shows the eastern half of the anticline (viewed

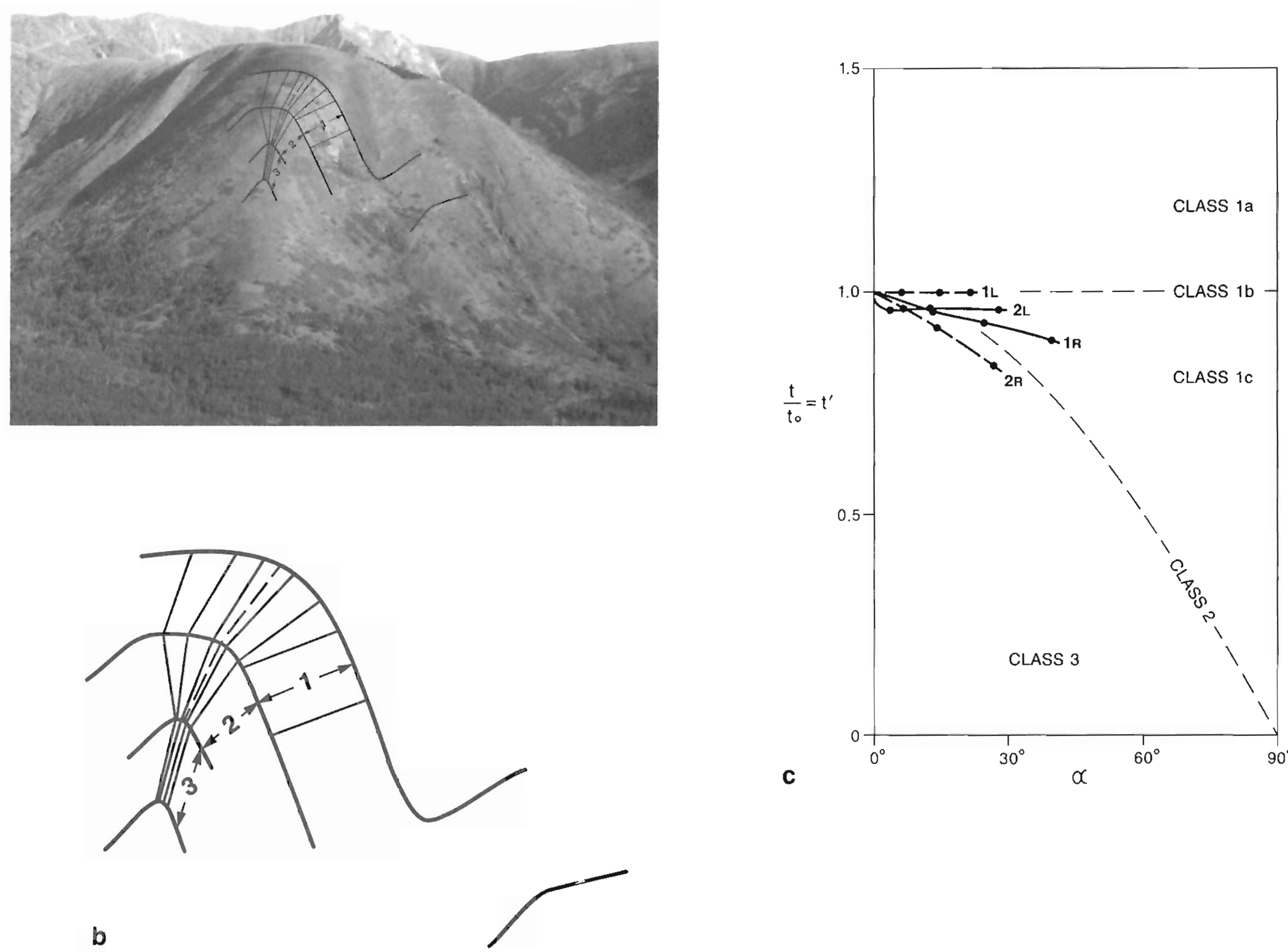


Figure 37. *(a) Northward view of anticline-syncline fold pair, Cretaceous Minnes Group, located on north side of the Halfway River valley. (GSC photo no. 657-3). (b) Dip isogons drawn across fold profiles traced from (a), dashed line is trace of axial plane. (c) Fold class plot for each numbered fold limb (L = left of axial plane, R = right of axial plane) on a fold class diagram allowing comparison of Foothills folds with the fundamental fold classes (see text for explanation).*

from the north), and Figure 42 shows part of the western half (viewed from the south). The anticline has a strike length of approximately 20 km and a maximum width of 5 km. Both limbs are vertical. In Figure 41, it is possible to note the change in stratigraphic character of the rocks toward the fold core. The outer units consist of massive carbonates that become progressively more carbonaceous and incompetent inward. The geometries described in Figures 37 through 40 apply to the Laurier Anticline also. Limbs are planar, hinge areas narrow, and most layers approximate a Class 2 form. Figure 42 illustrates that the same competent unit may be a composite of Class 1B and 1C (layer 1) and Class 3 (layer 2) forms despite the lack of an obvious incompetent unit separating them. In Figure 41, the dip isogons show an alternating diverging-converging pattern across adjacent units. In this case, the overall form is Class 2 but in detail the fold is a composite of alternating Class 3 and Class 1C folds.

At this point, it is clear that fold styles within the Cretaceous, Triassic and Mississippian stratigraphic successions are identical. They are composites with an overall Class 1C verging on Class 2 form. Hinge areas are narrow and unfaulted, limbs are planar and steep dipping, and the folds are either chevron or box shaped.

Westward, adjacent to the mountain front, the variation in fold shape increases coincident with an increase in the proportion of shale to carbonate and sandstone. Figure 43 shows a more westward facies of the Mississippian Prophet Formation with a greater proportion of dark weathering, calcareous limestone. Figure 44 shows disharmonic, tight, overturned to recumbent folds within shales and siltstones of the overlying Carboniferous Stoddart Group.

The folds shown in Figure 43 are part of a larger fold complex, shown in Figure 58, that is detached from flat lying carbonate rocks below. Despite the complex look of these folds and their vergence eastward, the geometric form of the folds is nearly the same as that for the upright box and chevron forms to the east. The dip isogons are weakly convergent across the more competent units and diverge slightly across the incompetent units. Plots of t/t_0 versus α show the competent layers to be Class 1C verging on Class 2, and the thicker, incompetent units to be Class 2 and Class 3 folds. Figure 58 illustrates two important features of Foothills folds: 1. folds are riding on detachment surface(s), and 2. there is no geometric method of predicting the level of the detachment surface. Therefore, it would be a mistake to assume that any of the large Foothills folds is concentric, and to use this assumption to calculate a depth of detachment.

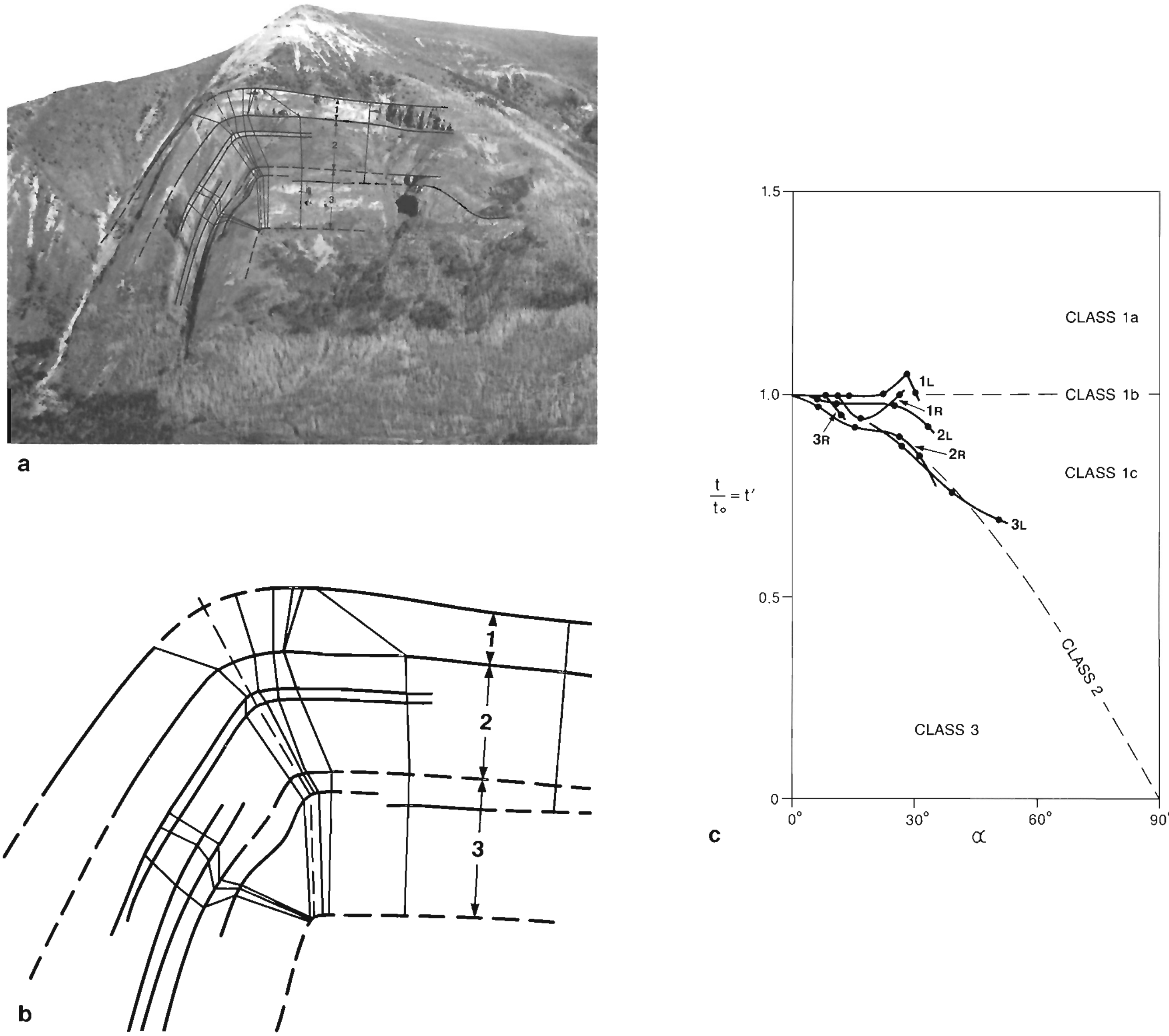


Figure 38. *(a) Northward view showing west half of box fold in Triassic strata, located on the north side of the Halfway River valley. Layer 3 is Liard Formation, strata beneath Layer 3 belong to the Toad Formation, and strata above Layer 3 belong to Baldonnel and Pardonet formations. Curvature of the trace of the axial plane (dashed line) is common to most large Foothills folds. (GSC photo no. 657-13). (b) Dip isogons drawn across fold profiles. (c) Fold class plot for each numbered fold limb.*

Wavelength. It has been demonstrated theoretically and in experiments using models, that a single layer embedded within a medium of lower viscosity will buckle with a characteristic wavelength that is a function of the relative viscosities of the materials as well as the thickness of the more competent layer (Biot, 1961; Ramberg, 1962). This approach has been extended to the analysis of multilayers, in which several competent layers having the same thickness are embedded within a less viscous material. The dominant wavelength of folds can be predicted using the following equation:

$$W_d = 2\pi t \left(\frac{n\mu^1}{6\mu_2}\right)^{1/3}$$

where W_d is the dominant wavelength, t is the thickness of the competent layers, n is the number of competent layers, and μ_1, μ_2 are the viscosities of the competent and incompetent material, respectively.

It is a very big step from the centimetre scale of experimental results in which viscosity, layer thickness and strain rate can be controlled, to a mountain belt, where these parameters can only be approximated or guessed at. It is not surprising then, that wavelengths are variable and the correlation with theoretical predictions weak. Figure 45 shows smooth curve envelopes of large fold profiles taken from structure sections (Fig. 73). Wavelengths vary between 4 km and 10 km, with most values near 7 km. Assuming a five layer model with two competent layers (discussed earlier) a viscosity contrast $\mu_1 : \mu_2$ of 5:1, and a competent layer thickness of 800 m, the wavelength predicted using the Biot equation is 6 km. If the viscosity contrast chosen were 10:1, W_d = 7.5 km. These predictions are in accord with some

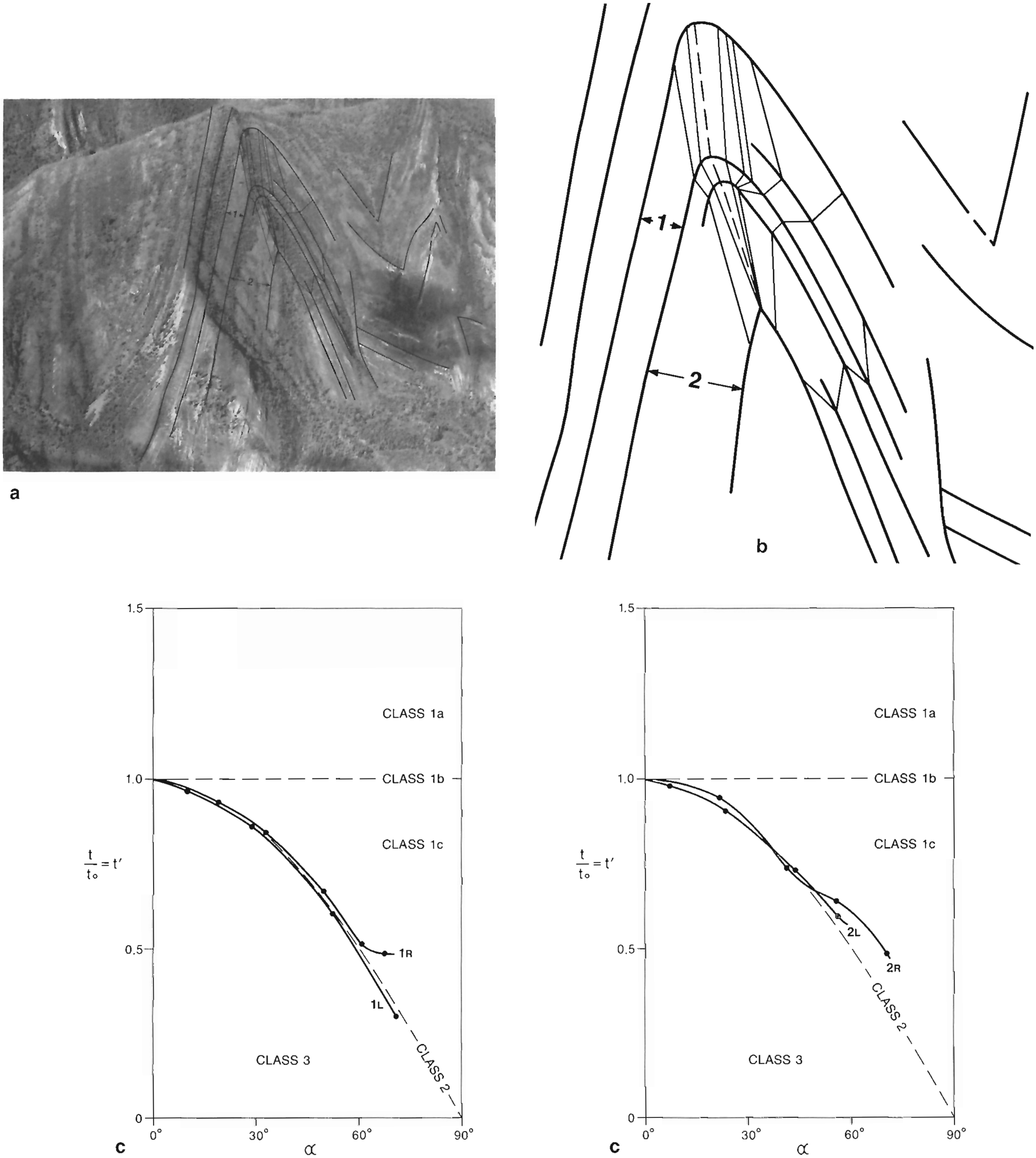


Figure 39. (a) Northward view of chevron folds in Triassic strata located on the north side of the Peace River valley. (GSC photo no. 657-19). (b) Dip isogons drawn across fold profiles. (c) Fold class plot for each numbered fold limb.

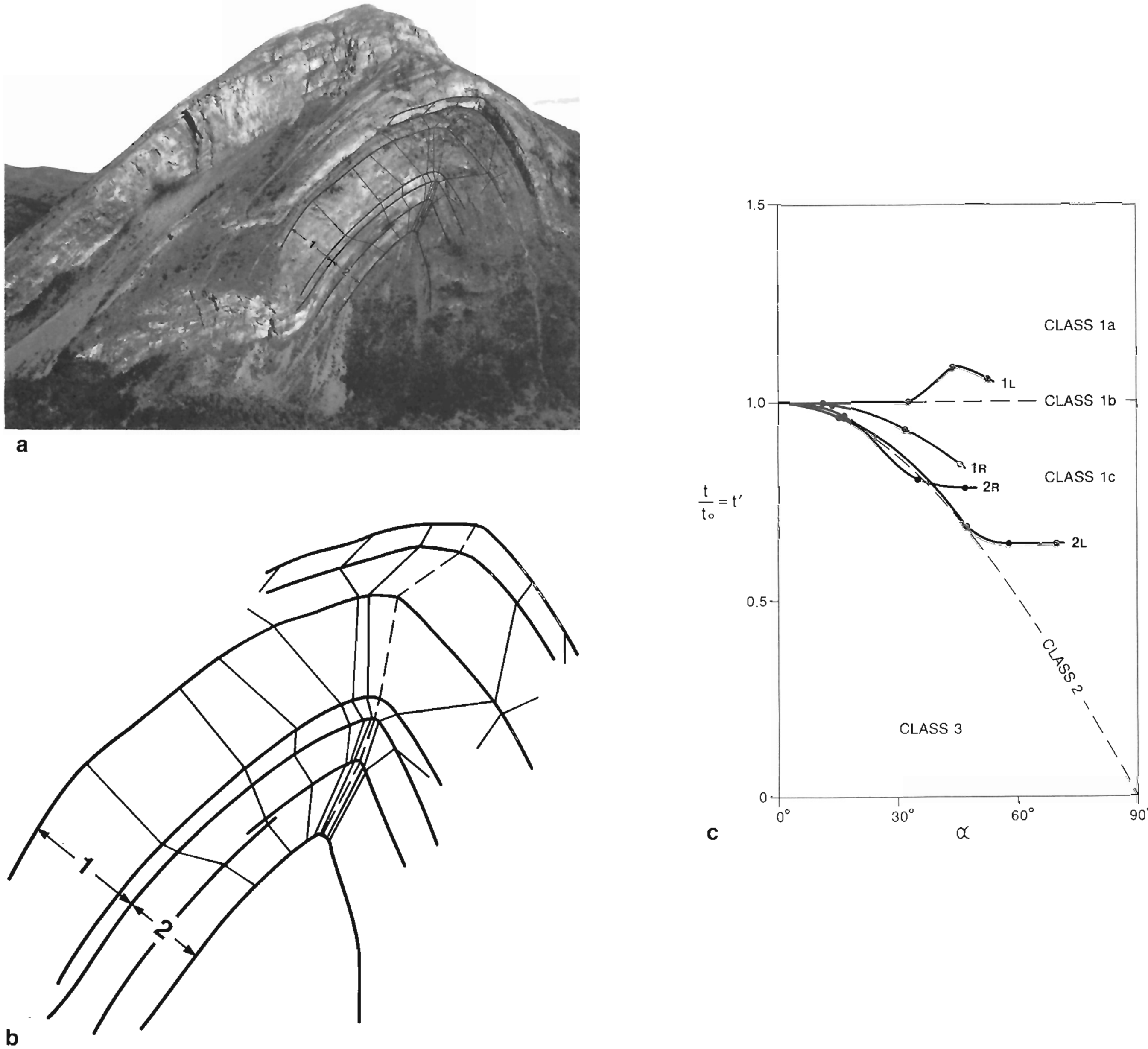


Figure 40. (a) Northward view showing an anticline of Mississippian Prophet Formation, located on the north side of the Halfway River valley. Besa River shale is exposed at the base of the slope in the core of the anticline; the upper cliff corresponds to map unit CP2 ; the middle recessive and lower cliff correspond to unit CP1 . Note curvature of the trace of the axial plane. (GSC photo no. 819-81). (b) Dip isogons drawn across fold profiles. (c) Fold class plot for each numbered fold limb.

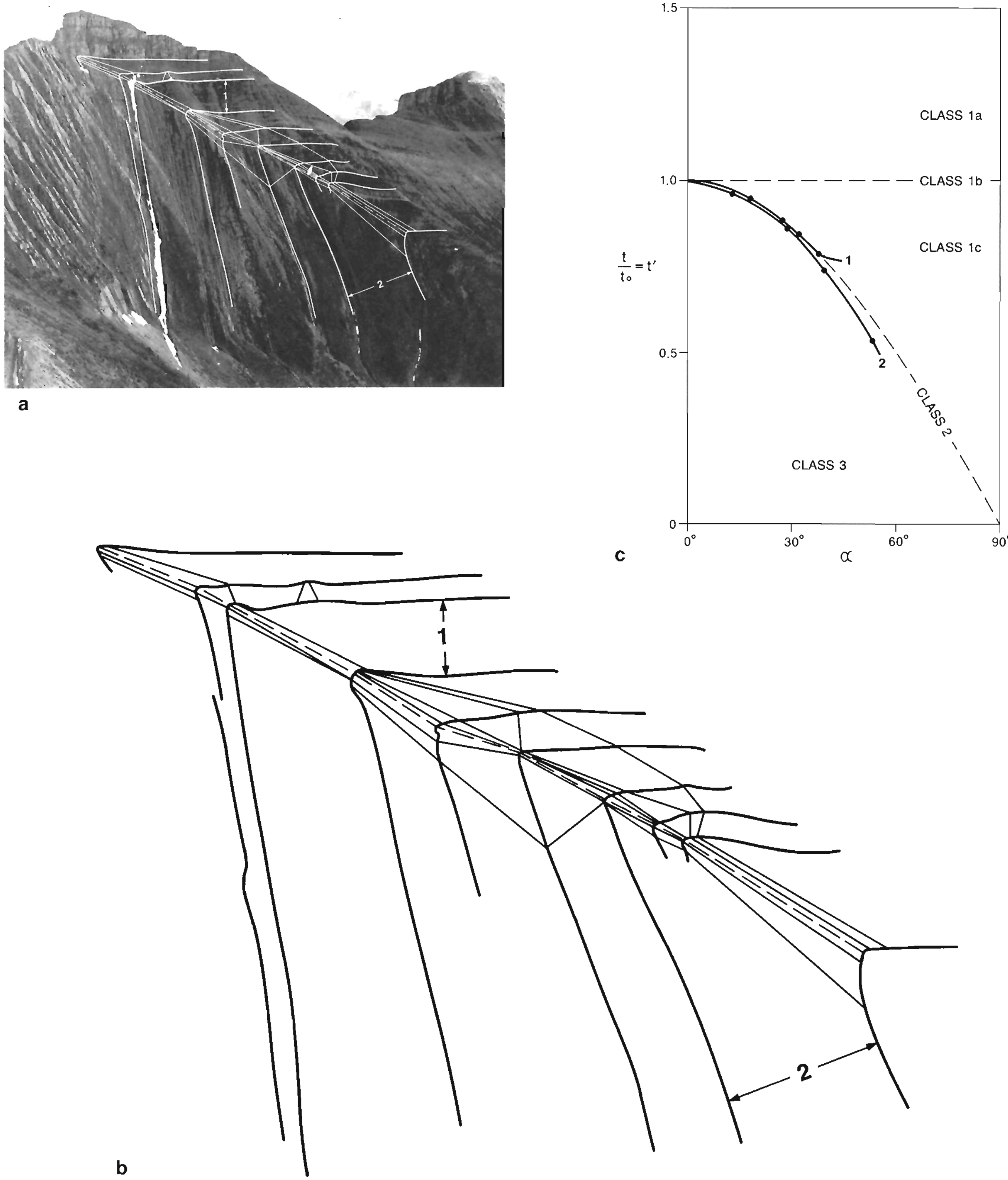


Figure 41. (a) Southward view of the eastern half of a box fold of Prophet Formation carbonate and underlying Besa River shale, located about 5 km south of the headwaters of Needham Creek. (GSC photo no. 819-33). (b) Dip isogons drawn across fold profiles. (c) Fold class plot for each numbered fold limb.

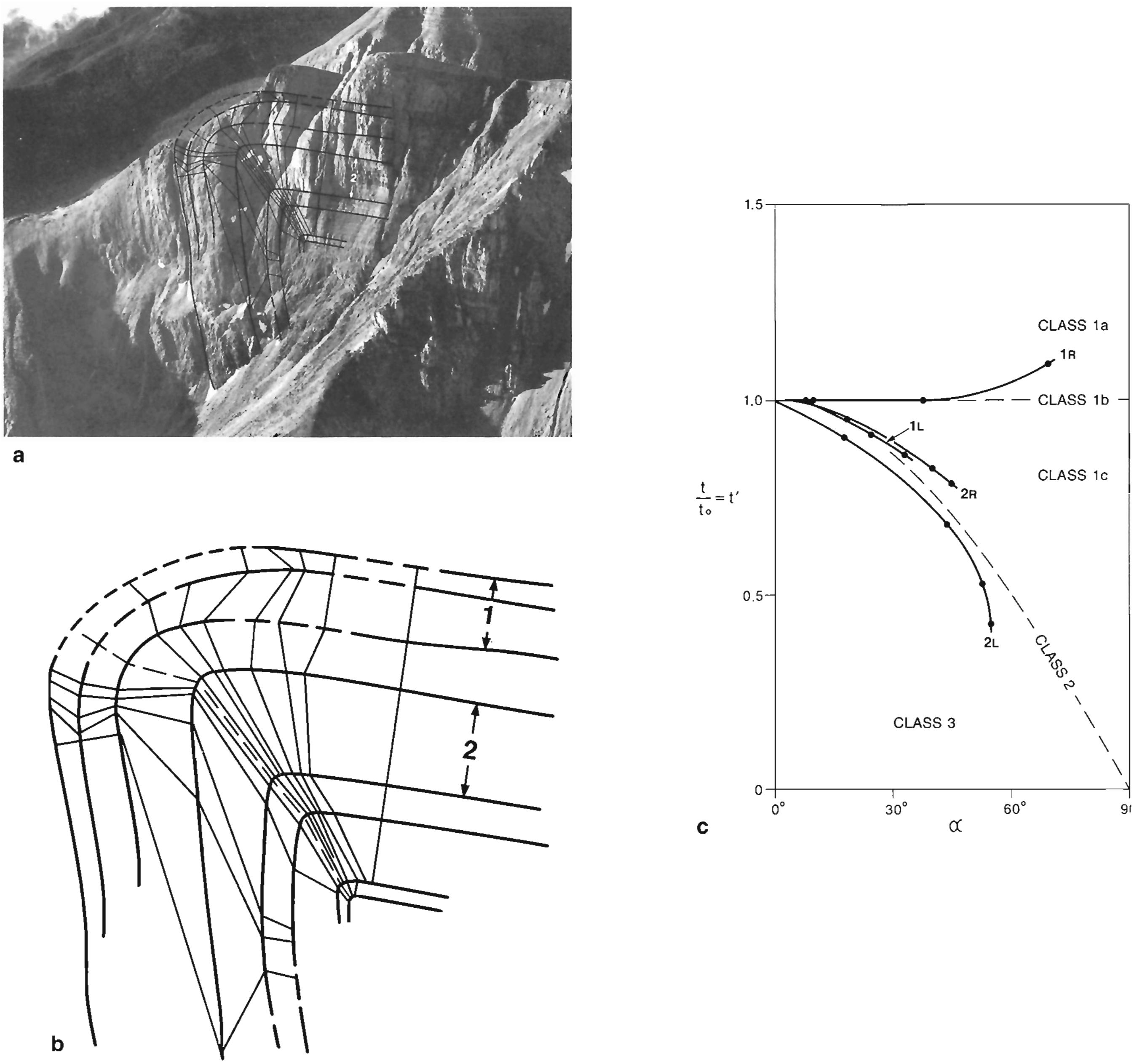


Figure 42. (a) Northward view showing the western half of the box fold shown in Figure 41. (GSC photo no. 819-81). (b) Dip isogons drawn across fold profiles. (c) Fold class plot for each numbered fold limb.

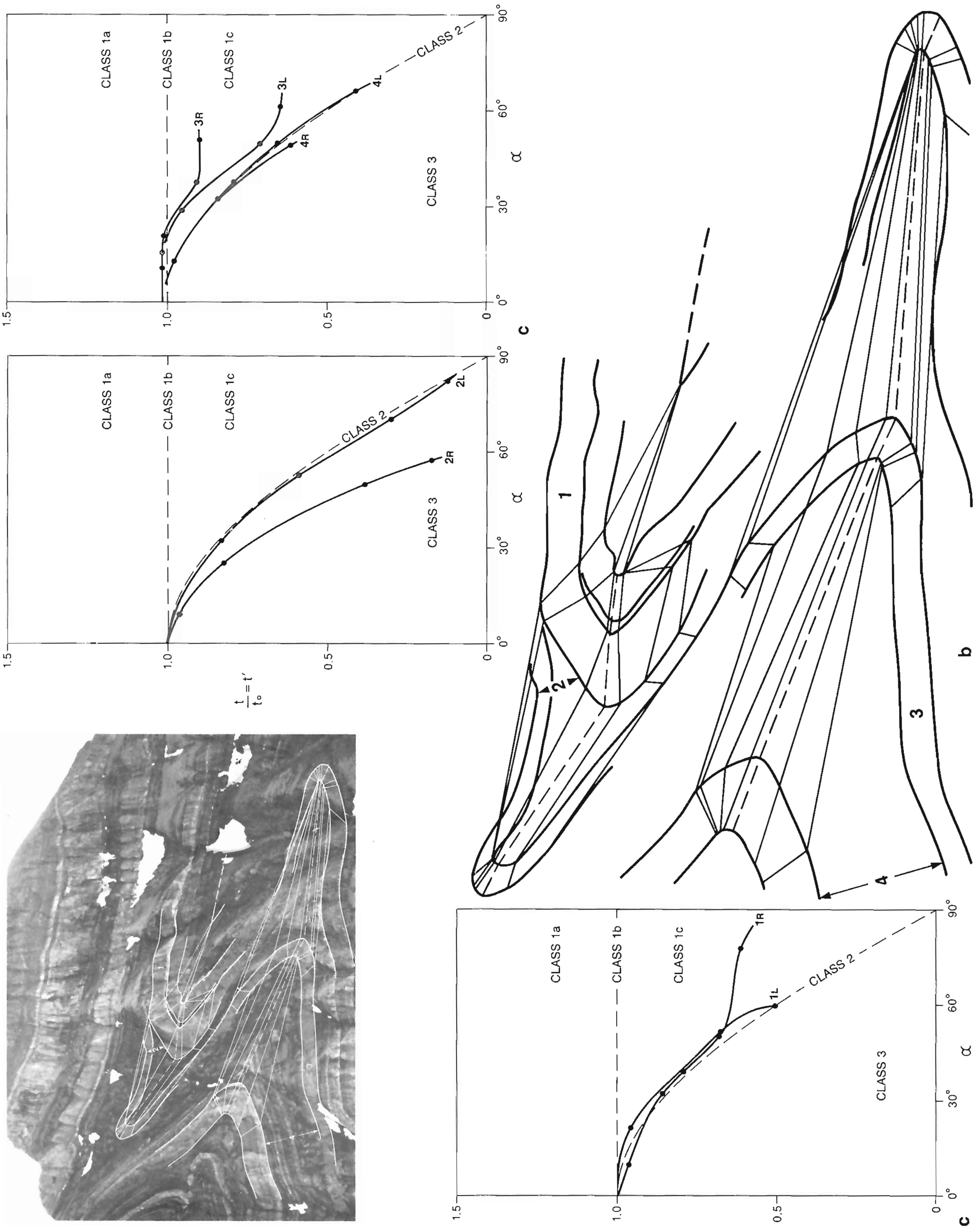


Figure 43. *(a) Detached fold complex of Mississippian Prophet Formation carbonates, headwaters of Nabesche River. (GSC photo no. 819-141). (b) Dip isogons drawn across fold profiles. (c) Fold class plot for each numbered fold limb.*

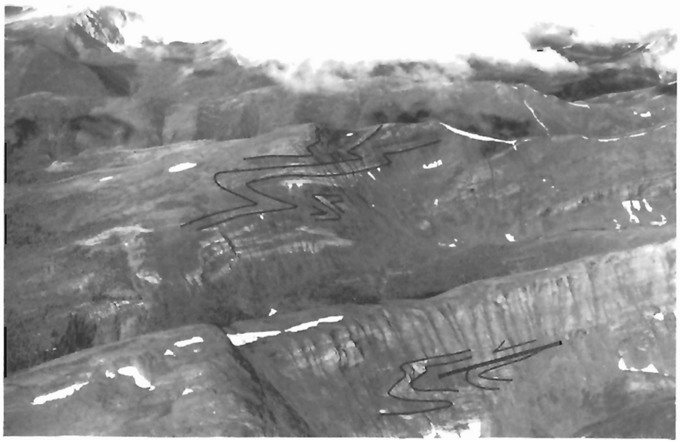


Figure 44. Southward view of overturned, disarticulated folds of Stoddart Group shale and siltstone, 5 km east of Mt. Burden. (GSC photo no. 819-71).

of the fold wavelengths measured, but it is difficult to know how much significance this agreement has. The theoretical treatment by Biot (1961) assumes constant layer thickness, constant viscosity contrast, homogeneity of individual layers, and a perfectly planar initial geometry. None of these assumptions applies in detail to the Foothills subprovince. Figure 3 illustrates the significant changes in stratigraphic thickness of the Prophet Formation, and the Minnes and Bullhead groups. A facies change occurs in the upper part of the Triassic succession in which the Liard, Charlie Lake and Baldonnel formations change laterally into the massive Ludington Formation, and it is unlikely that the pre-deformation geometry was perfectly planar. Results from experiments by Cobbold (1976) using models, suggest that initial perturbations, such as slight variations in layer orientation, can serve as nucleation sites. Consequently, the distribution of inhomogeneities within a layered succession may, in some instances, preempt the influence of layer thickness and viscosity contrast on fold wavelength development.

To test the influence of layer thickness and viscosity contrast, wavelengths should be measured along the length of the northern Foothills subprovince and compared with the stratigraphic succession to determine whether or not there is a systematic change in wavelength in response to changes in the character of the layered succession, and whether this variation agrees with theoretical predictions.

Rocky Mountains structural subprovince

Summary

The Rocky Mountains subprovince is dominated by several, large, thrust-bound anticlines and anticlinoria that comprise an overlapping succession of folded thrust sheets. The form of each folded sheet varies in accordance with the lithology of its strata. Those composed of middle Paleozoic carbonate rocks, like the Bernard Anticline, are broad, relatively simple structures, whereas those formed of less competent clastic facies, like the Laurier Anticlinorium, are internally complicated.

Smaller scale folds exposed on cliff faces are geometrically equivalent to fold forms described within the Foothills subprovince except that they are rarely upright. Even massive Devonian carbonate can deform into tight chevron-style folds that are composites of Class 2 and Class 1B forms. Figure 46 shows a recumbent fold with alternating Class 1B and Class 2 forms and evidence of substantial bedding plane slip. Figures 47 and 48 illustrate the capacity of dolostones to assume a chevron form but not without some disruption within the hinge zone along low angle faults. The more incompetent Kechika Group calcareous shales and siltstones may be intricately folded at this scale, and show evidence of penetrative deformation (Fig. 30). These folds are geometrically interesting but sufficiently small in scale to be relatively unimportant within the context of regional strain.

The remaining discussion is restricted to a description of the large folded thrust sheets.

Robb Anticline

The Robb Anticline (Fig. 49) is a broad fold of Silurian and Devonian dolostone and limestone that plunges at approximately 15 degrees south beneath Besa River shale at the Halfway River. It has a strike length of 12 km, an exposed width of 4 km, and is concentric in profile section. Its form and geometric relationship with structures adjacent to and below it are illustrated in Structure Cross-section 1 (A-A'; in pocket). Inclination of the western limb increases northward from about 15 degrees to 40 degrees. The eastern limb dips approximately 35 degrees, and adjacent to it is a smaller syncline-anticline pair (Fig. 59). There is little surface evidence along the eastern margin of the anticline to support the interpretation that it comprises a large allochthonous sheet, but at its northern limit the underlying Sidenius Thrust is well exposed and indicates minimum lateral displacement of 5 km (Fig. 59). The surface evidence for a 'folded thrust sheet' interpretation appears contradictory and will be analyzed in detail in the discussion of blind thrusts.

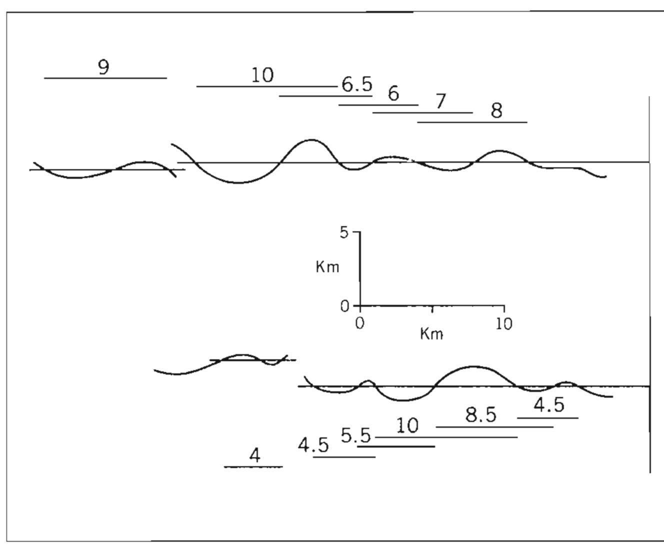


Figure 45. Wavelengths taken from smooth-curve approximations of Foothills folds using structure cross-sections in Figure 73. Horizontal lines denote individual wavelengths measured in kilometres.

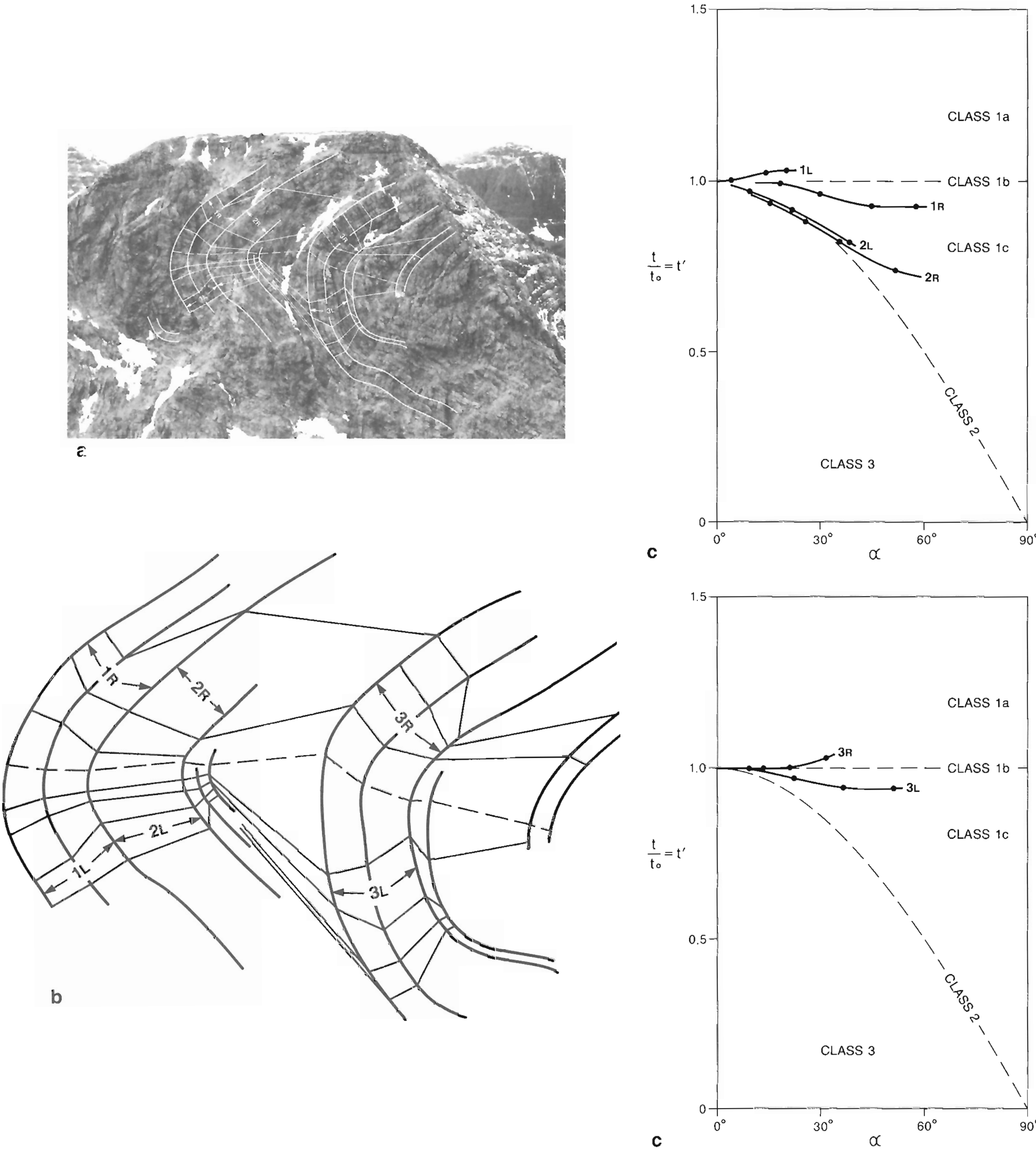


Figure 46. (a) Southwestward overturned, recumbent folds in Devonian carbonate strata, 4 km northeast of Mt. Kenny. (GSC photo no. 943-53). (b) Dip isogons drawn across fold profiles. (c) Fold class plot for each numbered fold limb.

Figure 47. *Faulted chevron folds in Stone Formation carbonate strata. (GSC photo no. 819-123).*

Laurier Anticlinorium

The Laurier Anticlinorium (Fig. 50a) is a large fold complex that extends from Mount Kenny 30 km south to Horn Creek. It lies structurally above the Robb Anticline, and is composed of Ordovician through Middle Devonian off-shelf clastic facies (with the exception of the Middle Ordovician Skoki carbonate) that are lateral facies equivalents of the shelf carbonate succession. Two profiles of the anticlinorium showing eastward overturned, complex folds are presented in Structure Cross-sections 2 and 3 (B-B' and C-C'; in pocket). These folds are well illustrated on the south slope of Calnan Creek, where a large, overturned, near isoclinal anticline is outlined by a Middle Devonian dolomitic quartz sandstone map unit (Fig. 50b). Several of the folds are cut by small thrusts with displacements of one kilometre or less.

Bernard Anticline

The Bernard Anticline (Figs. 50a, 51) emerges from beneath the southern limit of the Laurier Anticlinorium and becomes a broad, box-like fold that can be traced for 60 km to its southern limit at Ingatieff Creek, where it plunges beneath the Burden Thrust. The box-shaped profile is illustrated in Structure Cross-sections 4, 5, 6, and 7 (D-D', E-E', F-F', G-G'; in pocket). The anticline is similar to, but larger than box folds described from the Foothills subprovince. The steeply dipping east limb (Fig. 51b) is cut by a small thrust along its northern half. The Guilbault Thrust (Fig. 51c) truncates the western flank of the Bernard Anticline except near its southern limit, where the fault can be traced into the core of an adjacent syncline. The anticline is also cut, across strike, by the Bernard Fault (see discussion of transverse faults), a high-angle, south dipping(?) reverse fault that repeats a segment of the fold core.

Burden Anticlinorium

The Burden Anticlinorium is a complexly folded and faulted thrust sheet comprising a basal layer of Kechika calcareous shales overlain by Ordovician through Middle Devonian carbonate rocks. Its internal geometry is illustrated in Structure Cross-sections 8 and 9 (H-H', I-I'; in pocket). Unlike the folds described above, there is less structural continuity within the Burden sheet. Fold size and shape is variable, and most folds are cut by thrusts. In character with the other folds, the eastern flank of the anticlinorium consists of a steeply dipping limb, in this case truncated by a splay off the Burden Thrust. Some of the folds, especially within the Dunedin Formation, are nearly isoclinal and strongly overturned toward the east.

Balden Creek Anticline

All of the fold complexes described above occur along the eastern margin of the Rocky Mountains structural subprovince. The Balden Creek Anticline occurs adjacent to

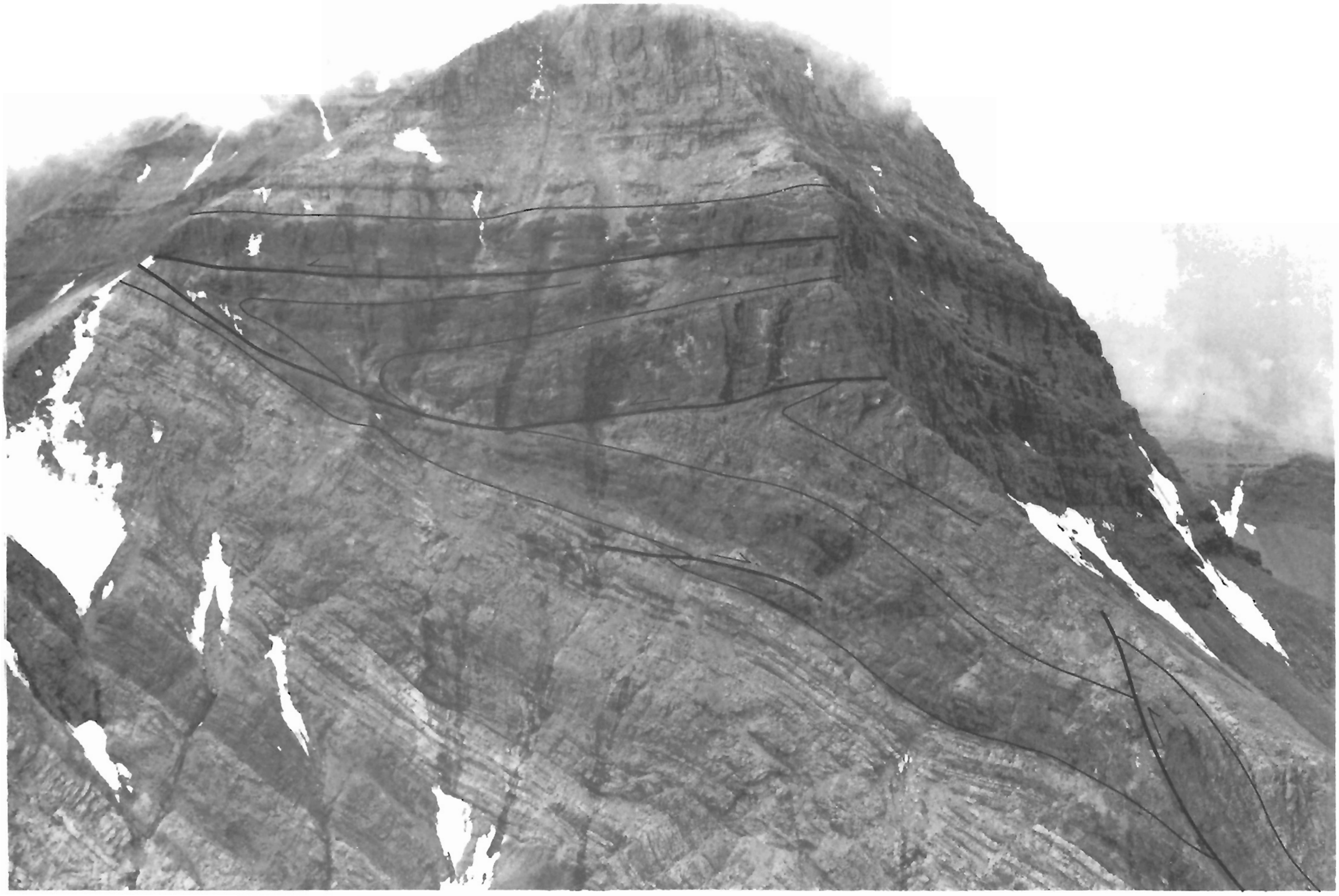


Figure 48. *Southwestward view of faulted recumbent chevron fold in Devonian carbonate strata. (GSC photo no. 943-50).*

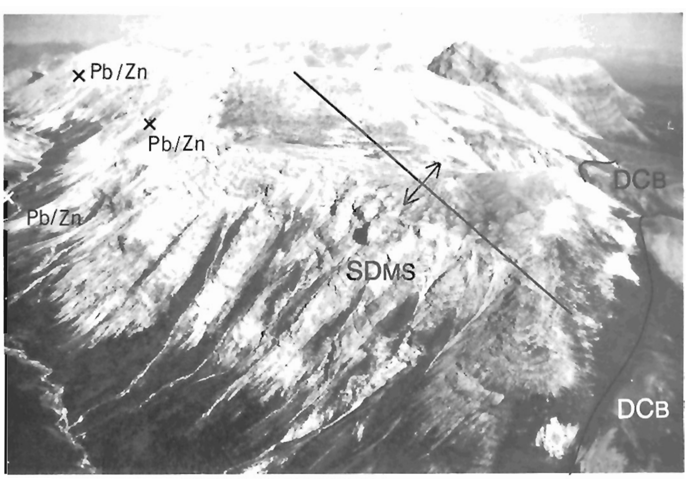
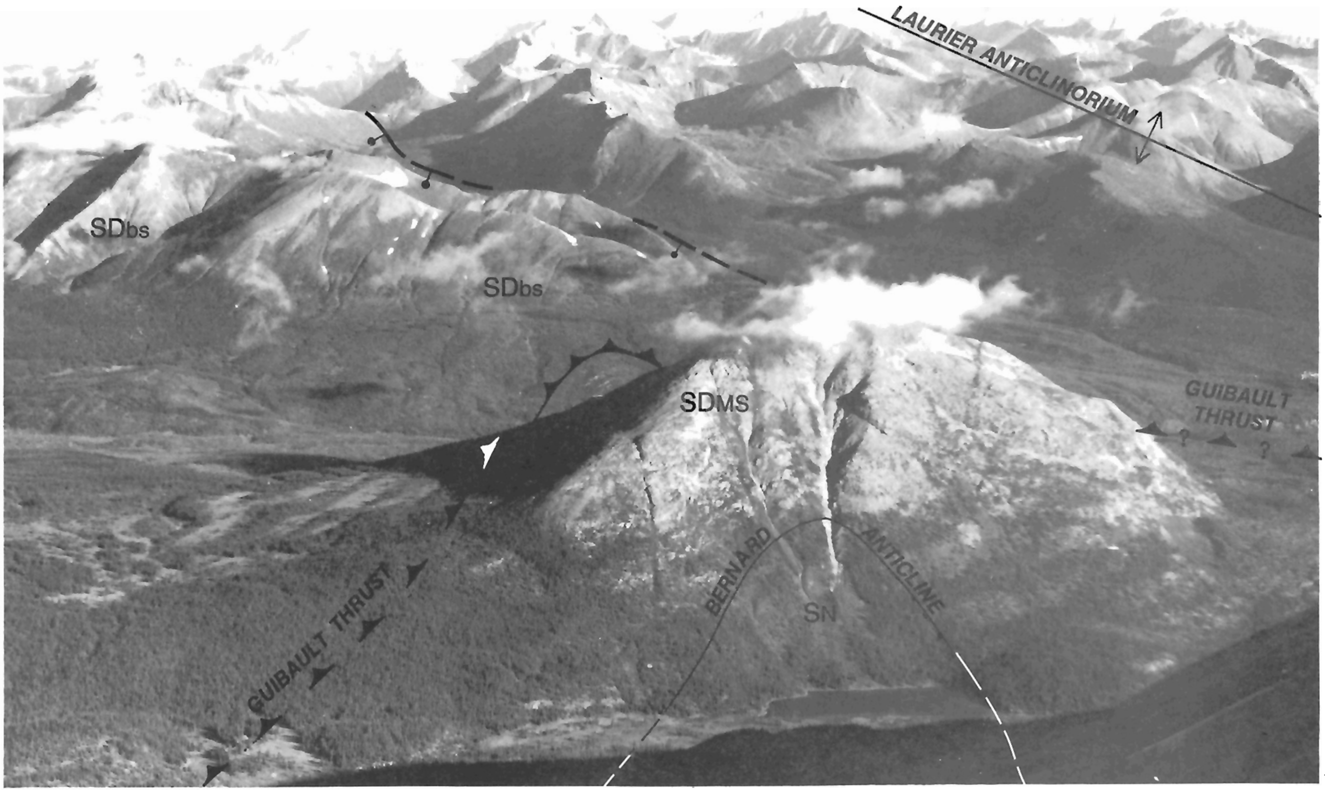


Figure 49. *Northward view of Robb Anticline, located on the north side of the Halfway River valley 10 km east of Robb Lake. Pb/Zn, lead/zinc showing;* SDMS, *Muncho–McConnell/Stone formations, undivided;* DCB , *Besa River Formation. (GSC photo no. not available).*

the western flanks of the Bernard and Laurier structures. It can be traced southward from the headwaters of Balden Creek to an abrupt fault termination along Aley Creek. Structure Cross-section 4 (D-D') illustrates the profile shape of the anticline. Kechika calcareous shales form the fold core and overlying Skoki carbonate forms the limbs (Fig. 51c). The eastern flank is cut by the Guilbault Thrust, and the western flank by a strand of the Burden Thrust. Near its southern termination, the anticline is overturned toward the west. At the headwaters of Guilbault Creek, it is overturned toward the east and maintains this attitude for most of its remaining length.

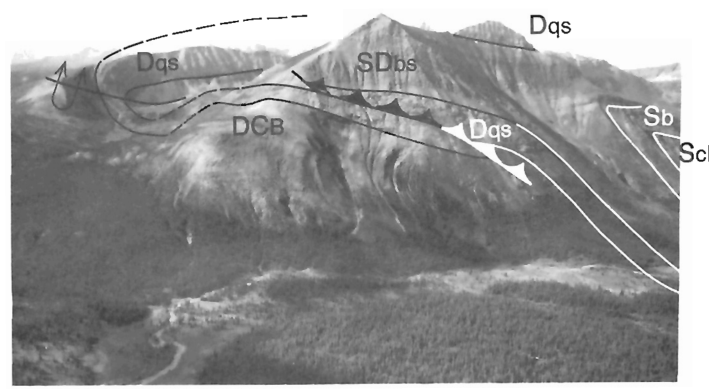
Other folds

Several other large anticlines and synclines are present between the northeasternmost strand of the Burden Thrust and the Herchmer Thrust to the southwest. The folds are outlined, for the most part, by Skoki carbonate, the only competent marker unit. Closed to isoclinal, eastward verging folds are common at all scales of size. Most are cut by thrusts, and few can be traced for any great distance. The lack of markers within the Kechika Group makes it appear deceptively simple at the map scale, but it is penetratively cleaved, and intricately folded in most areas.



a

Figure 50. (a, above) Northward view of the north plunging termination of the Bernard Anticline (centre foreground), and the overriding thrust panel of Silurian and Devonian off-shelf facies (SDbs *). Note the lack of internal organization of the Rocky Mountains relative to the Southern Rockies Front Ranges subprovince shown in Figure 25. (b, right) Southwestward view along the mountain front, of recumbent anticline composed of the brown siltstone unit (* SDbs *) and dolomitic quartz sandstone unit (* Dqs *). (* SN *, Nonda Formation;* SDMS *, Muncho– McConnell/Stone formations undivided;* SDbs *, brown weathering siltstone unit;* Dqs *, dolomitic quartz sandstone unit;* DCB *, Besa River Formation;* Sb *, Breccia unit;* Scl *, Carboniferous limestone unit). (GSC photo nos. 819-98 (a), 657-148 (b)).*



b

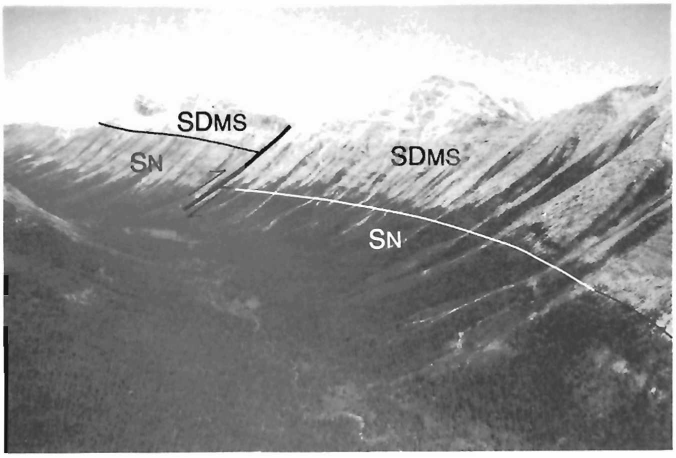
Models of chevron fold development

Experiments using models designed to investigate the mechanics and mechanisms of fold development have led to significant advances in our understanding of the folding process, especially with regard to development of the straight-limbed, narrow-hinged folds typical of the Foothills subprovince. Results suggest that this geometry is both stable and energy efficient, that limb dips of 60 degrees should be expected, and that an abrupt change in amplitude at the margin of a fold complex is reasonable.

One approach centres on the energy requirements of folding. It is reasonable to assume that a buckled layer will adopt the most energy efficient form. Using this premise, Bayly (1974) calculated the amount of energy required for the continued development of a variety of fold shapes at a variety of limb inclinations. The results of this work are summarized in Figure 52. It can be demonstrated that less energy is required for a chevron fold form to continue developing than for concentric folds with large rounded hinges; in addition, limbs become straighter and hinges narrower as limb inclination increases. This model is based on the following assumptions:

1. The volumes of competent and incompetent material are equal
2. The viscosity ratio is 400
3. The profile length of one wave is 40 times the thickness of the competent member.

These assumptions do not apply to the material properties of the northern Foothills, because the viscosity ratio is too high, and the wavelength to thickness ratio is too large. The close correspondence of predicted fold geometry with natural fold geometry however, supports Bayly's conclusion that straight



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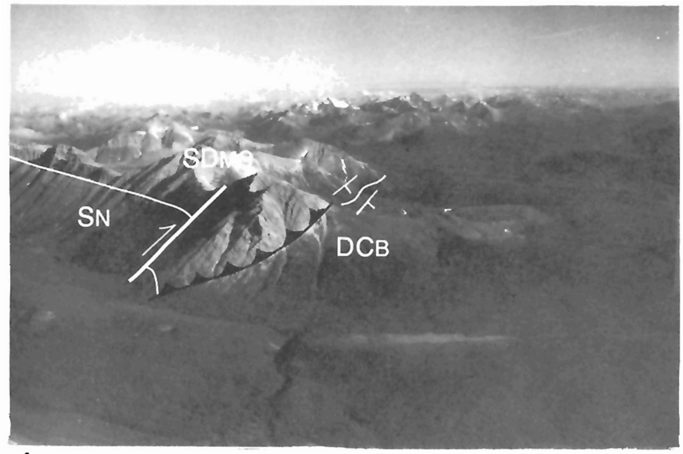

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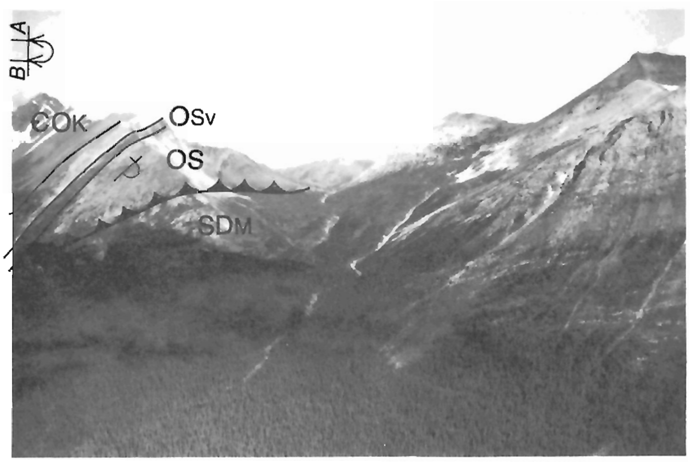


c

Figure 51. *(a) Northwestward view across the axis of the Bernard Anticline (SN , Silurian Nonda Formation; SDMS , Silurian and Devonian Muncho–McConnell and Stone formations). (b) View northward along the eastern margin of the Bernard Anticline showing a minor thrust (right foreground) placing Muncho–McConnell and Stone formations (SDMS) over Besa River Formation (DCB). This minor thrust dies out within 2 km northward (top centre of photo) where the eastern flank of the Bernard Anticline is unfaulted. (c) Northward view along the western flank of the Bernard Anticline: Guilbault Thrust (GT) places overturned Ordovician Skoki Formation (OS) carbonate onto the Muncho–McConnell Formation (SDM). Dark weathering unit at upper left is a volcanic layer (OSv) (approximately 50 m thick). East limb of Balden Anticline (BA) at left. (ЄOK , Kechika Group). (GSC photo nos. 657-193 (a), 819-94 (b), 657-164 (c)).*

limbs and narrow hinges do not require a local change in mechanical behaviour but are simply the most economical shape.

A further outcome of Bayly's work, which has been discussed in detail by Ramsay (1967, 1974), is that once a chevron fold form has been initiated (i.e. has a limb dip of about 5-10 degrees) progressively less energy is required to continue closing the fold up to a limb inclination of about 60 degrees. After this point the energy requirements increase. Ramsay (1974) demonstrates this in his geometric analysis, which shows that the rate of fold development increases quickly once limb dips are greater than 10 degrees (Fig. 53; from Ramsay, 1974). Once initiated, it will continue to develop with relative ease until steep limb inclinations are achieved, at which point it will begin to resist further closure and eventually "lock up". Ramsay's calculations assume that the low viscosity layers within the multilayered succession are free to flow into dilated zones generated within the hinge area and that bedding plane (flexural) slip occurs along the length of the limbs as well as in the hinge area. Viability of this latter assumption has been disputed by Bayly (1976).

Viewed from a perspective of energy requirements for fold development, the existence of a large change in fold amplitude at the eastern margin of the Foothills subprovince is predictable. It requires less energy for a fold to continue developing than for initiation of a new fold adjacent to it. Consequently one should expect the most easterly fold structure to achieve 60 degree limb dips (that is, begin to "lock up") before an adjacent fold begins to develop. Therefore, during the waning stages of deformation, folds already initiated along the fold belt margin will continue to develop because energy requirements are reduced, but no new folds will be initiated because the regional stress field is no longer adequate. The result is an abrupt change in fold amplitude as shown in Structure Cross-sections 2 through 9 (in pocket).

Ramsay's (1974) model of chevron fold development (Fig. 53) is based on a geometric analysis of naturally occurring folds. He suggests that the chevron shape is typical of multilayers with large ductility contrast between layers as long as the thickness of the competent layers does not change significantly – a sudden increase in thickness causes fold limbs to become rounded. The incompetent layers must be free to flow into areas of dilation that develop in the hinge zone between competent layers as they slowly collapse during fold closure. By analyzing the amount of bedding plane slip that must occur during fold development, the area of the zone of dilation that must be filled, and the consequence of changing bed thickness, Ramsay suggested the following generalizations. The chevron style is stable as long as the thickness of the competent layers is relatively constant (this restriction can be relaxed if the ratio of competent layer thickness to limb length ratio is less than 1:10); bulbous hinge forms or thrusts may accommodate the presence of a single very thick layer, whereas very thin layers may be extended or boudinage structures may result. Hinge dilation is a potential feature that is dependant upon the amount of shortening and the proportion of competent layer thickness to limb length. Maximum dilation occurs

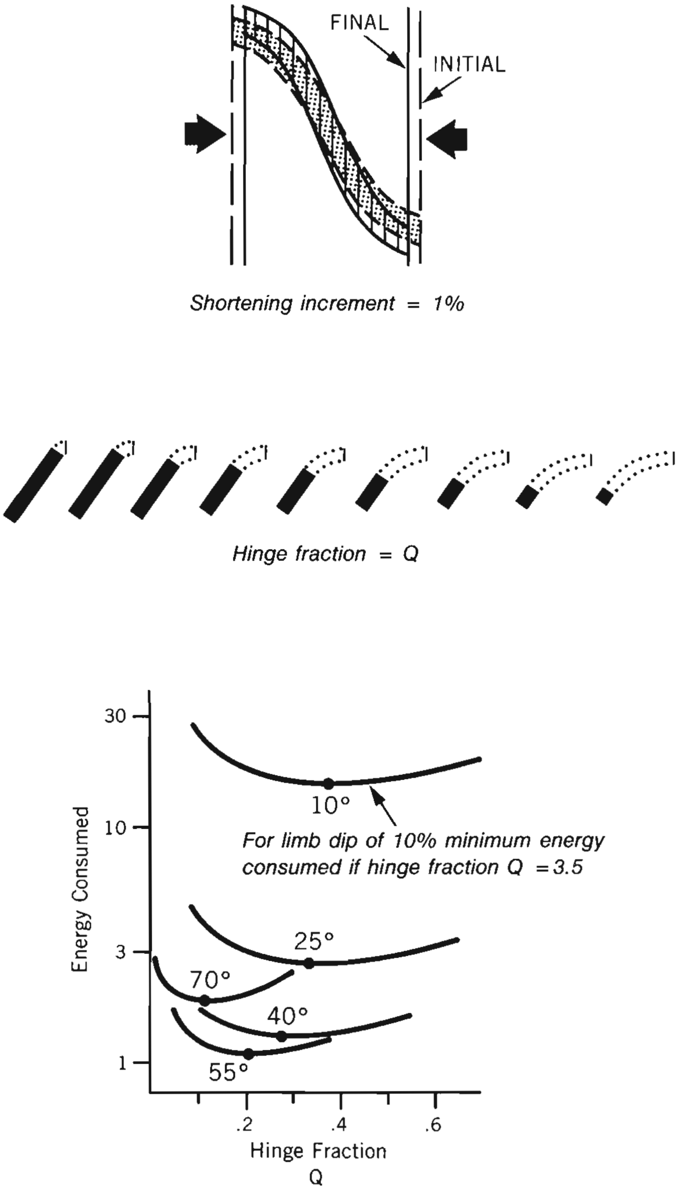


Figure 52. *Summary of results from Bayly (1974) illustrating the relationships during folding of a single layer between energy consumed, fold shape (hinge fraction Q), and limb dip. (a) Energy calculations are made for shortening increments of 1%. (b) Fold shapes used for the analysis are defined in terms of the ratio of hinge length to limb strength, called the hinge fraction Q. (c) Plot of relationship between energy consumed and hinge fraction Q shows that for given limb dip, indicated for each curve, there is a "most efficient" fold shape (identified by the black dot on each curve); as limb dip increases toward 55°, the energy required to fold the layer decreases and the hinge fraction Q decreases. Beyond limb dips of 55°, energy consumed increases but Q continues to decrease. The results of this study are: 1. folding of a single layer becomes progressively easier up to limb dips of around 60°; and 2. as the fold develops, it progressively assumes a chevron shape.*

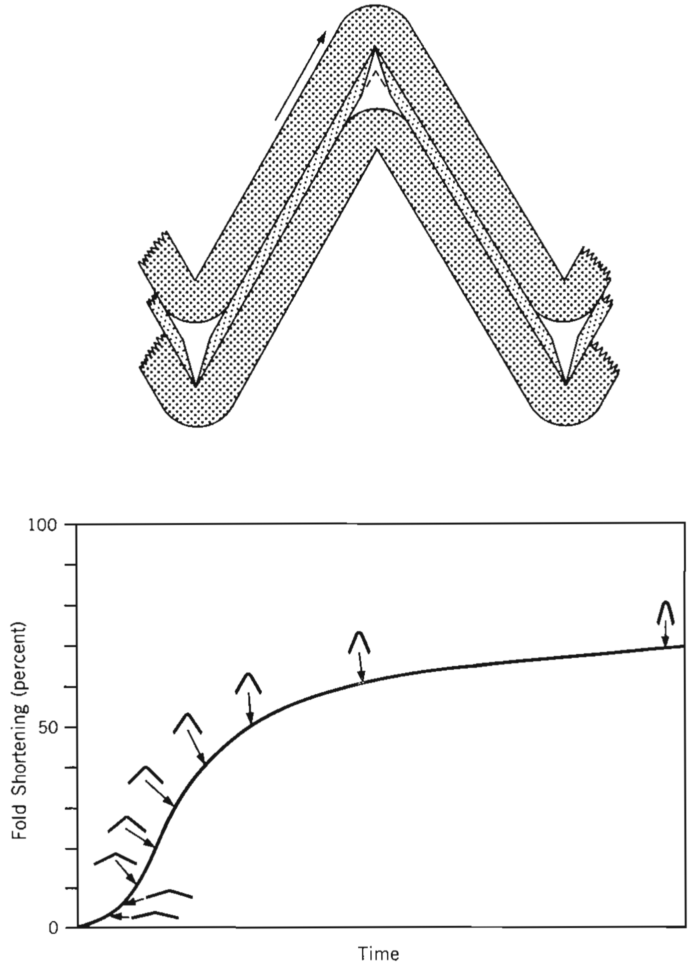


Figure 53. *(a) Geometric model adopted by Ramsay (1974) for an analysis of chevron fold development. Coarsely stippled layers represent competent strata; finely stippled layer represents incompetent strata, which, in the model, are allowed to flow into the hinge area. Resistance to sliding between competent and incompetent layers is assumed to be zero. (b) Results of Ramsay's analysis showing that a constant applied stress will cause increasingly rapid fold formation to limb dips of about 50° to 60°. Decrease in the rate of fold development is equally rapid beyond limb dips of 60° – the fold is said to "lock up".*

during the latest stages of fold development. Once formed, fold development accelerates rapidly until the interlimb angle approaches 60 degrees, at which point the fold begins to resist further closure and "lock up". Small-scale structures within the hinge zone may record a complex history of superimposed extension and contraction structures related to progressive deformation within the hinge.

At the regional scale, the Foothills subprovince does not conform to a regularly interbedded succession of competent and incompetent layers all with the same thickness. Yet the hinge effects, noted by Ramsay, resulting from the presence of an anomalously thick competent layer, are rarely seen. One should expect the Prophet Formation, which is up to 800 m thick, to exhibit a more rounded, bulbous, and faulted hinge form because the thickness to limb length ratio rarely exceeds 1:10. Figures 40 through 42 show little evidence of the features Ramsay describes. A possible explanation, illustrated by Figure 41, is that the Prophet Formation contains a sufficient number of ductile intervals to allow internal compensation within the hinge while at the same time increasing the effective thickness to limb length ratio well beyond 1:10. This logic applies to the upper part of the Triassic succession and to the Minnes and Bullhead groups: each contains numerous, relatively thin, ductile units that accommodate a chevron geometry without excessive distortion of the hinge areas. The middle Paleozoic carbonate succession is much more massive and homogeneous and one might apply Ramsay's conclusion that it is too thick and too resistant to sharp bending to allow a chevron geometry to develop. This may be the case for the Bernard Anticline, which is much wider than any Foothills anticlines, although at a smaller scale (Figs. 46-48) it can be demonstrated that a chevron form (with associated faulting in the hinge zone) can occur.

The energy analyses of Bayly and Ramsay presuppose existence of the chevron form at the time of fold initiation. Model experiments by Cobbold (1975, 1976) and Johnson (1977) demonstrate this may or may not be the case. Cobbold shows that both concentric-like and chevron-like fold forms may coexist within the same experiment, and that one form may change to the other as the experiment proceeds. He also points out that not all folds begin as sinusoidal waves (as predicted by the theory of elastic materials) but may nucleate at a point as a very small flexure that then spreads outward in all directions. This is the manner in which kink bands form and propagate (Patterson and Weiss, 1966) and leads to the question of whether or not some chevron and box folds could have resulted from the interaction of propagating kink bands on a very large scale. Faill (1969) has shown that the geometry of fold complexes within the Valley and Ridge province of the central Appalachians can be explained, geometrically, in terms of conjugate kink bands. A conjugate set of kink bands forms a box-shaped fold that narrows inward into a chevron shaped fold where the kink bands intersect (Fig. 54). Faill (*ibid.*) points out that several intersecting kink bands with different orientations will produce a fold complex with the same geometric properties as folds within the Valley and Ridge province. In other words, he has taken the kind of geometries produced by Patterson and Weiss (1966) at the millimetre and centimetre scale, and applied them to natural folds at the kilometre scale. Acceptance of this model carries with it the implication that folds nucleate at many points within the layered succession as very small, abrupt flexures that grow outward, progressively, and in all directions. Therefore, beds forming the kink band were forced to rotate through the hinges at the kink axes. Cobbold (1976) points out that under these circumstances there should be microscopic and mesoscopic evidence frozen within these beds in the form of vein fillings, numerous joints and fractures, and minor faults, to indicate that they were once severely bent. This does not appear to be the case within Foothills folds. A migrating fold hinge also requires sequential thickening and thinning as the hinge migrates along the beds. This process should lead to fine-scale disruption of original sedimentary features (if the flow is penetrative), as well as evidence of extension and contraction frozen into the rocks along the entire fold limb. Another feature that might be expected is variation of orthogonal thickness of the fold limb (slight pinch and swell resulting from uneven flow of material into and out of the hinge. This kind of supporting evidence is lacking, and suggests that hinges did not migrate through the rock succession to any significant degree. The size and shape of the hinge zone may have changed but the position has not.

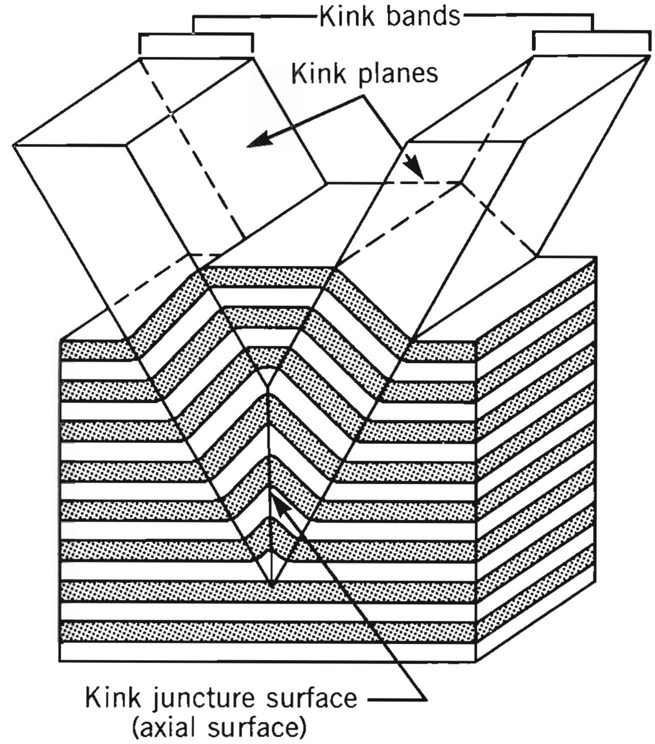


Figure 54. *The interaction of two oppositely inclined kink bands resulting in a change from box-fold to chevron-fold shape. Note that within the chevron part of the fold, shortening parallel to layering decreases progressively downward to zero (from Faill, 1969, Fig. 3).*

Figure 55 is a summary illustration of results from experiments using models by Johnson (1977) in which he employed elastic materials. It shows that folds may begin as sinusoidal waves, undergo differential wavelength changes to become concentric, and then yield within the inner portions of the fold hinge into a chevron form that propagates upward (and downward) along the hinge; the end result is a complex of chevron folds with steep limb dips. This progressive evolution of fold form requires a change in hinge geometry but not hinge location, which was established at the time of sinusoidal wave formation and is controlled by the thickness of the competent layer and the ductility contrast between it and the confining medium.

Scale model experiments by Dennis and Hall (1978) and Blay et al. (1977) have produced results similar in style to natural fold belts such as the Jura, the central Appalachians and the northern Rocky Mountain Foothills. Each model uses gravity as a driving force and each contains a weak basal layer over which the folded succession deforms and moves. The experimentally produced folds had box and chevron forms, formed serially, and were located at random – there was little correspondence between predicted and actual wavelength. Blay et al. (*ibid.*) noted that some folds developed at stress concentration points caused by uneven lubrication of the basal layer.

Most existing theoretical treatments of fold development are based on the initial assumption that a periodic sequence of folds appears instantaneously throughout the entire layered sequence with one small increment of deformation. These assumptions are a consequence of considering natural layered systems to be perfectly elastic;

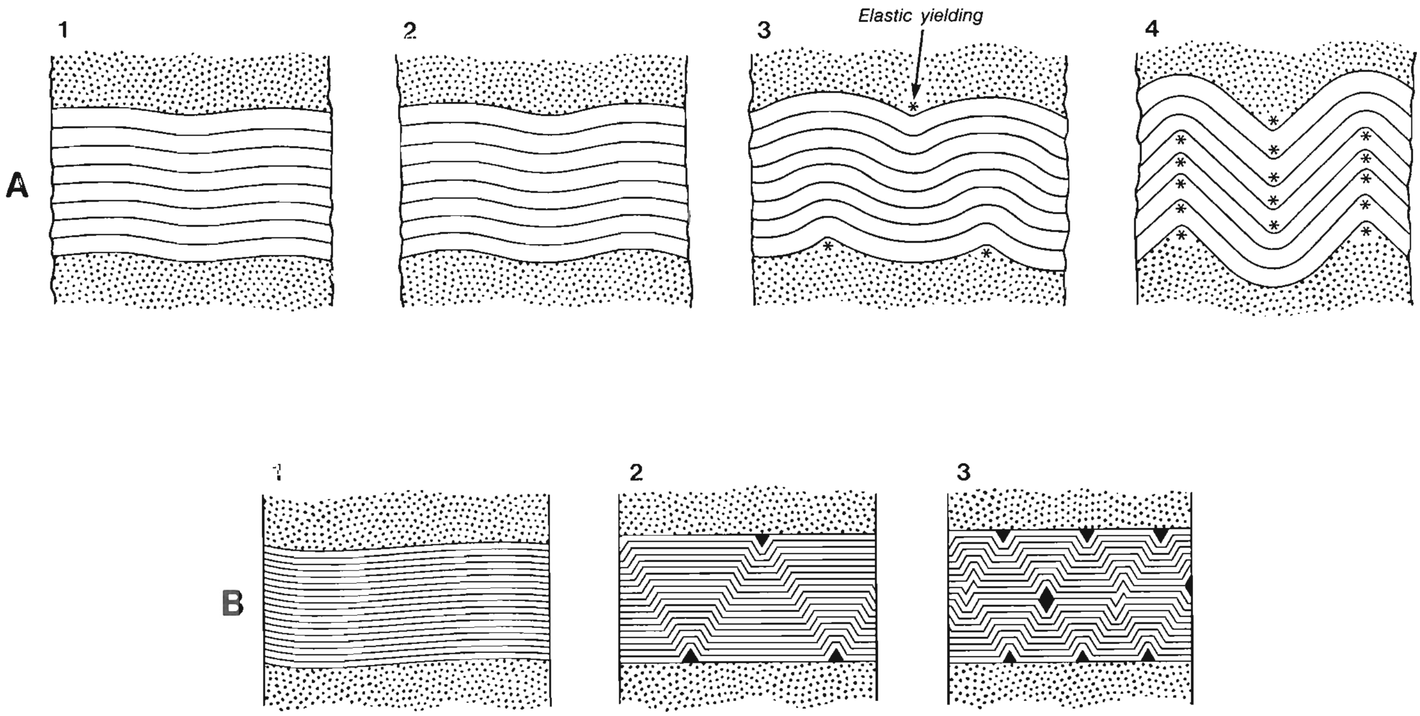


Figure 55. Sequence of fold forms developed when an elastic multilayer is compressed. Sequence A begins with a sinusoidal form, and evolves through a concentric form into chevron forms; sequence B begins with a broad sinusoidal form, evolves to kink forms which eventually intersect to form local chevron fold forms (from Johnson, 1977, Fig. 5).

but as Cobbold (1976) points out, natural systems contain inhomogeneities and undergo viscous as well as elastic deformation. The propagation rate of folds may be relatively slow compared with the rate of amplification. The experimental results of Blay et al. (1977) support this view. The results show that the stress front imposed through lateral compression of the model propagates through the model material at a rate that varies with the amount of lubricant on the basal layer and the type of inhomogeneities within the system. Once the stress front reaches the limits of the model, folds begin to form serially (not simultaneously) as chevron and box-shaped anticlines separated by broad, undeflected synclines. Studies of natural folds by Cobbold (1975) support the interpretation that a serial pattern of fold development may occur. His results indicate that folds may be grouped into complexes in which fold amplitude decays outward from the central fold. He experimented using a single, embedded wax layer in a less viscous wax matrix, with a small perturbation built into the otherwise planar layering. He found that an anticline formed on the site of the perturbation, and that subsidiary folds with progressively smaller amplitudes formed adjacent to the initial fold. Each fold complex formed in this way had an internal periodicity, but the fold complexes developed serially with no periodicity evident between complexes. Applied to natural fold systems, these experimental results suggest the following: 1. folds may form serially (west to east in the Foothills) rather than simultaneously; 2. the site of initiation of a fold complex may be caused by some original inhomogeneity within the layering (e.g. a facies change); and 3. individual folds within a fold complex may be periodic, but the complexes may not.

Foothills folds may be viewed as a succession of stress "release valves" (Fig. 56). Build-up of regional compressive stresses was accompanied by lateral propagation of the stress front eastward. This continued until a stratigraphic inhomogeneity such as a thickness or facies change was encountered, providing a focus of stress concentration. The rocks eventually yielded and a fold was initiated. Because the stress required to keep the fold developing is far less than that required to initiate it, the fold immediately and rapidly reduced the regional horizontal stress field. Presumably this process can occur over a very short period of time (in geological terms). Once limb dips of 50 degrees are attained, compressive stresses begin to build because more energy is required to keep the fold closing. Eventually, the fold locks, losing its ability to absorb stress and causing a resumption of lateral stress propagation until another inhomogeneity is encountered and a new fold forms.

Faults

Introduction

Thrusts are essential structures of the northern Rocky Mountains structural subprovince. Unlike the southern Rockies, where thrust-bound homoclines form discrete, rectilinear mountain ranges, thrusts in the north have less regular patterns. The overriding sheets are more deformed and difficult to recognize. The en echelon pattern of thrusts in the south is less obvious in the north. At the latitude of the Peace River, several thrusts can be recognized within a dominantly carbonate terrane, but at the latitude of Chowade River, one may traverse the Rockies subprovince and not encounter a single thrust of consequence. Stratigraphic facies changes account for differences in style within the map area as well as between regions, but facies changes do not explain the very conspicuous lack of thrusts along the mountain front. It is difficult to reconcile the Bernard Anticline as an allochthonous sheet transported several kilometres eastward when a thrust cannot be mapped along its eastern margin; one might equally assume the Bernard

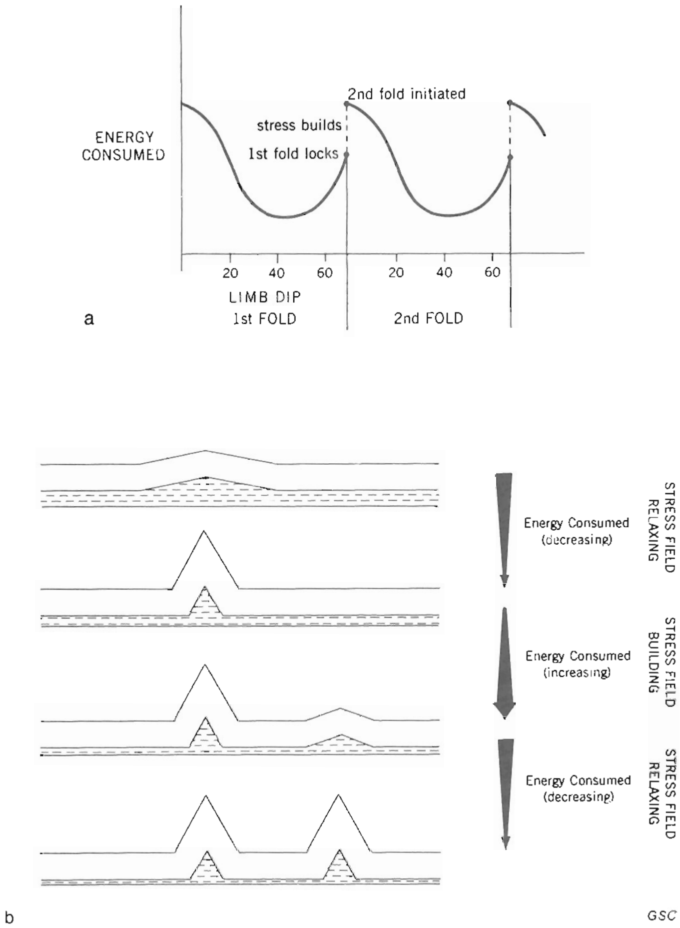


Figure 56. *(a) Diagrammatic plot of expected energy consumption (in relative units) versus limb dip for Foothills folds within the Halfway River map area. Energy required to initiate a fold is large, but drops significantly as the fold develops, until limb dips of about 60° are reached; the fold then begins to "lock up", fold development eventually stops and stress levels increase until large enough to initiate a new fold, beginning anew the cycle of stress release followed by stress build-up. (b) Diagrammatic representation of chevron fold development and its effect upon orogenic stress levels. Orogenic stresses relax during folding, and build between periods of folding.*

Anticline is a drape fold over a basement uplift like those proposed by Stearns (1975) for the (Laramide) Rockies of Colorado and Wyoming.

Most of the ensuing discussion will focus on field evidence that supports the interpretation of the northern Rocky Mountains as a "thin-skinned" tectonic regime. A key structural element on which this is based is the "blind" thrust: a non-outcropping thrust, the displacement of which is absorbed by disharmonic structure within the thrust sheet (Thompson, 1979, 1981).

The combination of folds with thrusts in some parts of the map area have produced unusual geometries, such as 90 degree fault truncations of bedding, and younger strata thrust onto older strata. Steep dipping contraction and extension faults are present in some areas; most are small except for the Bernard and Aley Creek faults, which are oriented transverse to the structural grain and displace the southward plunging portions of large anticlines.

The differences between the northern and southern Rocky Mountains are more apparent than real. Using the blind thrust model, it will be demonstrated that the principles of thin-skinned tectonics used in the preparation of structure cross-section interpretations in the south, are equally applicable in the north.

Blind thrusts applied to the interpretation of the mountain front

Blind thrusts

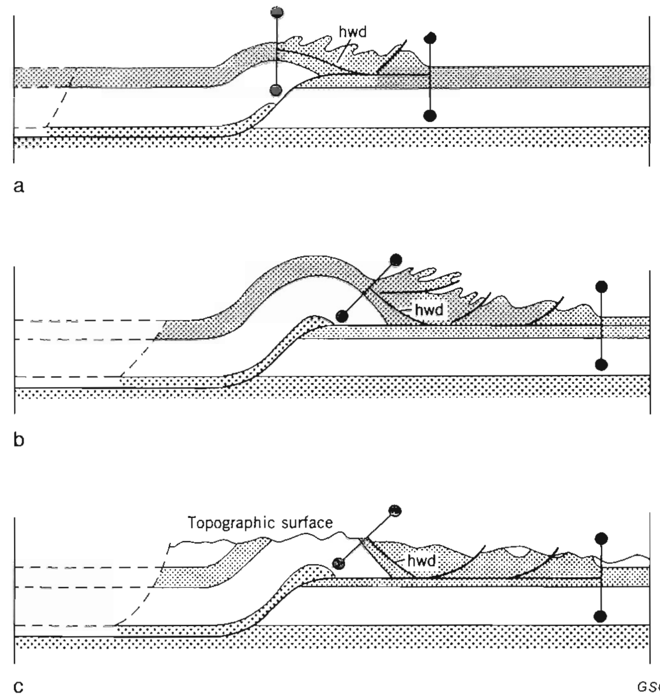
Blind thrusts are an important structural element of the northern Rocky Mountains subprovince. They account for some of the important differences in structural style between the northern and southern Rocky Mountains. Unlike thrusts that splay upward from a basal detachment zone to intersect the topographic surface, blind thrusts merge and flatten at some point in their upward trajectory into an "upper" detachment zone(s); consequently they are rarely observed at the surface.

Geometric considerations require that the displacement on a blind thrust be transferred into the overlying (hanging wall) beds through formation of a complex of disharmonic structures. Otherwise, balanced shortening from one stratigraphic level to another would not be possible (Dahlstrom, 1969). This is illustrated in Figure 57, a simplified diagrammatic representation of the blind thrust model.

As displacement on the blind thrust is initiated, a second detachment, within the overriding sheet, forms and permits incompetent strata to behave independently and disharmonically with respect to the underlying rigid (hanging wall) carbonates (Fig. 57a). If the amount of shortening due to disharmonic folding keeps pace with the amount of displacement on the blind thrust, it will not extend laterally beyond the limits of the fold complex; if this balance is not maintained, then a splay from the blind thrust will develop, and break to surface as a kind of "release valve" that accommodates the shortening imbalance. As displacement continues, the lateral limits of disharmonic folding are extended to keep pace with fault displacement (Fig. 57b). A field example of detached disharmonic folds (Fig. 58) shows Carboniferous carbonates have become folded over "undeformed" Carboniferous carbonates. According to the interpretation of Figure 57, the folds are absorbing displacement on a blind thrust (the detachment zone) that displaces (shortens) Devonian carbonates farther to the west (right).

By superimposing a hypothetical topographic surface onto Figure 57c it becomes apparent how misleading the surface geology may be. There is no structural break between the large surface anticline on the left and the fold complex on the right, and no way of determining that the large anticline is actually an allochthonous thrust sheet that is folded over a fault ramp or step at depth. Based on the surface evidence, one might argue in favour of vertical (basement) uplift as the agent responsible for the large anticline and that the fold complex adjacent to it is a consequence of gravitational sliding off the uplifted carbonate terrane.

Figure 57. Diagrammatic representation of the blind thrust model. The layers with patterns represent mechanically incompetent strata, separated by a rigid carbonate unit with no pattern. (a) illustrates the onset of displacement across the thrust accompanied by development of a hanging wall detachment (hwd) that allows the incompetent strata within the hanging wall plate to deform disharmonically and absorb displacement on the underlying thrust. The thrust ceases to exist at the point where shortening due to folding in the hanging wall equals displacement on the thrust – hence the pin on the right side of the section. In (b), continued thrust fault displacement increases the width of disharmonically deformed hanging wall succession. (c) illustrates the difficulty in deciphering the nature of the detachment of the mountain front anticline using surface exposures: the major thrust remains 'blind', and much of the shortening within the disharmonically deformed, incompetent unit may be difficult to assess unless good stratigraphic markers are present (from Thompson, 1981).



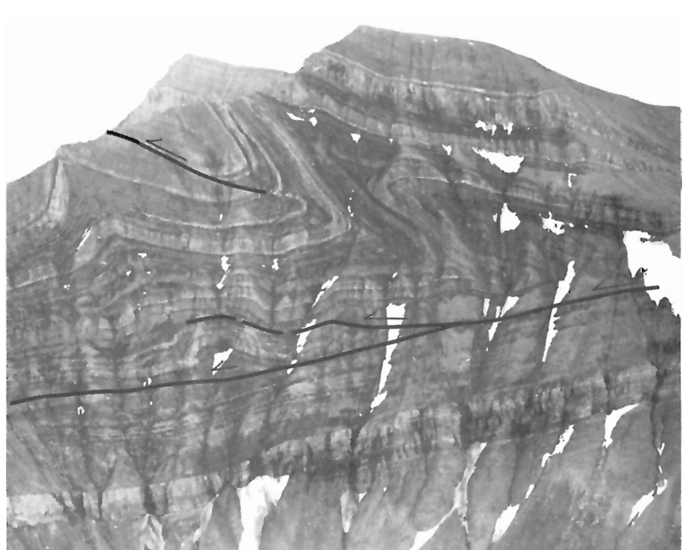
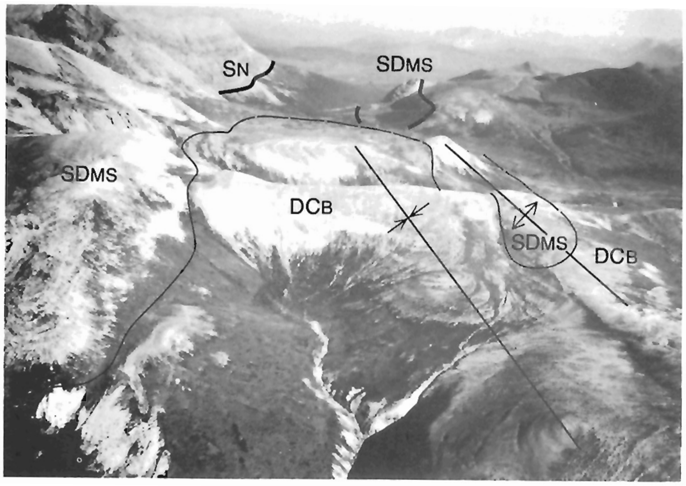


Figure 58. Southward view (3 km northeast of Mount Burden) showing disharmonic folds within argillaceous limestone and shale (Mississippian Prophet Formation) that are detached from shale and limestone within the Besa River Formation (from Thompson, 1979, 1981). (GSC photo no. 819-140).



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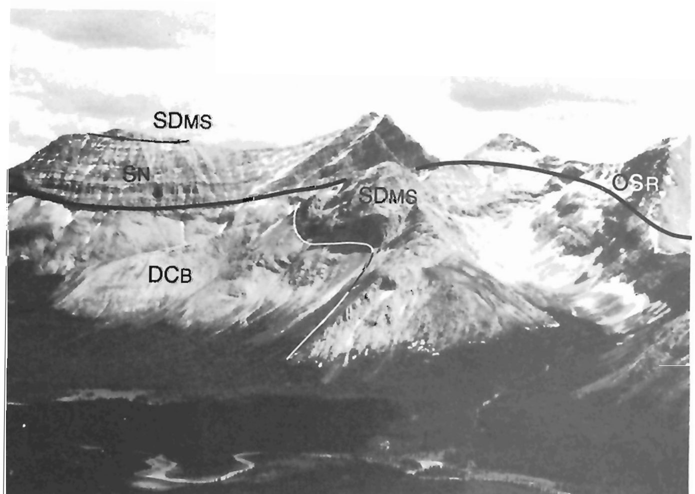


b

Figure 59. Up-plunge (a) and down-plunge (b) views of the Robb Anticline illustrating the allochthonous nature of the anticline despite the absence of a mountain front thrust. (a) Northward view showing the eastern, unfaulted limb of the Robb Anticline. The contact between Besa River shale (DCB) and Muncho–McConnell and Stone formations (SDMS) is stratigraphic. Thrusts in background appear minor, and large-scale transport is not obvious. (b) Southward view showing an erosional cut-away of the anticline on the south slope of Sidenius Creek (SN , Nonda Formation; OSDR , Road River strata). The anticline was produced by large-scale (minimum 5 km, west to east) transport across the Sidenius Thrust. The Thrust cannot be mapped southward along the eastern flank of Robb Anticline beyond the limits shown in (a). (GSC photo nos. 1172-1 (a); 1172-3 (b)).

Field evidence favouring the blind thrust model (Figs. 59, 60) is present in those areas where the plunging part of a mountain-front anticline is cut across by a deeply eroded valley that exposes lower structural levels. This occurs at the northern extreme of the Robb Anticline and at Mount Burden. Details of the structural geometry of these regions follows.

Relationship of the Robb Anticline to the Sidenius Thrust

Immediately north of the Halfway River, the Mountain front is defined by the moderately dipping east limb of the Robb Anticline, composed of Silurian and Devonian dolostone. A shallow syncline containing Besa River shale lies adjacent to the Anticline (Fig. 59a), and no structural break occurs at the mountain front. East of the syncline, the uppermost part of the carbonate succession reappears along the hinge of a narrow, low profile anticline before disappearing beneath a broad expanse of Besa River shale. Farther east, one encounters only Besa River shale or younger units forming Foothills folds. Approaching the Robb Anticline from the south leaves one with the impression that it represents a broad upwarp involving a relatively minor amount of horizontal shortening, and that the transition from Foothills to Rocky Mountains folds is gradual. However, if one approaches the same locality from the north, an entirely different perspective and set of geometric relationships are encountered. In Figure 59b, a view toward the south taken from Sidenius Creek, the Robb Anticline occurs in the centre of the photograph as the broad fold projecting across the skyline. On the left, the thick, cliff-forming carbonate rocks of the Silurian Nonda Formation are separated from underlying Devonian and Mississippian Besa River shale by the Sidenius Thrust, which can be traced 3 km laterally. Map relations demonstrate a minimum horizontal displacement of the Robb Anticline of 10 km from west to east, yet 5 km southward along strike, there is no exposed evidence of the allochthonous nature of this large fold. In Figure 59a, the Sidenius Thrust is shown in the background at right centre, where it places Devonian carbonates onto Besa River shale; it can be traced to the south for approximately 1 km, at which point it is "lost" within Besa River shales. What happens to the thrust? To answer this question one must travel south along the mountain front to Mount Burden where an explanation is preserved in spectacular cliff exposures.

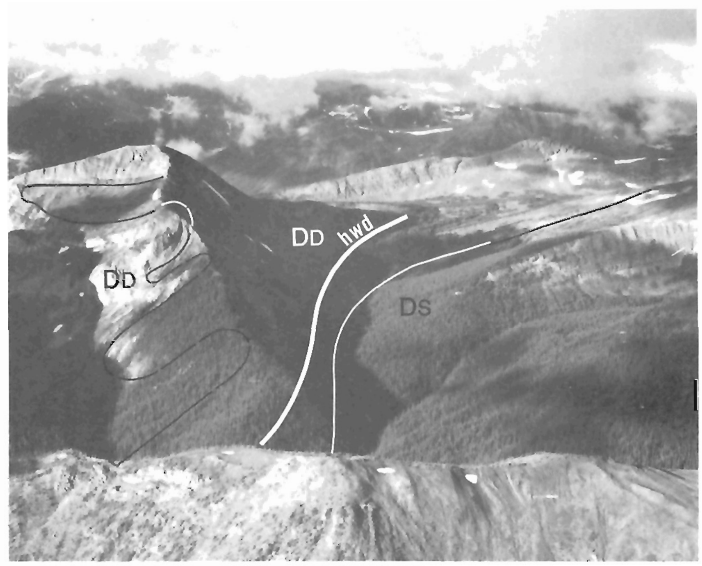
The Burden Thrust

The Burden Thrust is the largest thrust in the region, with a minimum horizontal displacement of 10 km. Over most of its exposed cross-section length, Ordovician Kechika Formation calcareous shales and siltstone are thrust onto Besa River shale, but at its eastern limit, the thrust cuts through the entire middle Paleozoic carbonate succession of the hanging wall plate (Fig. 60). It splays into a system of thrusts (Burden Thrust system) toward the northwest.

The problem with the Burden Thrust, as with the Sidenius Thrust, is that it disappears at its eastern end where it enters the Besa River shale in the hanging wall plate. The solution to this geometric dilemma is found within the hanging wall plate. Figure 60b is taken from the same camera location as Figure 60a, but is a view southward along the mountain front. In the centre middle ground, a line is drawn along the contact between the rigid carbonate succession of the Stone Formation and the less competent, carbonaceous limestones of the Dunedin Formation and younger units; this boundary defines the east dipping limb of the frontal anticline that we know is cut by the Burden Thrust in Figure 60a. In the left part of Figure 60b, a line is drawn that outlines a succession of stacked, recumbent folds in the Dunedin Formation. This succession has been shortened several times its original bed length, whereas the underlying, more competent, carbonate succession dips beneath it as an essentially undeformed plate. For this reason, a detachment must exist (labeled "hwd") above the

a

Figure 60. *(a) Southwestward view of Mount Burden (right foreground) showing allochthonous Burden Anticlinorium. The Burden Thrust (heavy dark line) can be traced from west to east (right to left) for more than 10 km, but cannot be traced eastward beyond the hanging wall cut-off of the Ordovician through Devonian carbonate succession (* ЄOK *, Kechika Group;* OS*, Skoki Formation;* SDM*, Muncho–McConnell Formation;* DS*, Stone Formation;* DD*, Dunedin Formation;* DCB*, Besa River Formation) or southward along the eastern margin of the anticlinorium. Near the eastern margin of the Burden Anticlinorium, there is a detachment (hwd = hanging wall detachment) above which less competent carbonate and shale of the Dunedin Formation deforms disharmonically with respect to the underlying Stone and older carbonates. (b) A closer view of geometric relations around the detachment as shown in the area indicated by the heavy arrow in (a). Detailed view toward the south of disharmonic, recumbent folds above the hanging wall detachment. The shortening accommodated within folds of Dunedin carbonate is interpreted as partially balancing displacement on the Burden Thrust. (GSC photo nos. 943-9, 943-10 (a); 819-80 (b)).*



b

Stone Formation that permits overlying, less competent units to fold independently of their substratum. This hanging wall detachment permits compensation of movement on the Burden Thrust (at depth) by disharmonic folding and preferential shortening of units within the hanging wall.

Interpretation of Bernard Anticline and Laurier Anticlinorium

Having established that the Robb Anticline is, in reality, a large, allochthonous, folded thrust sheet, and that the Sidenius and Burden thrusts are blind, one is led to the conclusion that other mountain front folds, such as the Bernard Anticline and the Laurier Anticlinorium, are large, allochthonous sheets even though erosion is not deep enough to expose the thrusts on which they were transported.

The Laurier Anticlinorium (see Structure Cross-sections 2 and 3; in pocket) forms a salient of Silurian and Devonian clastic facies that overlaps platform carbonate equivalents at its northern and southern terminations. At the northern end, it is clear that the fold complex is thrust over the Robb Anticline where the latter plunges steeply southward into the subsurface. The McCusker Thrust can be traced southeastward along the western flank of the Robb Anticline to near the Halfway River where it splays into two thrusts that cross the Halfway and then swing eastward along the south slope of the valley. This fault and its splays cut up stratigraphically in both hanging wall and footwall successions until both are "buried" in Besa River shale beneath the Laurier Anticlinorium. Cross-section profiles show that the Laurier Anticlinorium consists of a succession of smaller, eastward verging folds that cascade downward from west to east. At its southern limit, near Laurier Lake, it is difficult to separate stratigraphic from structural relationships. Lower and Middle Devonian shale and siltstone sit with apparent conformity on the northward-plunging nose of the Bernard Anticline. If one opts for a stratigraphic explanation, this contact represents an abrupt, northward facies transition from carbonates to clastics. An alternative explanation is that the apparently conformable contact is a continuation of the Guilbault Thrust from the south, which wraps around the north plunging Bernard Anticline and then cuts upsection into the Besa River shale in a manner similar to the McCusker Thrust. This latter interpretation is favoured for the following reason. About 100 m above the point where the Bernard Anticline plunges out, corals of Early Devonian age were collected; if the contact between carbonate and clastic facies is stratigraphic, then the carbonates must be older than those Early Devonian corals. However, if one follows the carbonate/clastics contact to the southeast along the eastern flank of the Bernard Anticline, Besa River shale overlies Middle Devonian Dunedin carbonate with apparent conformity. To the northeast, along the margin of Laurier Anticlinorium, Besa River shale overlies Middle Devonian clastic rocks. Therefore, for a stratigraphic interpretation to apply, a very abrupt change in facies would have to occur within a strike length of about 1 km or less, around the northeastern margin of Bernard Anticline. This occurrence is not supported by field observations. Instead, the contact is interpreted as an extension of the Guilbault Thrust, and the Laurier Anticlinorium is thought to be a large, complexly folded and faulted allochthonous sheet that structurally overlaps laterally equivalent carbonate rocks at either end (see Structure Cross-sections 2 and 3).

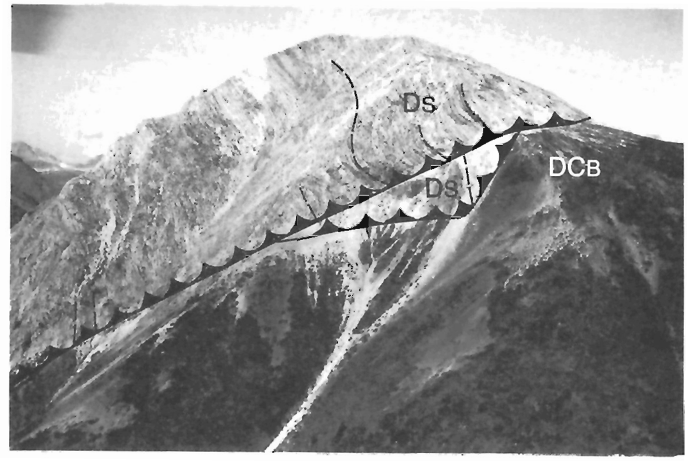
The Bernard Anticline is interpreted in a similar manner – as a large, allochthonous thrust sheet soled by a blind thrust that is structurally lower than the Burden and Guilbault thrusts (see Structure Cross-sections 4-7).

Thrusts that break the surface

West of the mountain front anticlines, two large thrust systems strike northwestward toward Mesilinka map area (94 C). The Burden Thrust can be mapped west from Mount Burden to Wicked Lake; there it changes direction to the northwest and splays into three thrusts. The one on the west enters the Ospika River valley near the headwaters of Aley Creek and carries Lower Ordovician, Kechika calcareous shales over Middle Ordovician Skoki carbonate. The two more eastward faults splay into four thrusts at Gavreau Creek. Between there and Aley Creek, the map relationships are complex because some splays cut older folds (see Structure Cross-section 4), but north of Aley Creek, the splays rejoin to form a single thrust that continues up Balden Creek and over to the Ospika River valley near the mouth of Denkman Creek. The thrusts cannot be mapped across the Ospika River valley because of poor exposure. However, on strike to the northwest, Gabrielse (1975) has mapped the Trident and Ospika thrusts, which can be traced northwestward into Ware map area (Taylor, 1983). It seems reasonable to conclude that the Ospika and Trident thrusts are northwestward extensions of the Burden Thrust complex.

The Herchmer Thrust forms a second large thrust complex in Halfway River map area. It was mapped and named by Gabrielse (1975) who has traced it from the Rocky Mountain Trench southeastward across Ware and Fort Grahame map areas. This is a very large thrust that places Upper Proterozoic Misinchinka over Upper Cambrian and Lower Ordovician Kechika Group shale. The Herchmer Thrust enters Halfway River map area near the mouth of Gavreau Creek, trends southeastward to the Peace River and enters Pine Pass map area (McMechan, 1987) along the east slope of Colin Creek. Fault splays that imbricate Cambrian carbonate and quartzite map units occur beneath the main fault. These splays rejoin the main thrust near Peace River.

Other thrusts exposed within the Rocky Mountains subprovince are small in comparison to the Burden and Herchmer structures. Several thrusts cut the Burden



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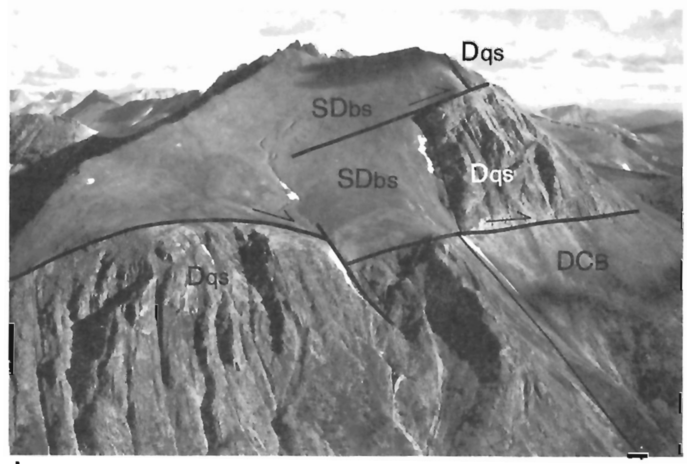

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Figure 61. *(a) Northward view along the eastern flank of the Bernard Anticline showing Stone Formation thrust onto Besa River Formation; and thrust cuts perpendicular to bedding of the Stone Formation (* DS *, Stone Formation,* DCB *, Besa River Formation). (b) Northward view along the eastern flank of the Laurier Anticlinorium. Massive beds of the dolomitic quartz sandstone unit stand vertically and have been cut at a right angle by subhorizontal thrusts (* SDbs *, brown weathering siltstone unit,* Dqs *, dolomitic quartz sandstone unit,* DCB *, Besa River Formation). (GSC photo nos. 657-191 (a), 819-11 (b)).*

Anticlinorium where it forms a large salient of middle Paleozoic carbonate rocks near the Peace River, but displacements are small.

In summary, the Rocky Mountains subprovince is cut by six major thrust complexes: the Sidenius, McCusker, Guilbault, Burden and Herchmer thrusts, and an unobserved blind thrust beneath the Burden Anticline. Displacement across each of these structures is greater than 5 km, and may be much greater for the Burden and Herchmer faults.

Thrust-to-bedding truncation angles

Thrust fault trajectories have a "staircase" geometry comprising long bedding plane glide zones linked by short inclined ramps. In most restored cross-sections, the ramps are interpreted as cutting bedding at 30 degrees (approximately). This truncation angle is usually inferred because hanging wall and footwall truncations are rarely preserved

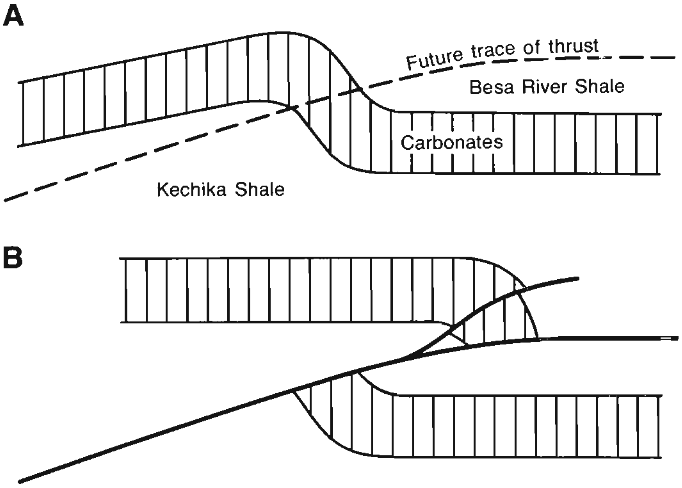


Figure 62. Diagrammatic representation of propagating, monoclinal warp cut by thrust. Fault-to-bedding truncation angle is large. Note that fault does not step upward through the carbonate succession but remains nearly planar and subhorizontal because it cuts stratigraphically up by cutting laterally through inclined strata.

for direct observation. In the Halfway River area, several hanging wall thrust truncations can be studied, and in each case the thrust cuts at or near a right angle to bedding. This is true for the Sidenius Thrust (Fig. 59b) at its southern termination, the Burden Thrust where it ramps through the middle Paleozoic carbonate succession at the mountain front (Fig. 60a) and for smaller thrust splays, like the two shown in Figure 61. In restored section, these fault truncations plot as vertical steps that appear mechanically unreasonable. The angular relations are easily explained if the thrust is interpreted to cut through a fold or fold complex. Most thrusts can be mapped along strike into folds (e.g. the Guilbault Thrust) that project beyond the thrust tip. As the thrust tip propagates laterally, it presumably is preceded by a propagating fold or fold complex. Therefore, the thrust grows laterally and longitudinally into inclined beds, and may truncate them at a right angle. Figure 62 is a diagrammatic illustration of how this may occur when a simple monoclinal flexure is cut by a thrust, and the field relations illustrated in Figure 61 support this illustration. In Figure 61a, the thrust has cut across the east dipping limb of the Bernard Anticline; it is interpreted as a relatively late splay with less than 1 km of displacement. To the south, along the trend of the fold limb (Fig. 27), kink bands have formed prior to formation of new subsiding thrusts. In Figure 61b, horizontal thrusts cut vertically dipping beds within an overturned anticline along the eastern limit of the Laurier Anticlinorium. This interpretation may be applied to the much larger Burden Thrust, which maintains a very flat trajectory through the middle Paleozoic carbonate succession, suggesting that the carbonates were steeply inclined prior to projection of the fault.

The intensity of footwall and hanging wall deformation adjacent to some major thrusts, which themselves are not severely deformed, suggests that significant deformation may occur prior to rupture along a discrete thrust plane. The hanging wall plate of the Guilbault Thrust contains isoclinally folded, Middle Ordovician Skoki carbonate (Fig. 63). The footwall of the Sidenius Thrust contains tight, overturned folds of Middle Devonian carbonate rocks (Fig. 64). The entire Laurier Anticlinorium is a complex of closed to isoclinal, overturned folds.

The idea of a thrust propagating its way up through an essentially undeformed rock succession is an end member interpretation of a more general situation, in which thrusts and folds develop contemporaneously. Folds may be a result of thrust displacement of the hanging wall over an inclined ramp, folds may precede the propagating tip of a thrust, and thrusts and thrust splays may cut through a well developed fold complex.

Faults transverse to the tectonic strike

Two large faults cut across the tectonic strike. The Bernard Fault is a steep, south dipping, reverse fault that repeats the southern part of the Bernard Anticline. The Aley Creek Fault is a less well documented, steeply dipping fault that forms the southern termination of the Balden Creek Anticline.

The Bernard Fault can be traced eastward from Wicked Lake, where it is truncated by the Burden Thrust, along the headwaters of Nabesche River to the east flank of Bernard Anticline. It is interpreted as a pre-fold and -thrust structure for the following reasons: 1. it cuts up and down stratigraphically in both hanging wall and footwall successions; 2. older rocks are exposed south of the fault, 3. it is truncated by the Burden Thrust, and 4. it has a near vertical dip where exposed (Fig. 65). Irish (1970) interpreted the Bernard Fault as a thrust continuous with the Guilbault Thrust – a reasonable interpretation at the scale of his work. However, more detailed mapping shows that the Guilbault Thrust terminates in a fold complex several kilometres to the north of the Bernard Fault and that the stratigraphy on the north side of the Bernard Fault is continuous to Wicked Lake.

The Aley Creek Fault is not exposed but was mapped on the basis of the abrupt southward termination of the Balden Anticline at Aley Creek. It is also interpreted as a relatively early structure, because it is truncated on the east by the Guilbault Thrust and on the west by a splay off the Burden Thrust.

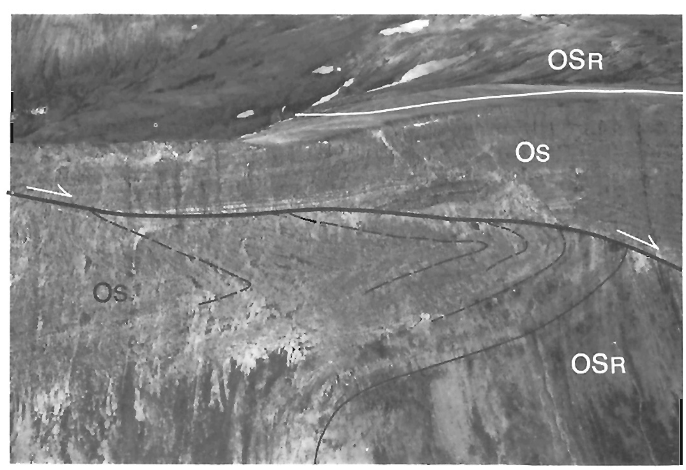


Figure 63. Westward view of Skoki Formation 5 km southeast of Mt. Kenny showing thrust cutting across upper limb of recumbent chevron fold (OSDR, Road River strata, undivided; OS , Skoki Formation). (GSC photo no. 819-23).

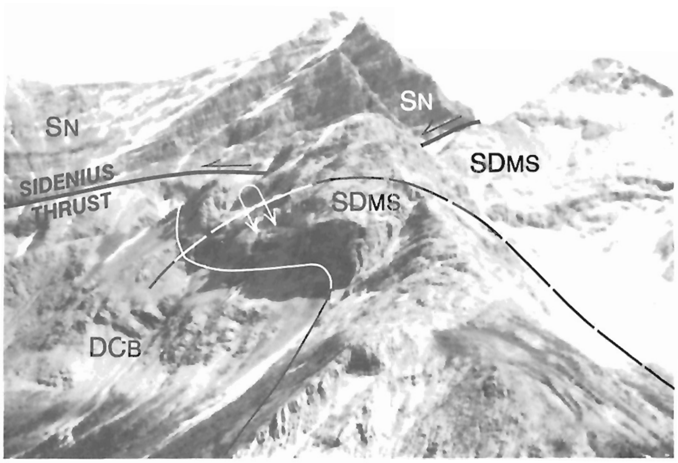


Figure 64. *Southward view showing fold of Stone and Dunedin formations within footwall of Sidenius Thrust (northwestern border of map area; see Fig. 59). (SN , Nonda Formation; SDMS , Muncho–McConnell/Stone formations, undivided; DCB , Besa River Formation). (GSC photo no. 1387-7).*

Steep dipping extension and contraction faults

Steep dipping extension and contraction faults are common within the carbonate successions. Within the Bernard Anticline, numerous steep dipping faults are present (Fig. 66). They rarely have strike lengths in excess of 3 km and offsets of more than 500 m. Some are parallel to the structural grain while others cut across it – there is no definite pattern. North striking contraction faults may become shallow, west dipping thrusts along strike. On the north slope of Lapierre Creek, a small contraction fault can be traced northward to the mountain front where it is a subhorizontal thrust (compare Figure 67a, b, c). This type of event also occurs within the Burden Thrust sheet. A vertical fault with only a few hundred metres of west-side-up displacement can be mapped from the northeast flank of Mount Burden southward to the mountain front where it too becomes a "flat" thrust. The mechanical significance of this along-strike change in fault attitude is not understood.

Principles of thin-skinned detachment tectonics applied to structure section preparation

Introduction

A critical step in the development of a regional tectonic model and synthesis is the preparation of structure cross-sections. The objective of this section is to combine the basic elements of surface structural geometry established for the Halfway River area into a sequence of integrated two dimensional interpretations. Surface geometric data provide the constraints used to project and portray structures in the subsurface.

The approach adopted below is a "thin-skinned detachment" model (Rogers, 1949), in which the key structural elements are rootless folds and gentle- to steep-dipping thrusts that flatten downdip into a master detachment. This model has been applied successfully to the foreland regions of many alpine-type belts, and perhaps with greatest precision to the southern part of the Canadian Rocky Mountains (Dahlstrom, 1969) where exploration for oil and gas has provided a rigorous test of the principles of detachment tectonics.

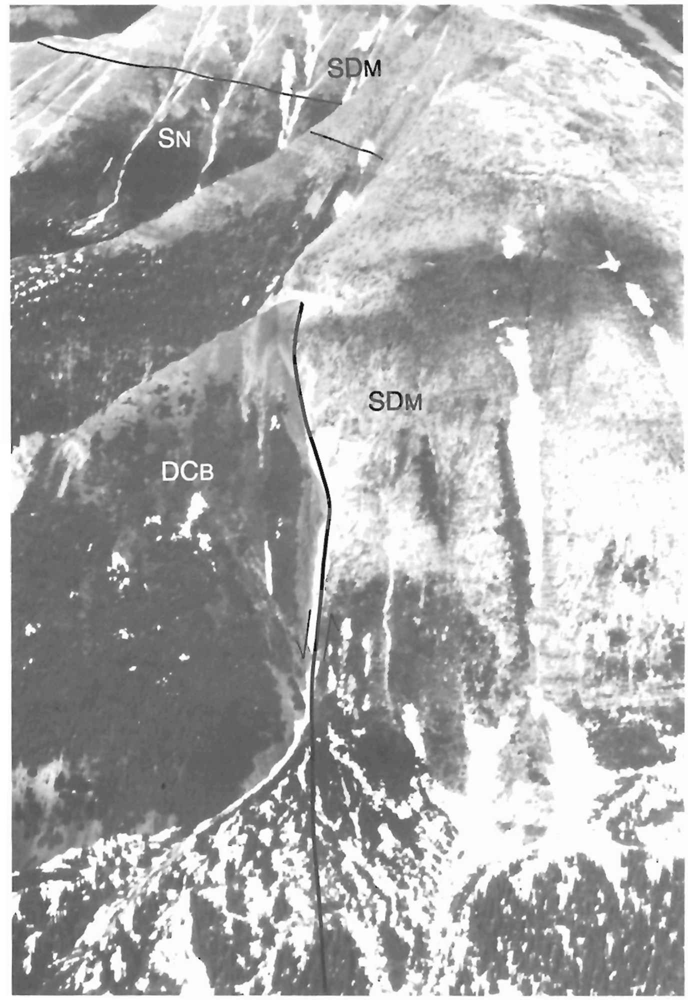


Figure 65. *Eastward view of Bernard Fault, south side (right) up. Besa River Formation (DCB) adjacent to Nonda Formation (SN) in foreground. In background, thin, dark marker at top of photo is contact between Muncho–McConnell Formation (SDM) and Nonda Formation. (GSC photo no. 2112-2).*

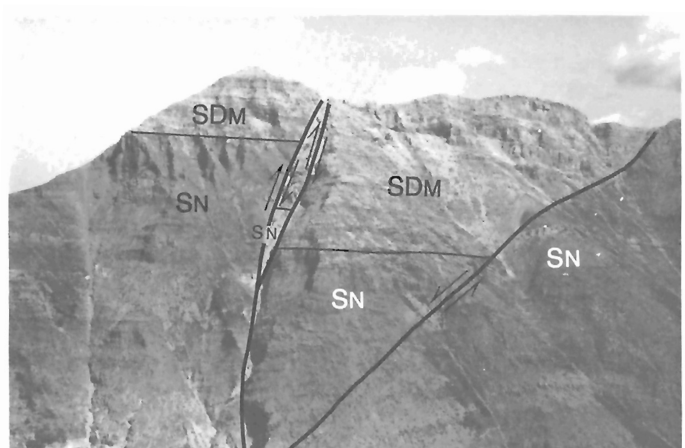
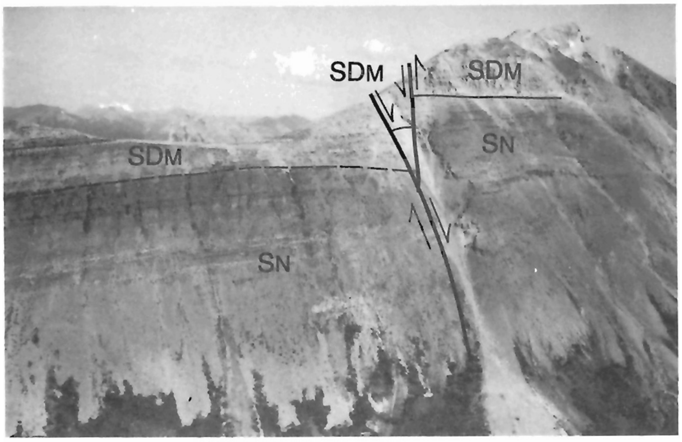


Figure 66. *Northward view along axis of Bernard Anticline. Dark grey Nonda Formation (SN) overlain by lighter grey Muncho–McConnell Formation (SDM). Extension fault starting at ridge crest on right dips to west and is cut (below field of view) by steep, west dipping contraction fault that displaces west side up. (GSC photo no. 657-260).*



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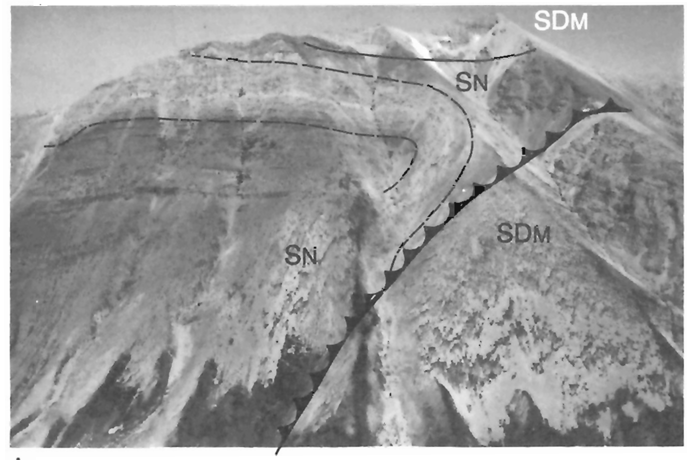

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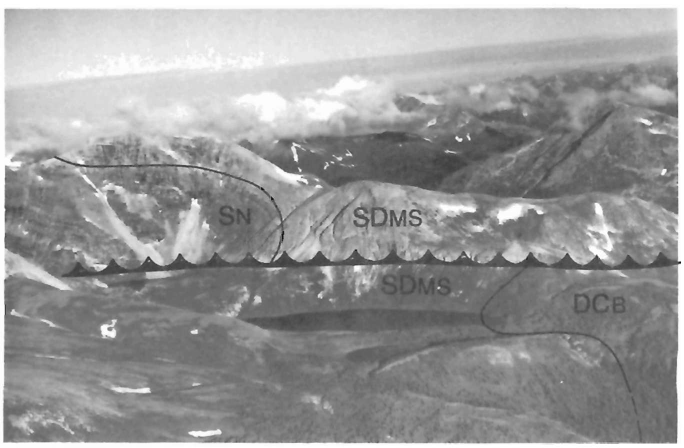

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Figure 67. Sequence of photos showing along-strike change in character of thrust fault located on northeast flank of Bernard Anticline (sequence taken from south to north and viewed northward). (a) Vertical fault with east side up; dark grey Nonda Formation (SN) overlain by light grey Muncho–McConnell Formation (SDM). (b) Steep, west dipping, contraction fault that cuts up through lower limb of anticline in hanging wall. (c) Subhorizontal, west dipping thrust at mountain front placing Muncho–McConnell/Stone Formation (SDMS) onto Besa River Formation (DCB). (GSC photo nos. 657-188 (a), 657-187 (b), 819-154 (c)).

Model constraints

The cross-section interpretations (Structure Cross-sections 1-9) that accompany the surface geological maps are constrained within the geometric framework already discussed and summarized below:

1. The fold style found in the Foothills – planar, steep-dipping limbs and narrow hinge zones – affects all units down to and including the Mississippian Prophet Formation.

2. Foothills folds are detached at the level of the Besa River shale, which is a detachment zone of regional extent.

3. Blind thrusts are responsible for the overlap of thick sheets of lower and middle Paleozoic rocks, and the development of large, mountain front anticlines.

4. Displacement on blind thrusts is linked mechanically to shortening accommodated by Foothills structures.

5. The Kechika Group is a major detachment zone within the Rocky Mountains subprovince.

6. Thrusts may cut through previously folded units, and cut-off angles may approach 90 degrees.

Preparation of a balanced cross-section

A balanced cross-section is one in which area is conserved, that is, cross-section area before deformation equals cross-section area after deformation. This idea assumes that there is no net gain or loss of material during deformation, only physical rearrangement. A balanced section is, at the very least, geometrically possible. If drawn with adequate attention to local geological constraints it should also be geologically reasonable. An unbalanced section is geometrically impossible and therefore geologically impossible regardless of the care used in its preparation. Not all sections can be rigorously balanced, but most sections can be constrained by balancing methods.

In unmetamorphosed foreland terranes, where most rock units undergo little or no penetrative flow, bed thickness is conserved during deformation, reducing the balancing procedure to a one dimensional problem of ensuring that bed length is conserved. Two methods are commonly used to maintain bed length: 1. the 'conservation of fault slip' method, which requires that fault displacement (measured along the fault) of every formation boundary cut by the fault is constant; and 2. the 'sinuous bed method', which requires that bed length is constant for each formation when measured across faults and around folds. Ideally, either approach can be used, and they should have the same net result, but in practice this is not always the case, as

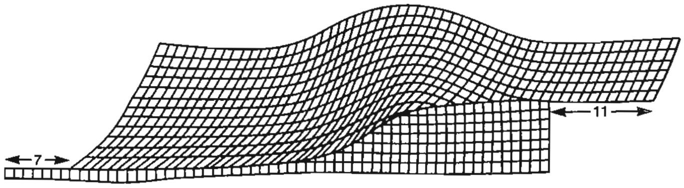


Figure 68. Illustration of how fault displacement can change along fault trace (in cross-section) because of simple shear strain within the displaced sheet as it moves over the fault ramp. Rectangular grid was stamped onto stack of paper that was then cut by a simulated thrust fault. Leading edge of the fault moved forward 11 units, whereas the trailing edge moved only 7 units (from Elliott, 1976).

illustrated in Figure 68. A third and potentially more powerful method of cross-section balancing is the 'area method' which was hinted at by Bucher (1955), applied without explanation by Gwinn (1970), and explained in detail by Elliott (1977) and Hossack (1979). For areas in which thick shale successions are present, or where thrust faults and folds are difficult to map due to a lack of exposure, the sinuous bed method is inaccurate for two reasons: 1. bed thickness may have changed because of penetrative flow, and 2. the method is very sensitive to details of the surface structure. The area method is not affected by thickness changes and is relatively insensitive to surface details, because it involves comparison of the cross-section area in the deformed state with the area of the same cross-section in the undeformed state. The difference in areas is a measure of the additional material required to produce the deformed cross-section, and this quantity, when recast in terms of original orthogonal thickness, has a length equal to the total shortening within the deformed cross-section. Figure 69 is a hypothetical cross-section l_u units wide, in which the total area A_d is calculated between a deformed marker unit and the basal décollement. Orthogonal thickness of the deformed succession is t_o units (obtained from surface stratigraphic sections and well intersections). Prior to deformation, a section l_u units wide would have had an area

$$A_u = l_u \times t_o$$

The area between the dashed line at t_o units and the marker unit is the excess area A_x produced because of tectonic thickening. This excess area can be thought of as the extra rock pushed into the cross-section; before deformation it also had an orthogonal thickness of 't_o' units, and an undeformed length l_x. Because

$$l_x = \frac{A_x}{t_o}$$

the original length l_o of the deformed cross-section is

$$l_o = l_x + l_u$$

$$= \frac{A_x}{t_o} + l_u$$

The usefulness of this method is illustrated in Figure 69b, in which a very generalized surface is used to

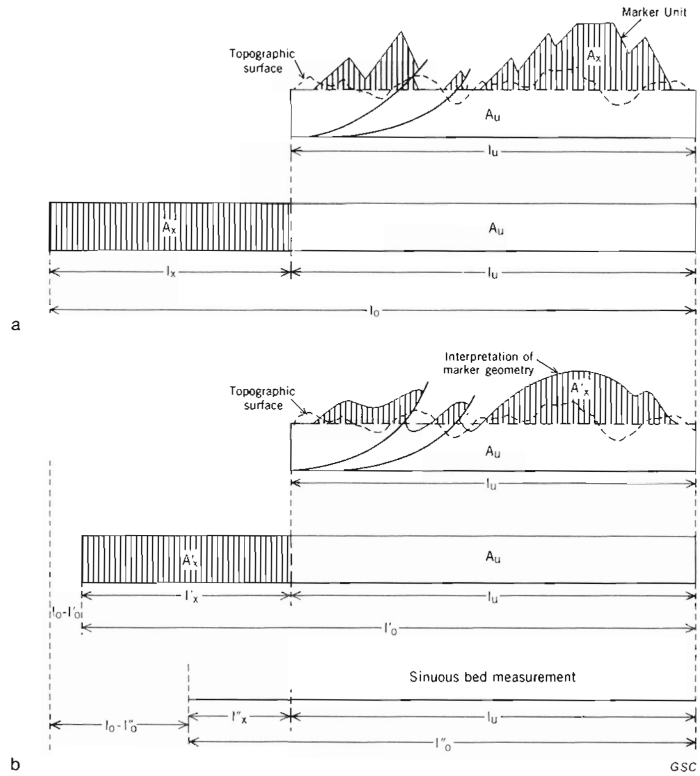


*Figure 69. Illustration of use of area method in cross-section balancing. (a) Geometry of marker unit assumed known (**A_u**, original area of strata **l_u** units long and with an undeformed thickness of **t_o** units; **A_x**, excess area of strata required to fill space between marker unit and original top of undeformed succession; **l_x** is length required to make area equal to **A_x**; **l_o**, total length of section in undeformed state. (b) Geometry of marker unit shown as a smoothed (interpreted) version of (a). Excess bed length within (b) is calculated using sinuous bed method and area method and the results compared. Sinuous bed method suggests that there has been very little shortening (e.g. only **l_x''** units), whereas the area method gives a shortening value (**l'_x**) much closer to the real value (**l_x**).*

approximate the form of the marker unit – an interpretation one might be forced to make if surface exposure was very poor. Using the sinuous bed method, undeformed length l_o', determined by measuring along the generalized surface, is 31 units. Using the area method, l_o' is 38 units, a much closer estimate of the actual value of 40 units calculated in Figure 69a. A section drawn on the basis of 38 units of shortening will have a subsurface interpretation that is more reasonable and it will also force the geologist drawing it to reconsider his generalized surface interpretation.

Both the sinuous bed and area methods were used in the preparation of structure sections across the Halfway River map area. Where surface structure is well exposed within competent units the sinuous bed method was used. Across portions where surface control was poor but the level of a subsurface datum could be calculated, the area method was used to calculate the amount of tectonic thickening and the section drawn to reflect that value. Annotations on each cross-section indicate which portions were balanced using the area method.

Theory put to practice

Construction of the Foothills segment

Surface geology of the Alberta plateau region adjacent to the Foothills subprovince is poorly exposed. This is compensated for, however, by numerous gas exploration wells that provide good control on the depth and thickness of the Cretaceous and the upper portion of the Triassic strata. Few of the wells penetrate to the Carboniferous Prophet Formation (=Debolt Formation of subsurface terminology). Wells were projected along strike into the planes of Structure Cross-sections 4, 6 and 9 and the well intersections joined. Some low-amplitude folds are present, separated by broad, undeflected areas. The origin of these folds is not entirely clear; they may relate only to deformation farther west in the Foothills subprovince or they may reflect the influence of high angle 'basement' faults, original sedimentary thickening and thinning, differential compaction, or a combination of all of these.

The first large deflection of bedding occurs at the Foothills front, where the east limb of a large box or chevron fold rises abruptly (Fig. 27). In most sections, the stratigraphy is well enough exposed so that the details of the fold or fold complex can be drawn with confidence. The limbs of box folds are shown to intersect at depth but the kink band interpretation of Faill (1969) was not followed. Instead, the resulting chevron fold is interpreted to maintain its amplitude downward, so that the amount of shortening across the structure is the same at all levels down to a detachment zone. Preparation of the surface part of the section across the eastern Foothills is straightforward. The question that arises is: at what level are the folds detached?

It was stated previously that the level of detachment of Foothills folds is the Besa River shale. This conclusion was arrived at in the following manner. The broad, flat-bottomed synclines can be used as a crude measure of whether or not there has been significant deflection of the layering above (or below) the regional westward dip beneath the undeformed plains region. If units within the base of the syncline are essentially "ongrade", one may measure downward, using orthogonal thicknesses, to any desired stratigraphic depth within the limits of known stratigraphic thicknesses. Most of the large Foothills synclines within the eastern and central part of the subprovince plot at or near the regional slope projected into the line of section from the east (Fig. 70), demonstrating that the synclines have not been raised by tectonic thickening at deeper levels, or lowered because of downbuckling. Determining the level of structural involvement beneath the anticlines is more difficult because the chevron fold model, adopted here, can, in theory, extend to any depth. However, two gas exploration wells indicate that Devonian carbonate units below the Besa River shale are not involved. The HB Headstone Creek well (see location on Structure Cross-section 2) was drilled into the core of a large anticline at the eastern margin of the Foothills subprovince. Drilling was terminated at a depth of 2600 m below sea level without intersecting Devonian carbonate. If the Devonian carbonate succession had been involved in the folding, it would have been intersected above the -2000 m level. The Schooler Creek well is located in the central part of the Foothills on the eastern flank of a large, thrust faulted anticlinorium (see geological map and Structure Cross-section 4). This well was drilled to -3500 m depth, and did not encounter the Devonian carbonate succession[1]. Given the 40 per cent shortening calculated across this structure, and assuming the base of the Devonian to be at 3700 m, one can only suggest that the Besa River shale is substantially thickened above an essentially uninvolved Devonian carbonate succession.

On the basis of this slim evidence, a regional detachment is postulated at the level of the Besa River shale. However, it is a reasonable assumption on other, more subjective grounds: 1. The Besa River shale has the mechanical properties of a weak basal layer across which detachment would be expected; 2. the Middle Devonian carbonate succession is thicker and more competent than overlying units, making it a mechanical impediment to the downward development of chevron folds (Ramsay 1974, Cobbold 1975); and 3. if the carbonates were thrust up into the cores of anticlines, drilling suggests that there is very little room for large thrust overlaps. Undoubtedly the Devonian carbonates have some structural relief, but it is considered minor in comparison with that of overlying units.

On each of the structure cross-sections, the method used to balance shortening down to the Mississippian Prophet Formation is indicated. In areas where surface geological control is poor, the area method was used to calculate a minimum value of shortening required to complete the section down to the top of the Devonian carbonates (Fig. 71).

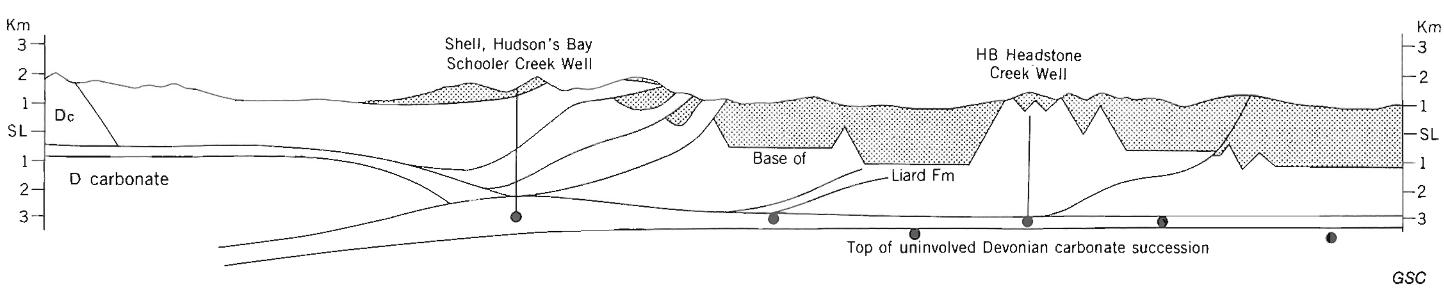


Figure 70. *Comparison of the regional dip of the Middle Devonian surface projected into the plane of cross-section from the Plains region on the east, with the depth to the top of the Middle Devonian estimated by measuring stratigraphically down beneath flat-bottomed synclines (dots) and using available drillhole data (wells labeled).*

[1]The Schooler Creek well has since been redrilled to a depth of 4500 m subsea. It appears that Devonian carbonates are at least that deep and that the assumption of an undeformed Devonian carbonate "basement" at between 3500 and 4000 m is a conservative estimate.

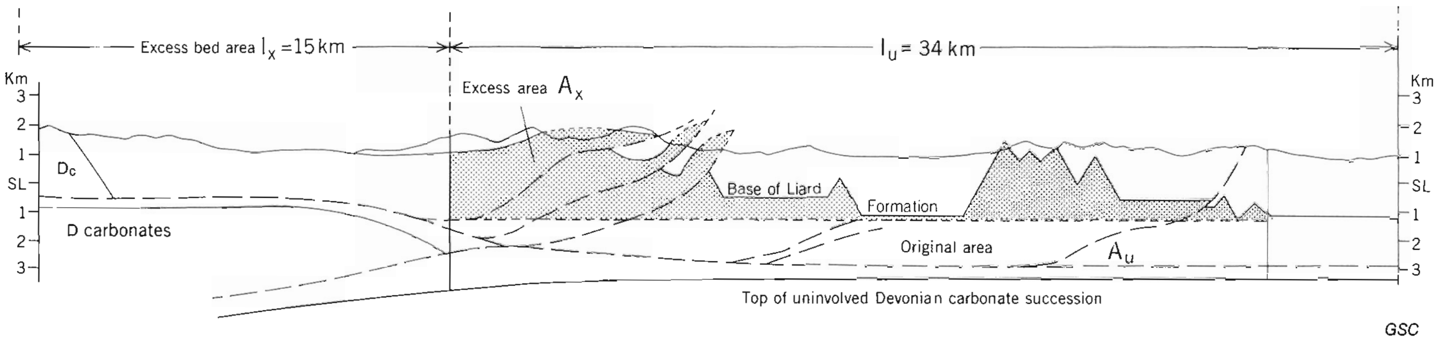


Figure 71. *Sample calculation of shortening across central and eastern portion of the Foothills structural subprovince using area balancing method.*

The concept of a relatively undeformed Devonian basement beneath the Foothills breaks down along the western part of the subprovince, where the synclines are raised well above "regional". In this area, deeper stratigraphic levels must become involved to partially fill the "hole" between the base of the Prophet Formation and the projected surface of the Devonian carbonate succession. This is easily accommodated by introducing a thick thrust sheet of Devonian carbonates on a blind thrust. There is good surface evidence that this type of thrust repetition occurs. At Mount Bertha, 15 km north of the Halfway River, a large, allochthonous sheet of Silurian and Devonian carbonates is exposed in the western margin of the Foothills (Fig. 72). This is evidence that thrust faulting within the Devonian succession does occur along the western margin of the Foothills, and the Bertha Thrust sheet may be interpreted as a splay off blind thrusts in the subsurface.

Therefore, involvement of the Devonian carbonates is interpreted as occurring (in a substantive manner) along the western margin of the Foothills where the bases of synclines no longer remain at "regional".

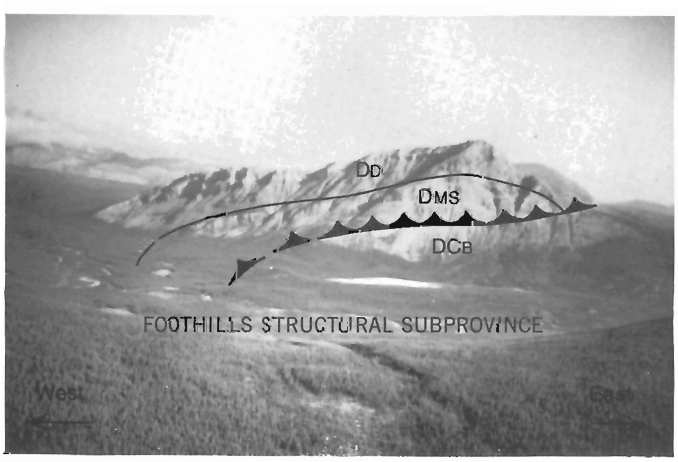


Figure 72. *Northward view of Mount Bertha showing Muncho–McConnell (SDM), Stone (DS) and Dunedin (DD) formations thrust over Besa River Formation (DCB). This fault is interpreted as a splay from a larger 'blind thrust' within the subsurface. (GSC photo no. 2112-3).*

Construction of the Rocky Mountains segment

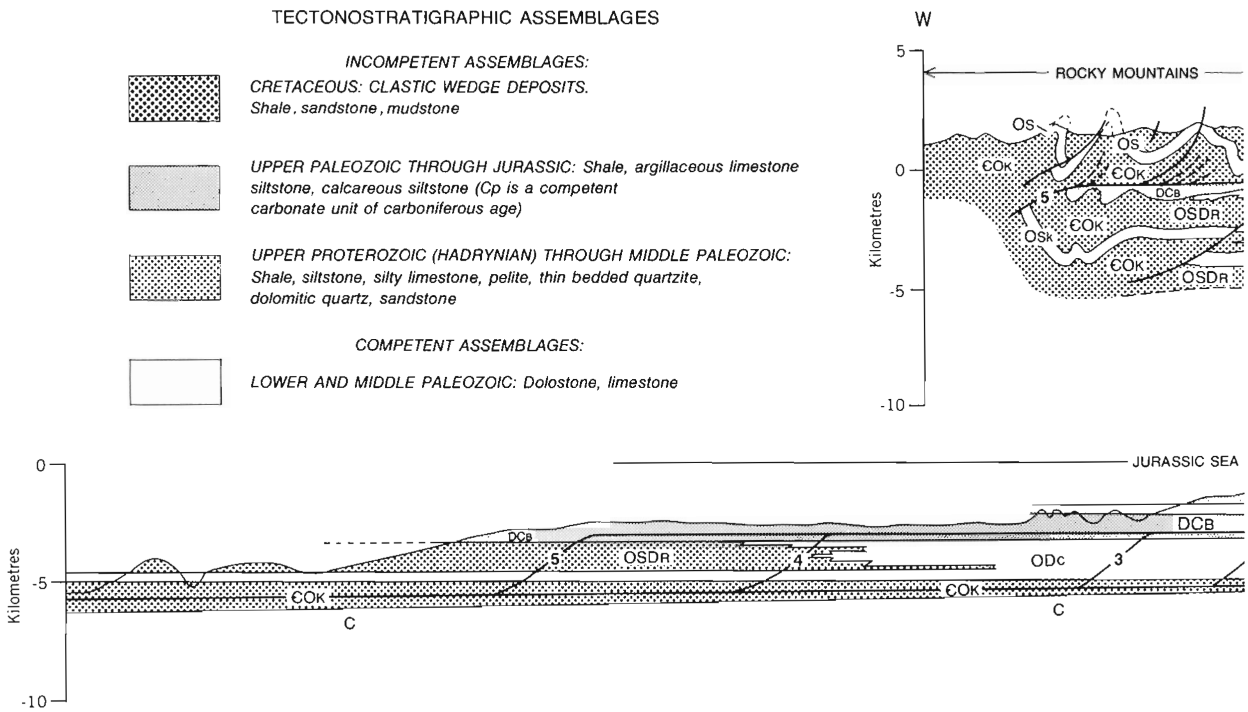
Continuation of the cross-section interpretations westward across the Rocky Mountains subprovince is straightforward. The mountain front is defined by large, allochthonous anticlines transported on blind thrusts that follow a glide zone within the Ordovician Kechika Group; the thrusts cut abruptly through the middle Paleozoic carbonate succession at the mountain front, where they are deflected into the Besa River shale. A hanging wall detachment (hwd) is shown within the allochthonous thrust sheet between the carbonates and the Besa River shale, and is joined to the main blind thrust, which is projected eastward beneath the Foothills. The underlying carbonate succession is interpreted as the westward extension of structurally lower thrust sheets imbricated on a blind thrust(s) beneath the western margin of the Foothills.

West of the mountain front, thrusts and folds are well exposed and, using the geometric constraints established previously, the section can be completed down to the level of the Precambrian Misinchinka Group phyllite and grit. Rigorous balancing west of the mountain front was not attempted; however, bed lengths between major synclines were balanced where possible.

Palinspastic restoration and calculation of net supracrustal shortening

The cross-sections can be unfolded and unfaulted to make a palinspastic restoration showing the original undeformed length, the trace of fault trajectories, and the spatial relation of vertical and horizontal stratigraphic variations to structure. This procedure incorporates the assumption made earlier that bed thickness of the more competent marker units remained essentially constant throughout deformation.

Figure 73, taken from Thompson (1979, 1981), is a palinspastic reconstruction of Structure Cross-sections 2 and 4 in which the eastern margin of the map area is used as a vertical datum, and the top of the Fernie Formation as a horizontal datum. The Fernie Formation is a convenient choice because it represents a sea level datum just prior to the onset of folding and thrusting, and therefore permits an assessment of the initial character of the miogeocline. Extrapolation of this datum well beyond the existing surface exposures of Mesozoic rocks, combined with the lack of information concerning the nature and thickness of upper Paleozoic and Mesozoic rocks in the western region, make the western portion of the reconstruction less reliable.



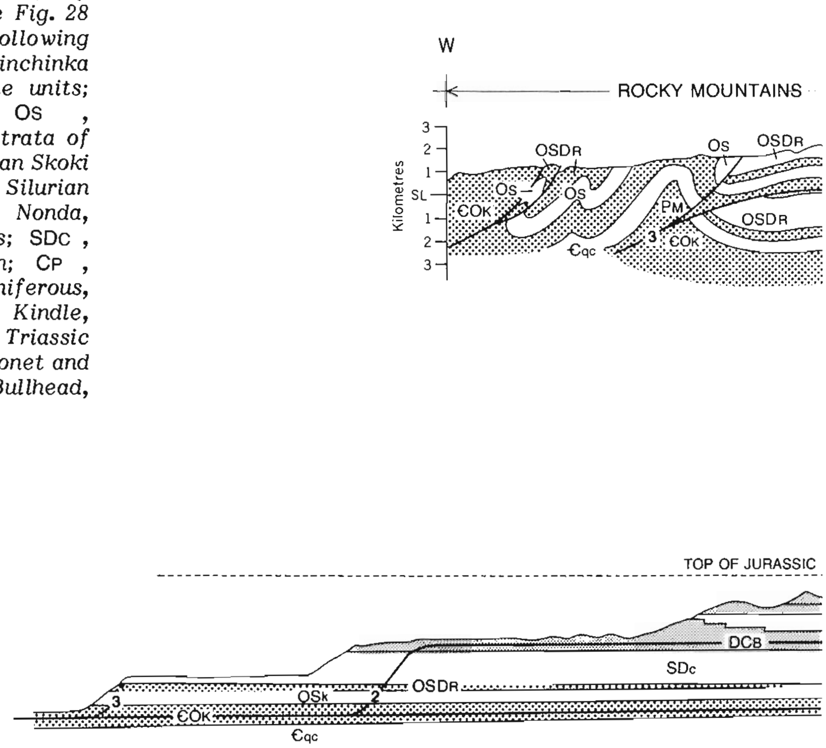
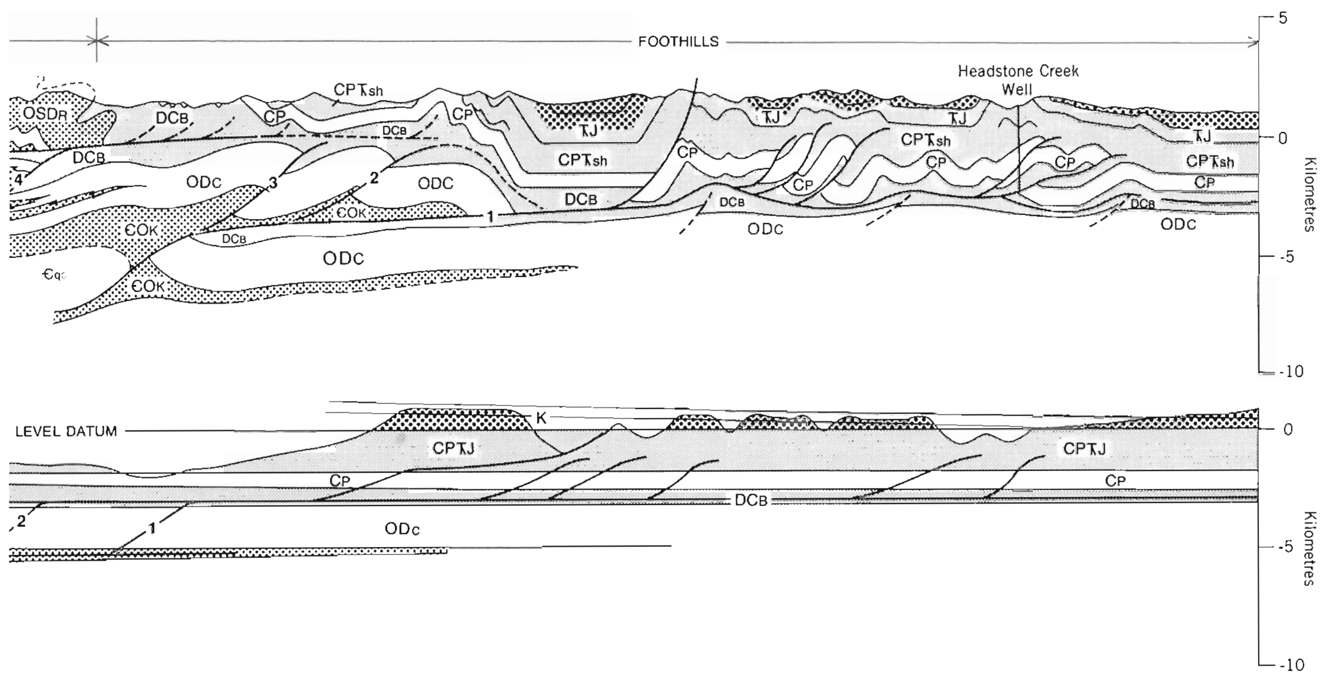
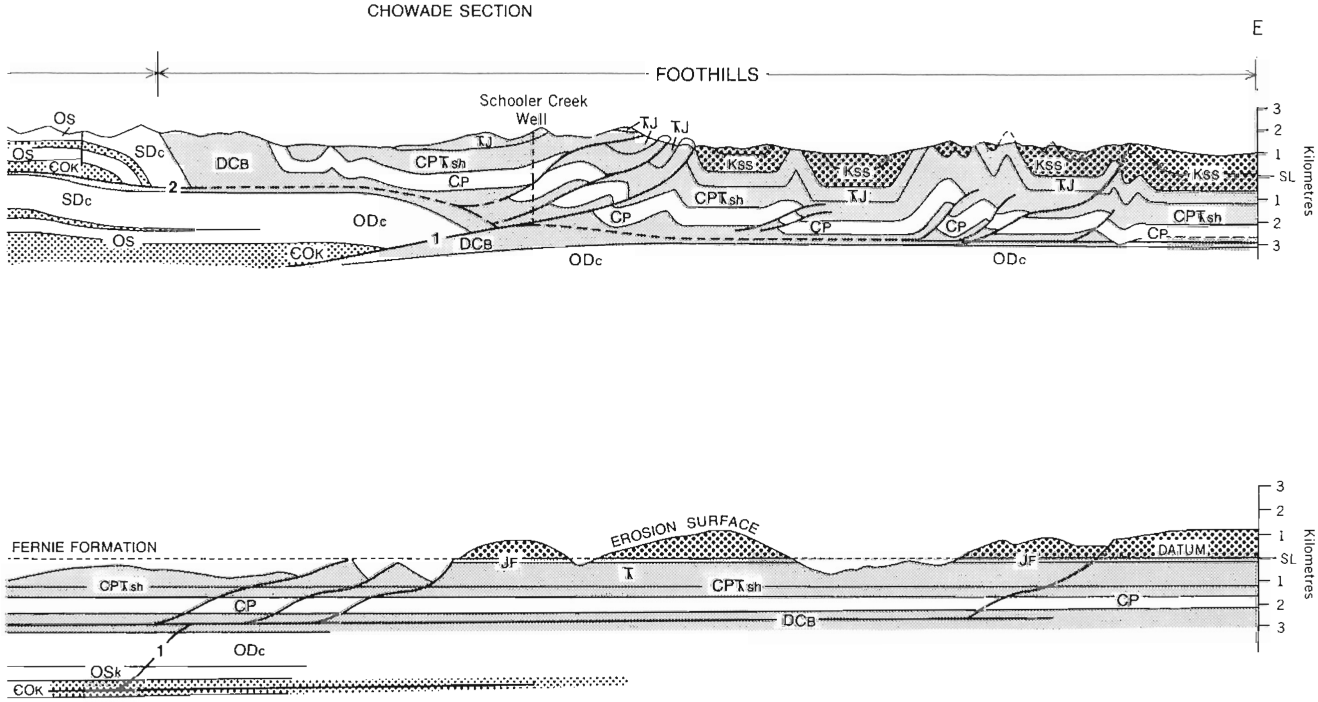


Figure 73. *Generalized structure cross-sections accompanied by palinspastic reconstructions through Halfway River map area (from Thompson, 1979, 1981; see Fig. 28 for section locations). Symbols refer to the following stratigraphic units:* PM *, Upper Proterozoic Misinchinka Group;* Ꞓqc *, Cambrian quartzite and carbonate units;* ꞒOK *, Cambro-Ordovician Kechika Group;* OS *, Ordovician Skoki Formation;* OSDR *, Road River strata of Ordovician through Devonian age;* OS *, Ordovician Skoki Formation;* ODC *and* SDC *, Ordovician or Silurian through Devonian carbonate units – Skoki, Nonda, Muncho–McConnell, Stone and Dunedin formations;* SDC *, Devonian and Carboniferous Besa River Formation;* CP *, Carboniferous Prophet Formation;* CPTsh *, Carboniferous, Permian and Triassic shales – Stoddart Group, Kindle, Fantasque, Grayling and Toad formations;* ŦJ *, Triassic and Jurassic Liard, Charlie Lake, Baldonnel, Pardonet and Fernie formations;* KSS *, Cretaceous Minnes, Bullhead, and Fort St. John groups.*

LAURIER SECTION



CHOWADE SECTION



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Bed lengths were measured along competent marker units, around folds and between faults, and plotted relative to the vertical and horizontal datums. Fault trajectories were plotted from measurements made on the structure sections to show the relative positions of ramps. Facies changes are shown where applicable, and the upper limit of preserved rock is also shown to illustrate the relative effects of individual structures on the quantity of material removed by erosion.

The amount of shortening across the Foothills segment varies between 22 and 30 km, depending upon the section. The variation is large and reflects the many uncertainties involved in cross-section preparation. These values do not include the shortening accommodated within the thickened mass of Besa River shale shown at the mountain front, which leads to the problem of how to distribute and balance the shortening on the blind thrusts with that across the Foothills segment. Shortening is portioned in the following manner: Foothills folds west to and including the last major undeflected syncline are balanced against displacement on blind thrusts beneath the western margin of the Foothills; Foothills structures west of this point to the mountain front are interpreted to balance with fault displacement on the blind thrust at the mountain front. This type of balanced mechanical link between thrusts and folds gets around the problem of underthrusting the carbonate sheets because the compensating fold structures are above and in front of the advancing thrust sheets.

On this basis, the 15 to 20 km of shortening calculated for the central and eastern portions of the Foothills should compensate for a similar amount of displacement on blind thrusts beneath the western margin of the subprovince. The approximately 10 km of shortening estimated on thrusts beneath the mountain front anticlines should be compensated for by an equal amount of shortening within the thickened mass of Besa River shale adjacent to the mountain front, together with the shortening within the folds of Carboniferous Mississippian Prophet carbonates. Therefore, total shortening across the foothills to and including the mountain front is 25 to 30 km.

Values for shortening west of the mountain front were not calculated.

Application of the blind thrust model to other parts of the northern Rocky Mountains

Comparisons of structural style north of the Halfway River map area reveal no significant changes (Fig. 28). The Foothills and Rocky Mountains subprovinces continue without interruption; the mountain front continues to be characterized by large anticlines – the largest of which is the Tuchodi Anticline; and large mountain front thrusts are absent. These observations suggest that whatever geometric model is applied to the Halfway River area should apply, without major modification, to the immediate along strike continuation of the belt. If it does not, the premises upon which the model was derived require re-evaluation.

A test of the blind thrust model across the Tuchodi Anticline was made by Thompson (1981) and the results are summarized below. A cross-section was constructed in which the Tuchodi Anticline was interpreted as a folded, allochthonous thrust sheet transported on a blind thrust that was deflected into Besa River shales beneath the Foothills subprovince (Fig. 74). So that the fault displacement could balance with the amount of shortening within the Foothills subprovince, a thick, subsurface, Proterozoic sedimentary succession was postulated. The results of this exercise were: 1. 21 km of shortening was calculated across the Foothills subprovince; 2. 19 km of displacement was required on the blind thrust beneath the Tuchodi Anticline; and 3. deformation of the crystalline basement surface was not a geometric necessity. On this basis, it appears reasonable to assume that the northern Rocky Mountains area is a thin skinned tectonic regime in which blind thrusts have played an important role. A more sophisticated but conceptually equivalent geometric solution for the Tuchodi Anticline and adjacent structures has recently been prepared by Gabrielse and Taylor (1982).

A hypothesis to explain the regional northward divergence in structural trend

One of the striking patterns evident on a regional compilation such as Figure 28, is the marked northward divergence of structural trends between the southern tip of Bernard Anticline and the northern limit of Muskwa Anticlinorium. Within the eastern, mainly carbonate portion of the Rockies, structural trends vary between north and north-northwest; to the west, trends are northwest, consistent with the regional grain of the Cordillera. This divergence is especially well displayed within the deformed Proterozoic sedimentary rocks of the Muskwa Anticlinorium. The Tuchodi Anticline has a definite north-northwest trend whereas structures west of it have a northwest trend.

Palinspastic restorations of structure cross-sections (Thompson, 1981) demonstrate that the amount of supracrustal shortening does not change significantly along strike, precluding any notion of a hinged rotation of the thrust belt in a clockwise direction about a vertical axis located near the southern end of the Bernard Anticline. It will be argued that the divergence in structural trend was caused by an old (pre-Silurian) basement ramp that has a north-northeast trend and extends from north of the Tuchodi Anticline southward (en echelon?) to near the Peace River. Displacement of allochthonous sheets up and over the basement ramp forced them to bend along the north-northwest trend even though the direction of transport was toward the northeast. Old north-northeast fault trends associated with the ramp would have served as pre-existing lines of weakness that localized later thrust faults associated with Jurassic through early Tertiary deformation. The basement ramp is analogous, therefore, to a thrust fault ramp oriented obliquely to the direction of transport.

A basement ramp is indicated in the restored section drawn across the Tuchodi Anticline (Thompson, 1981; Fig. 74). Profound, pre-Silurian, eastward beveling of the Proterozoic and early Paleozoic sedimentary successions occurs across the strike of the Tuchodi Anticline. In Late Jurassic time, just prior to deformation, this beveling is reflected in a substantial eastward rise in the level of crystalline basement. The trend of this basement feature is interpreted to be coincident with the trend of the Tuchodi Anticline for two reasons: 1. pre-Ordovician dykes, suggestive of crustal distension, have a north-northwest trend, and 2. a large normal fault that is pre-Silurian in age and across which stratigraphy is missing on the east side, also has a trend coincident with the Tuchodi Anticline (Taylor and Stott, 1973, Map 1343A). This surface evidence suggests a period of crustal distension and faulting – east side up – along a north-northwest trend prior to deposition of Silurian carbonate rocks. A thrust propagating at or near the basement sedimentary rock interface might be deflected up over the ramp rather than cut through it; in the process the allochthonous thrust sheet would be forced to fold in

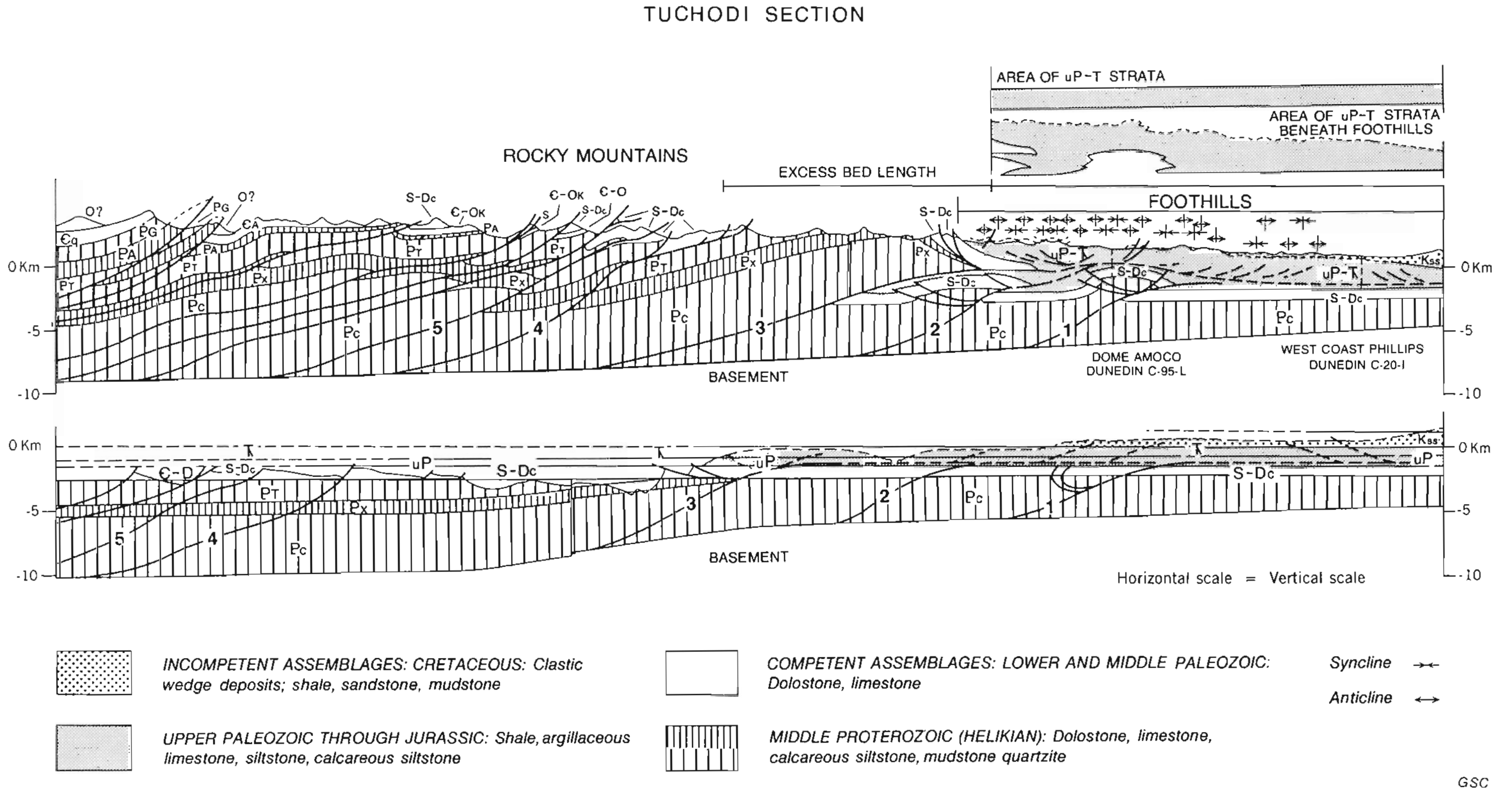

Figure 74. Generalized structure cross-section accompanied by palinspastic reconstruction through Tuchodi Lakes map area (from Thompson, 1981; see Figure 28 for location of section). Symbols refer to the following stratigraphic units: Pc *, mid-Proterozoic Chischa Formation;* Px *, mid-Proterozoic Tetsa, George, and Henry Creek formations;* PT *, mid-Proterozoic Tuchodi Formation;* PA *, mid-Proterozoic Aida Formation;* PG *, mid-Proterozoic Gataga Formation;* ЄOK *, Cambro-Ordovician Kechika Group;* SDc *, Silurian through Devonian carbonate units – Nonda, Muncho–McConnell, Wokkpash, Stone, and Dunedin formations;* uPT̄ *, upper Paleozoic through Triassic Besa River, Kindle, Fantasque, Grayling, Toad, Liard, Charlie Lake, Ludington, Baldonnel, and Pardonet formations;* Kss *, upper Cretaceous Fort St. John Group, Dunvegan and Kotaneelee formations (see Taylor, 1973 for stratigraphic descriptions).*

accordance with the ramp geometry (see Figure 75 for diagrammatic illustration of this principle). In addition, the thrust might follow the path of a pre-existing, high angle fault which would affect the angle at which the thrust cut up through the sedimentary package. Both possibilities lead to surface trends parallel with the pre-existing basement ramp.

Persistence of the north-northwest trend southward to near the Peace River suggests that the old basement feature has a considerable strike length.

A comparison with the southern part of the Canadian Rocky Mountains

Most variations in structural style along the strike of the Canadian Rockies are attributable to changes in the overall mechanical character of the layered rock sequence. In the north, there is a greater proportion of incompetent relative to competent rock units that provided greater potential for the formation of décollement zones as well as structural disharmony between different stratigraphic levels. But other differences, such as the narrower width of the northern Rockies, the reduced amount of supracrustal shortening, and the smaller volume of Late Cretaceous and early Tertiary foredeep deposits are not as readily explained. Changes northward in the lithostratigraphic character of the miogeocline should not have affected the amount of shortening or the width of the belt, only the surface expression of the shortening.

A one-to-one comparison of the northern Rockies with the southern Rockies is not necessarily valid, because it does not take into account the potential for variation in the rate at which deformation may have proceeded through the miogeocline. For instance, if the miogeocline had an essentially constant width from south to north, and if the duration of orogenic events was the same along the miogeocline, a variation in the intensity of the orogeny along the strike would be measurable in terms of how far across the miogeocline strain was transmitted. Figure 76 (Thompson, 1979, 1981) is an attempt to measure that variation.

The miogeocline has a hinge area beyond which most units thicken rapidly. In Figure 76, palinspastic plots of the miogeocline for the south (Price and Mountjoy, 1970) and for the north (Thompson, 1979) are compared by aligning their hinges (for Paleozoic successions) and then superimposing the position of the structural belts. The eastern limit of strain is the edge of the Foothills subprovince.

Three, interdependent observations are apparent:

1. Deformation in the north did not progress as far cratonward as in the south. A 20 km width (at least) of deformed cratonic and clastic wedge deposits is structurally involved in the southern cross-section but remains undeformed in the northern cross-section.

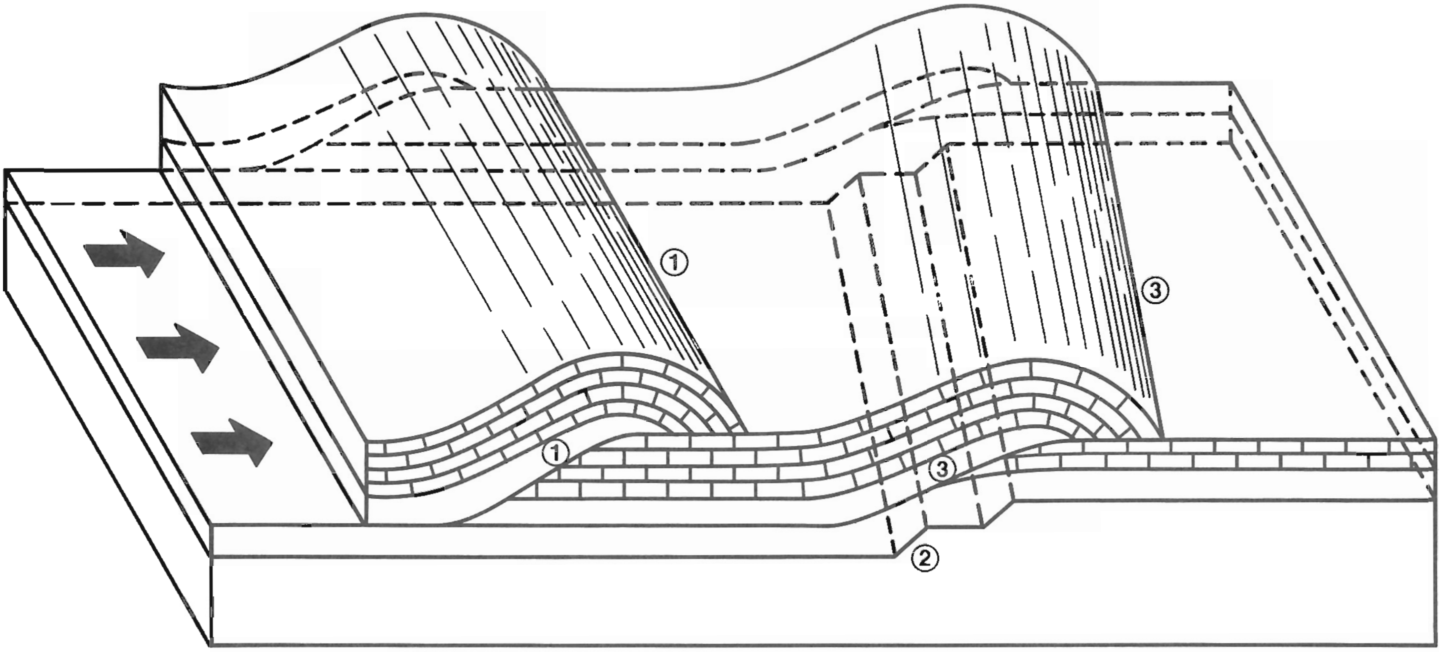

Figure 75. Illustration of the potential influence of old basement features on structural trends formed during a younger thrust and fold event: Thrust (1) cuts up through layered succession, forming fault ramp and anticline trending perpendicular to the direction of maximum applied horizontal stress; deformation then propagates eastward but is inhibited by a pre-existing basement deflection (2). Thrust (3) is forced to cut upsection along the deflection; displacement along thrust (3) forms an anticline trending parallel to the ramp. This trend is inherited from the basement deflection and is oblique to the direction of maximum applied horizontal stress. The northward divergence of structural trend in the northern Rocky Mountains is interpreted as having originated in this manner.

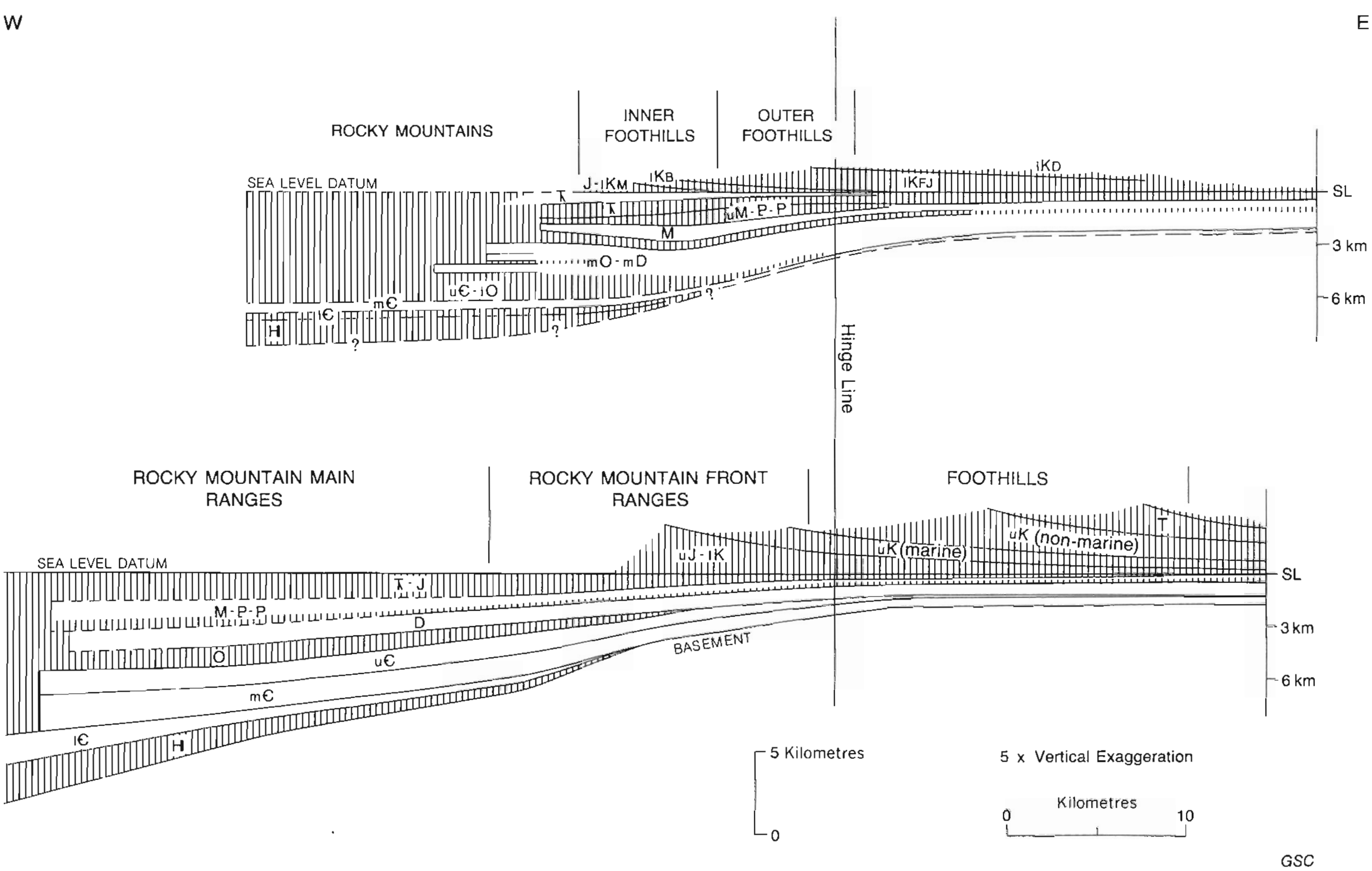

Figure 76. Comparison of restored sections of miogeocline for the Halfway River area and for the southern Rocky Mountains (southern section taken from Price and Mountjoy, 1970). Hinge lines (for lower and middle Paleozoic strata) used as common datum for comparative purposes (from Thompson, 1981).

2. The northern Rocky Mountains structural subprovince occupies a cross-section stratigraphic position within the miogeocline equivalent to the Main Ranges structural subprovince in the south.

3. In the north, there is no analogue for the closely spaced, imbricated sheets of cratonic, Paleozoic, carbonate rocks that characterize the southern Rockies Front Ranges structural subprovince, because platform carbonate rocks occupying an equivalent cross-section position in the north remain essentially undeformed.

The eastern limit of deformation in each cross-section (Fig. 76) may be regarded as the preserved external limit of a strain front that passed progressively from west to east through the miogeocline. The narrower width and decreased amount of foreshortening in the north can be related directly to the observation that a narrower width of the sedimentary prism was deformed; if strain had persisted farther eastward in the north, the platform carbonate succession that currently forms a basement to Foothills folds, would have become imbricated and uplifted as deeper décollement surfaces were extended eastward, resulting in a thrust-faulted structural subprovince similar in character to the Front Ranges of the southern Rocky Mountains. The thickness and lateral extent of the young portion of the adjacent clastic wedge deposits would also have been affected. Increased loading of the lithosphere by thrust sheets during Late Cretaceous and early Tertiary time would have caused a deepening and enlargement of the adjacent foredeep basin (Price and Mountjoy, 1970) and increased the potential for accumulation of a thicker, more extensive clastic wedge sequence of that age. Instead, the northern Canadian Rocky Mountains represent a less complete stage in structural development of the foreland belt, a stage that preceded development of a Front Ranges-type structural domain. This does not mean that deformation ended sooner; on the contrary, Lower Cretaceous foredeep deposits form a major component of the Foothills structural subprovince (Stott, 1975) which was deformed in Late Cretaceous (and early Tertiary?) time, synchronous with the major Late Cretaceous – early Tertiary tectonic pulse to the south. However the intensity of that pulse decreased northward.

The similarities between the northern and southern Rocky Mountains are more apparent if one restores the deformed edge of the southern Rockies to a more westerly position and unfaults most of the Front Ranges subprovince in the process. Then the important points of reference are the Purcell and Muskwa anticlinoria. Within such a framework, a crude symmetry exists between north and south.

ECONOMIC GEOLOGY

GEOLOGY AND CHARACTER OF CARBONATE-HOSTED LEAD-ZINC OCCURRENCES

Introduction

Discovery of lead-zinc mineralization near Robb Lake in 1971 sparked an exploration and staking rush in the northern Rocky Mountains of British Columbia that led to the recognition of a potential new lead-zinc belt. Numerous carbonate-hosted showings were discovered, but few of them are sufficiently large or rich to merit further development at present. One group of showings in which interest remains reasonably high is at Robb Lake, where 6.1×10^6 short tons (5.5 Tg) of 7.3 per cent combined lead-zinc are reported in rocks of Devonian age (Northern Miner, January 30, 1975).

Macqueen and Thompson (1978) described the regional geological setting and general character of some of the more significant carbonate-hosted lead-zinc occurrences. They related the emplacement of lead and zinc sulphides, and the petroleum residues associated with the sulphides, to the total diagenetic history of the source and host rocks, linking the mineralizing process to the geological conditions necessary for maturation, migration, and emplacement of petroleum. This approach predicts that mineralization occurred long after deposition of the host rocks. Consequently, the breccias, hosts of the mineralization, were the result of a late breccia forming mechanism – perhaps hydraulic fracturing – associated with burial-induced high pore-fluid pressures. A brief account of their results follows.

General remarks

Table II provides information on seven lead-zinc occurrences in the Peace River – Prophet River area (Fig. 77). The following generalizations are apparent:

1. All occurrences are carbonate-hosted, and are located at or near the carbonate shelf edge, with the possible exception of the Richards Creek showings
2. The Muncho-McConnell/Stone interval appears to have the most promising economic potential; however, any part of the Upper Silurian through Middle Devonian carbonate sequence may be mineralized
3. Mineralization occurs in a variety of textures and spatial settings: as part of a breccia matrix, as vein and fracture fillings, as open-space (vug, etc.) fillings, as fine grained replacements of host carbonate, and as massive, coarse crystalline pods
4. Mineralogy is simple: sphalerite and/or galena with or without pyrite or marcasite; sparry white dolomite, bitumen, and quartz are the common gangue minerals. Sphalerite is 5 to 10 times more abundant than galena
5. None of the occurrences in either assemblage show evidence of significant carbonate solution, nor are they spatially related to overlying unconformities
6. There is no obvious evidence of control of mineralization by Mesozoic – Tertiary tectonism
7. Fluorite and barite, although known in this stratigraphic interval in a number of localities north of the Peace River – Prophet River area (Taylor et al., 1975; Morrow et al., 1976), do not appear to be closely associated with lead-zinc mineralization.

Muncho-McConnell, Stone occurrences

The most common form of sphalerite and galena mineralization in this interval is that associated with crackle, mosaic, and rubble breccia[1], which are most extensively developed in the Robb Lake area. The Mount McCusker showings occur in breccias in the lowest exposed part of the Muncho-McConnell Formation. The Richards Creek and Lady Laurier Lake showings occur as fracture fillings.

In breccia occurrences, anhedral crystals or aggregates of crystals of sphalerite occur as rims on breccia fragments, or as scattered 'blebs' within the white sparry dolomite that forms a matrix to the breccias. Sphalerite crystals and aggregates range from pinhead size to several centimetres; larger masses are very rare. Sphalerite is most commonly anhedral, and seems to have been extensively shattered. Colloform textures have been reported at Robb Lake, but are extremely rare. Galena is much less abundant in most showings but occurs commonly as euhedral 'blebs' from pinhead to (rarely) 1 to 2 cm cubes, and mostly in close association with sphalerite.

Dunedin occurrences

Mineralization in the middle and upper parts of the Dunedin Formation at Redfern Lake (Fig. 77) consists of widely scattered, disseminated pore fillings and replacements within fine to medium crystalline dolomite, associated with euhedral crystals of quartz that have grown within the dolomite as a replacement. At Robb Lake, sphalerite mineralization occurs in the middle part of the Dunedin in close association with white sparry dolomite, which occurs as pseudobreccia replacements of Dunedin limestones and dolostones, and as fracture fillings. At the Nabesche River occurrence (Nabe Property) in the upper unit of the Dunedin Formation[2], sphalerite occurs as rare blebs or euhedral crystals up to 4 cm in size, within the dolomitized parts of the unit. Pyrite, bladed quartz crystals, and white sparry dolomite are intimately associated with the sphalerite; galena is unknown.

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[1]Crackle breccia: little displacement of fragments; mosaic breccia: fragments largely but not wholly displaced; rubble breccia: fragments completely displaced, no fragments match.

[2]The upper part of the Dunedin Formation was called the Slave Point Formation by Macqueen and Thompson (1978, p. 1744).

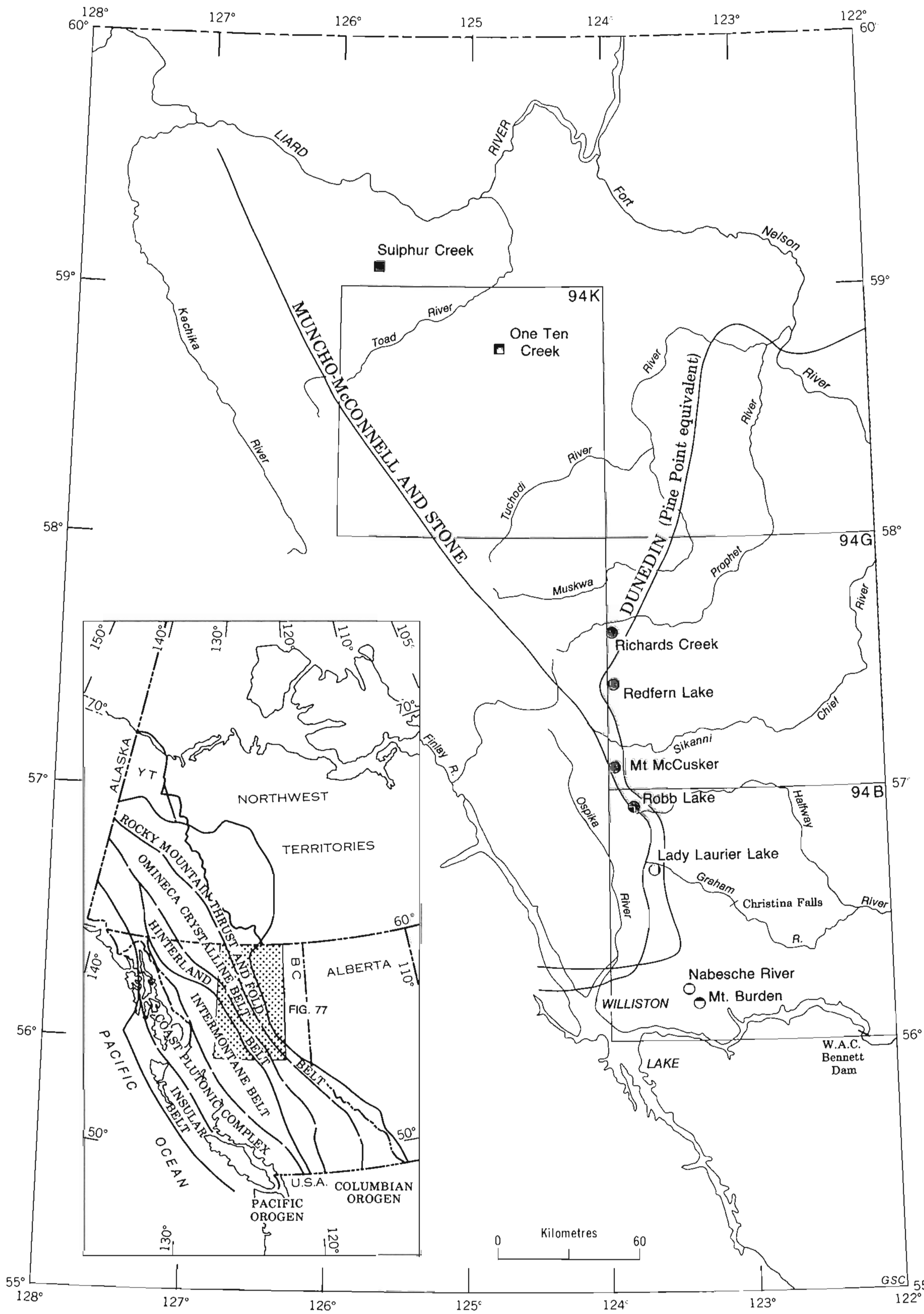

Figure 77. *Map of northeastern British Columbia showing location of carbonate-hosted lead-zinc occurrences (solid circles) and breccia-hosted barite occurrences (solid squares) relative to major shelf to off-shelf facies transitions (inset) (from Macqueen and Thompson, 1978). Solid lines represent western limit of carbonate facies (see Fig. 4).*

TABLE II

Setting and character of zinc-lead showings, northeastern British Columbia (after Macqueen and Thompson, 1978)

Showing	Formation, stratigraphic position and setting	Host rock	Mineralized zones		
			Type	Geometry	Associated minerals, hydrocarbons
Richards Creek	STONE FM., upper ≈ 200 ft.; DUNEDIN FM., lower ≈ 100 ft.; (Dunedin Fm. overlain by Besa River Fm.)	Dolostone, fine to medium crystalline	Fracture and vein fillings; small massive pods; open space fillings; part replacements of host dolostone; mineralization fracture controlled	Locally discordant; individual occurrences small, with no obvious lateral continuity	Pyrite, marcasite, quartz, bitumen (?)
Redfern Lake	DUNEDIN FM., middle (?) and upper parts; very close to lateral facies change to Besa River Fm.	Dolostone, medium crystalline carbonaceous; limestone	Disseminated pore fillings and replacements in dolostone; fracture fillings in limestone; mostly sphalerite	Unknown	Closely associated with quartz; bitumen
Mt. McCusker	MUNCHO - McCONNELL FM., lowest ≈ 500 ft.	Dolostone, fine to medium crystalline	Medium to coarse crystalline aggregates in matrix of angular dolostone breccias; as rinds on breccia fragments	Breccias locally discordant, but make up a zone with lateral extent of several thousand feet	Quartz, minor pyrite, bitumen (?)
Robb Lake	MUNCHO - McCONNELL, STONE INTERVAL, upper two thirds; middle part of Dunedin Fm.; (Dunedin Fm. overlain by Besa River Fm.)	Dolostone, fine to medium crystalline quartzose	Angular breccias, crackle and mosaic; pseudobreccia; crystal aggregates of sphalerite and galena in breccia matrix; sphalerite rinds on breccia fragments; fine grained disseminations in carbonaceous zones	Breccias discordant, considerable lateral and vertical dimensions; possible stratigraphic control	Quartz, calcite, pyrite, bitumen
Lady Laurier Lake	MUNCHO - McCONNELL FM., lower third	Dolostone, fine to medium crystalline	Unknown	Unknown	Quartz
Nabesche River	SLAVE POINT FM., lower part	Dolostone, originally echinoderm mudstone	Blebs associated with white dolostone	Irregular cross cutting patches	Smithsonite, bitumen
Mt. Burden	STONE FM., upper ≈ 100 ft.	Limestone, micritic	Scattered blebs associated with white dolostone; mainly galena	Irregular cross cutting patches	Quartz, bitumen, smithsonite; pyrophyllite in overlying Besa River Fm.

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Apparent paragenesis

Regional observations of host rocks and relationships evident at and between the individual showings described in Table II can be used to construct a possible paragenesis.[1] In Figure 78, the first or early group of minerals listed represents the path from freshly accumulated lime sediments to lithified carbonate rocks composed of dolomite or low-magnesium calcite. The second group of minerals includes white sparry dolomite, sulphides, quartz, bitumen, and calcite, and is interpreted as representing separate and perhaps much later processes that took place within the carbonate sequences after lithification. Sulphides are always found in association with white, coarse crystalline, sparry dolomite, but there is much more white sparry dolomite in the Upper Silurian through Middle Devonian sequence than sulphides. In mineralized zones, quartz is most common within dolomite beds rich in quartz sand; it commonly lines open spaces and appears to have crystallized later than most of the white sparry dolomite. Bitumen is commonly associated with the sulphides but is not present everywhere. Calcite appears to be the latest-forming gangue mineral, and commonly fills small cavities or lines open spaces.

The suggestion that white sparry dolomite, sphalerite, galena, quartz, and pyrobitumen are all related in some manner and are of "late" origin is supported by the intimate association of these elements with one another as observed in outcrops, hand specimens, and thin sections (e.g., Fig. 3 in Macqueen, 1976). In Figure 78, a time gap is inferred between early and late processes; the length of this gap is unknown.

Sulphur isotope results

Values of $\delta^{34}S$ for sulphides fall within the range of +5.0 to +17.5 per mil. (Table III), close to values reported from "type" Mississippi Valley lead-zinc deposits (see Heyl et al., 1974) and from Cambrian carbonate-hosted zinc-lead deposits of the southern Rockies (Evans et al., 1968). The data in Table III (Morrow et al., 1976) suggest that the barite and sulphide deposits are genetically distinct.

Metal content of carbonates and shales

Metal content of Upper Silurian and Devonian carbonate rocks is low (Table IV), supporting the conclusion that the carbonate host rocks were not the metal source. Metal content of Besa River shale is also low (less than 100 ppm Zn; Pb at detection level).

The Robb Lake occurrences

Robb Lake is the largest and most promising area of mineralization discovered within the shelf carbonates of northeastern British Columbia. There are at least twelve separate showings exposed along the axis and western flank of the Robb Anticline (Fig. 5 in Macqueen and Thompson, 1978). Each showing is associated with a breccia body in which spalerite and galena form part of the breccia matrix (plates XVA, XVB in Thompson, 1974; Fig. 10 in Macqueen and Thompson, 1978). The difficulties of assessing the potential of this deposit are twofold: 1. it is not possible to

[1]A table showing regional mineralogy and Zn, Pb and Cu content of the Besa River and adjacent formations has been placed in the Depository of Unpublished Data. Photocopies are available, at a nominal charge, from the Depository of Unpublished Data, C.I.S.T.I., National Research Council of Canada, Ottawa, Canada K1A 0S2.

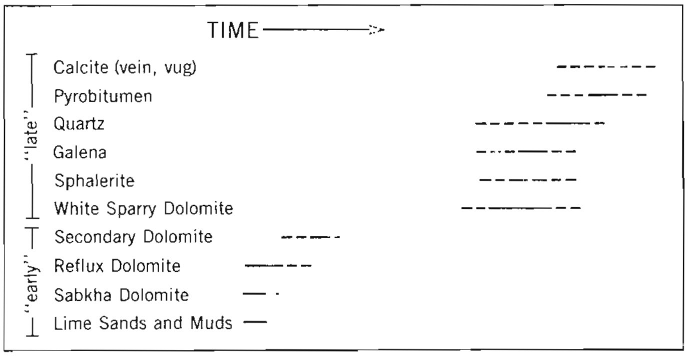

Figure 78. Apparent paragenesis in Upper Silurian through Middle Devonian mineralized carbonates, northeastern British Columbia. Elements designated as 'early' involve path from unconsolidated sediments to lithified rocks; those designated as 'late', developed or were emplaced within lithified rocks. Possible time gap between 'early' and 'late' elements is speculative (from Macqueen and Thompson, 1978).

TABLE III

Metal contents in selected carbonate units
(from Macqueen and Thompson, 1978)

Sample	Location	Formation	Stratigraphic level	Mineral	$\delta^{34}S$‰[1]
1	Robb Lake	MUNCHO-McCONNELL, STONE INTERVAL	Breccia zone in creek, one third below top	Sphalerite, red	+17.5
2				Sphalerite, yellow	+15.7
3			North face showing	Sphalerite, red	+13.5
4			As sample 1	Galena	+9.6
5	Mount Burden	STONE	Uppermost part	Galena	+6.5
6					+6.2
7	Nabesche River	SLAVE POINT	Lower part	Sphalerite, red	+5.0
8				Sphalerite, yellow	+13.8
9	Robb Lake	BESA RIVER	Basal bed, section 1	(Bitumen)	−4.2

[1]Analyses completed at Stable Isotope Laboratory, Department of Physics, University of Calgary. Values represent averages of duplicate analyses.

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predict the dimensions and geometry of individual breccia bodies; and 2. there are no trace element distribution patterns or alteration zones or halos associated with mineralized breccias to serve as proximity indicators. Consequently, evaluation of economic potential necessitates grid drilling.

Origin of the host breccias

Breccia origin(s) has not been established. Macqueen and Thompson (1978, 1979a, b) used geological field relations, an analysis of insoluble residues, burial history, diagenetic history and inferred time relationships to support a relatively late development of the breccias – perhaps by a hydraulic fracture mechanism. Sangster (1979) disputed this analysis, preferring a solution collapse origin associated with erosion and paleokarst development prior to deposition of the Dunedin Formation (Sangster and Lancaster, 1976). Sangster's thesis is based largely on the textural similarity of Robb Lake breccias with those developed within Ordovician

TABLE IV

Sulphur isotope data
(from Macqueen and Thompson, 1978)

	n	Zn(ppm)			Cu(ppm)		
		Range	$\overline{X}$	σ	Range	$\overline{X}$	σ
NONDA FORMATION (SILURIAN), Robb Lake[2]							
<50% carbonate[3]	2	3,26	–	–	19,38	–	–
50-90% carbonate[3]	14	<4-113	25	39.2	6-18	12	<4
>90% carbonate[3]	10	<4-43	13	15.5	6-12	9	<4
All samples >50% carbonate[2]	24	<4-113	20	31.6	6-18	11	<4
STONE, MUNCHO-McCONNELL FORMATIONS, Robb Lake[4]							
All samples	37	<4-150	33	44.0	<4-44	12	9.3
<50% carbonate[3]	6	<4-140	38	55.3	20-44	30	8.3
50-90% carbonate[3]	17	<4-150	42	50.9	5-26	11	5.2
>90% carbonate[3]	14	<4-90	21	26.7	<4-12	7	2.4
White sparry dolomite and country rock, DEVONIAN, northeastern B.C.[5]							
	n	Zn(ppm)			Pb(ppm)		
		Range	$\overline{X}$	σ	Range	$\overline{X}$	σ
White sparry dolomite							
>90% dolomite[6]	19	4-28	11	8.2	4-9	6	1.5
80-90% dolomite[6]	14	4-86	21	20.3	4-8	6	1.0
Country rock	15	<4-20	4	4.6	<4-10	5	2.3

[1] Elemental analyses by atomic absorption: analyst R.R. Barefoot, formerly of Geochemistry Section, I.S.P.G., Calgary. Methods described in Macqueen et al. (1975).
[2] Samples from measured section in Robb Lake area, 56°57′N, 123°45′W. Pb values below limit of detectability for all samples (= 4 ppm).
[3] Carbonate content determined semiquantitatively by X-ray diffraction using peak-height ratios.
[4] Samples from measured section in Robb Lake area, 56°57′N, 123°45′W. Samples mostly light yellowish brown dolomite typical of the interval, but white dolomite (see text) not excluded. Pb values below limit of detectability (= 4 ppm) for all samples.
[5] Samples from five stratigraphic sections in Macqueen and Taylor (1974). Pine Point Formation, 35 samples; Stone Formation, 12 samples; Beaverhill Lake Formation (Upper Devonian), 1 sample. No significant difference in samples from different formations.
[6] Dolomite content determined semiquantitatively by X-ray diffraction using peak-height ratios.

GSC

rocks of eastern Tennessee. The comparison may be valid but the conclusion that this supports a solution collapse origin does not necessarily follow. The mode of origin of eastern Tennessee breccias remains moot (Callahan, pers. comm., 1979).

Origin of the metals and mode of emplacement

The model proposed by Macqueen and Thompson (1978; see also Macqueen, 1976) is based on models for the migration and accumulation of hydrocarbons into carbonate reservoirs, namely, that zinc and lead were stripped from the host shale during dewatering, and flushed into the nearest available reservoir, in this case open breccias within the Muncho-McConnell, Stone and Dunedin formations. The close spatial association of bitumen with sphalerite in the breccia matrix is critical to the argument. That hydrocarbons and metals were emplaced within the breccias at about the same time suggests they were carried within the same plumbing system and ultimately flushed from the same source. Subsequent discovery of significant lead and zinc occurrences within Upper Devonian clastic rocks to the northwest, in Ware map area (McIntyre, 1980), supports the notion that off-shelf clastic facies may represent a potential metal source; however, there is no proof, as yet, that the Robb Lake occurrences were so derived. Other factors may be important, such as episodes of crustal extension producing steep dipping fault conduits capable of localizing the flow of meteoric, epithermal and even crustal fluids.

Postscript

There are significant differences in Pb isotope data between Devonian carbonate-hosted and Devonian shale-hosted deposits in northeastern British Columbia: both types indicate multistage histories of development, but carbonate-hosted Pb is more radiogenic and more variable in composition (Morrow and Cumming, 1982). One stage in the development of carbonate-hosted Pb "may have been a period of residence of Pb in uranium-rich Paleozoic shale, which caused the development of radiogenic leads that were emplaced later in the carbonate-hosted lead-zinc environments" (Morrow and Cumming, *ibid.*, p. 1070). These conclusions lend conditional support to the idea of a relatively late stage emplacement of mineralization for Robb Lake and related carbonate-hosted deposits in northeastern British Columbia.

Property descriptions
(see Figure 79 for property location)

Poco

Location. 94 B/3W; lat. 56° 10.5' - 13'
long. 123° 20.4' - 23.7'

At approximately 6500 feet (1980 m) elevation; 1 mile (1.6 km) east of Mount Burden

Owner. 1972 - 1974; Union Oil Co. of Canada Ltd., Calgary
1975 - present; British Newfoundland Exploration Ltd., Vancouver

Metals. Lead, zinc

Description. Minor sphalerite and galena as disseminations and vein fillings within brecciated Dunedin Formaton dolostone adjacent to (in hanging wall of) the Burden Thrust

Work done. Surface mapping; soil and rock geochemistry; four IEX diamond-drill holes totalling 269 m

Status. Inactive

References. British Columbia Ministry of Energy, Mines and Petroleum Resources, Annual Reports, 1972, p. 460; 1973, p. 384; 1974, p. 288; 1975, p. E154; 1976, p. E168; 1977, p. E210

Cliff, Lost

Location. 94 B/4E; lat. 56° 02' - 05.5'
long. 123° 32' - 37.5'

Between elevations 2300 (701 m) and 6000 feet (1828 m) on the north shore of Williston Lake, 2 miles (3.2 km) east of Wicked River

Owner. Trojan Consolidated Mines Ltd., Vancouver

Metals. No showings reported

Description. Claims underlain by Muncho-McConnell, Stone and Dunedin formations

Work done. Surface mapping, silt and soil sampling

Status. Inactive (lapsed?)

References. British Columbia Ministry of Energy, Mines and Petroleum Resources, Annual Report, 1972, p. 460

Rush

Location. 94 B/4E; lat. 56° 02.4' - 06'
long. 123° 33' - 36'

Between 3000 feet (914 m) and 5500 feet (1676 m) elevation north of the Peace River between Hamlyn and Cowart creeks

Owner. Sicintine Mines Ltd., Vancouver

Description. Claims underlain by Stone and Dunedin formations

Metals. No showings reported

Work done. Stream silt sampling

Status. Inactive (lapsed?)

References. British Columbia Ministry of Energy, Mines and Petroleum Resources, Annual Report, 1972, p. 461

WI, Zrr

Location. 94 B/5E; lat. 56° 15' - 17'
long. 123° 37.5' - 39.8'

At approximately 3500 feet (1066 m) on Wicked Lake

Owner. Cordilleran Engineering Ltd., Vancouver

Metals. Copper, lead, zinc

Description. Showings in Dunedin Formation in footwall of Burden Thrust. Pyrite and/or marcasite occur as pods, veins and disseminations in brecciated dolostone near fault. Thick gossans are associated with the mineralization. Smithsonite and/or hemimorphite occur along fractures, on bedding surfaces, and as part of breccia cement

Work done. Surface mapping; trenching; soil and rock geochemistry

Status. Inactive

References. British Columbia Ministry of Energy, Mines and Petroleum Resources, Annual Reports, 1972, p. 461; 1974, p. 288; 1977, p. E210

Rush

Location. 94 B/5E; lat. 56° 22' - 25'
long. 123° 35.5' - 39'

Between 5600 feet (1707 m) and 6500 feet (1981 m) elevation on the north side of Gauvreau Peak, at the headwaters of Nabesche River

Owner. Spartan Explorations Ltd., Vancouver

Metals. No showings reported

Description. Claims underlain by Muncho-McConnell and Stone formations

Work done. Surface geological mapping

Status. Lapsed

References. British Columbia Ministry of Energy, Mines and Petroleum Resources, Annual Report, 1972, p. 461

Lara

Location. 94 B/5R; lat. 56° 27'
long. 123° 36'

Between 4300 and 6000 feet (1311 and 1829 m) near Aley Creek

Owner. Moneta Porcupine Mines Ltd., Vancouver

Metals. No showings reported

Description. Adjacent to Burden Thrust separating Ordovician and Silurian shales from Muncho-McConnell Formation

Work done. Surface geological mapping; soil sample survey

Status. Lapsed

References. British Columbia Ministry of Energy, Mines and Petroleum Resources, Annual Report, 1972, p. 461

Nabe

Location. 94 B/6W; lat. 56° 15' - 20.3'
long. 123° 23.2' - 28'

Between 5000 and 6000 feet (1524 and 1829 m) elevation at headwaters of Nabesche River

Owner. Cominco Ltd., Vancouver, BX Development Ltd

Metals. Zinc

Description. Minor sphalerite in reefal dolostone of the Dunedin Formation

Work done. Surface mapping; soil geochemistry; seven diamond-drill holes totalling 927 feet (283 m)

Status. Inactive

References. British Columbia Ministry of Energy, Mines and Petroleum Resources, Annual Report, 1972, p. 462

Pres, Quille

Location. 94 B/6W; lat. 56° 22.3'
long. 123° 27'

Owner. Cordilleran Engineering Ltd.

Metals. No showings reported

Description. Claims underlain by Stone, Dunedin and Besa River formations along eastern flank of Bernard Anticline

Work done. Surface geological mapping; silt and soil geochemistry

Status. Lapsed

References. British Columbia Ministry of Energy, Mines and Petroleum Resources, Annual Report, 1972, p. 463

Brin

Location. 94 B/12E; lat. 56° 42' - 40'
long. 123° 36' - 35'

8 km southeast of Lady Laurier Lake, at the headwaters of the Graham River

Owner. British Newfoundland Exploration Ltd., Vancouver

Metals. Lead and zinc

Description. The claims are underlain by Nonda, Muncho-McConnell, Stone and Dunedin formations. Galena and sphalerite occur within breccias in the Dunedin Formation

Work done. Surface mapping; stream, soil and rock geochemistry; one diamond-drill hole totalling 48 m

References. British Columbia Ministry of Energy, Mines and Petroleum Resources, Annual Reports, 1972, p. 462; 1973, p. 387; 1975, p. E155

Ash

Location. 94 B/12E; lat. 56° 42.8' - 45.7'
13E long. 123° 35.5' - 38.7'

11 km east-northeast of Mount Lady Laurier

Owner. Bralorne Resources Ltd.; Acheron Mines Ltd.; Advent Developments Inc.

Metals. No showings reported

Description. Claim area underlain by Besa River Formation

Work done. Soil geochemistry

Status. Lapsed

References. British Columbia Ministry of Energy, Mines and Petroleum Resources, Annual Reports, 1972, p. 463; 1973, p. 387

Linda, Ace

Location. 94 B/12E; lat. 56° 42' - 41'
long. 123° 44' - 42'

Owner. Bernard J. Marini

Operator. British Newfoundland Exploration Ltd., Vancouver

Metals. Zinc

Description. LINDA claims are underlain by brown siltstone unit; ACE claims are underlain by Muncho-McConnell and Stone formations. Minor sphalerite occurs in fractures in calcareous shale of LINDA claims; limonitic gossan was found on ACE claims

Work done. Surface mapping; one diamond-drill hole totalling 183 feet (56 m) on ACE claims

Status. Lapsed

References. British Columbia Ministry of Energy, Mines and Petroleum Resources, Annual Reports, 1974, p. 289; 1975, p. E156

Ospika

Location. 94 B/5W; lat. 56° 21'
long. 123° 46'

11 km east of Ospika River, 2 km north of Gavreau Creek at approximately 1830 m elevation

Owner. J.R. Woodcock

Operator. Aquitaine Co. of Canada Ltd., Calgary

Metals. Zinc

Description. Sphalerite and smithsonite in fractured dolostone of the Skoki Formation

Work done. Silt and soil geochemistry

Status. Lapsed(?)

References. British Columbia Ministry of Energy, Mines and Petroleum Resources, Annual Report, 1975, p. E154

Cyn

Location. 93 B/5E; lat. 56° 16'
long. 123° 31'

Headwaters of most northerly branch of the West Nabesche River

Owner. British Newfoundland Exploration Ltd., Vancouver

Metals. Zinc

Description. Sphalerite in shaly limestones of the Besa River Formation

Work done. Surface geological mapping; silt and soil geochemistry

Status. Lapsed

References. British Columbia Ministry of Energy, Mines and Petroleum Resources, Annual Report, 1975, p. E155

Bob

Location. 94 B/13E; lat. 56° 45.6' - 46.5'
long. 123° 37.6' - 39.8'

11 km east of Mt. Lady Laurier

Owner. Anchor Mines Ltd., Vancouver

Metals. No showings reported

Description. Claim group underlain by Besa River Formation

Work done. Silt and soil geochemistry

Status. Lapsed

References. British Columbia Ministry of Energy, Mines and Petroleum Resources, Annual Report, 1973, p. 388

Van, Doll, Last, Fine

Location. 94 B/13E; lat. 56° 52.5' - 58.0'
long. 123° 30' - 38'

14 km east of Robb Lake

Owner. T. Rolston

Operator. T. & C. Management Ltd., Vancouver; Texal Development Ltd., Vancouver; Gold River Mines Ltd., Vancouver

Metals. No showings reported

Description. Claims underlain by Prophet and Besa River formations

Work done. Airborne magnetometer survey; VLF EM surveys

Status. Lapsed

References. British Columbia Ministry of Energy, Mines and Petroleum Resources, Annual Report, 1973, p. 388

Ron

Location. 94 B/13E; lat. 56° 59'
long. 123° 41'

10 km northeast of Robb Lake at 5000 feet (1525 m) elevation

Owner. Buckhorn Mines Ltd., Vancouver

Metals. Zinc, lead

Description.	Brecciated dolostones in Muncho-McConnell Formation containing fine sphalerite and galena
Work done.	Surface geological mapping; soil geochemistry; four diamond-drill holes totalling 815 feet (248 m)
Status.	Lapsed
References.	British Columbia Ministry of Energy, Mines and Petroleum Resources, Annual Reports, 1973, p. 389; 1974, p. 289; 1975, p. 156

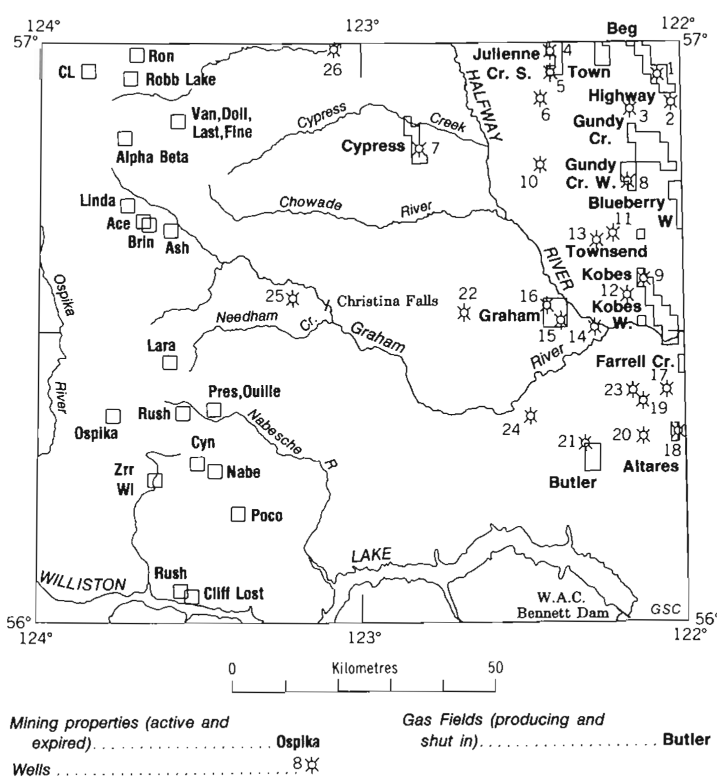
Alpha, Beta

Location.	94 B/13; lat. 56° 49.1' - 52.1' long. 123° 40.8' - 45.5'
	Between 4600 and 7600 feet (1402 and 2318 m) elevation on Calnan Creek, 6 miles (9 km) southwest of junction of Calnan Creek with Halfway River
Owner.	Milestone Mines Ltd.; Pan Ocean Oil Ltd., Calgary
Metals.	No showings reported
Description.	Claims underlain by Skoki Formation, graptolitic shale and quartzite map unit, carbonaceous limestone unit, Silurian breccia unit and brown siltstone unit
Work done.	Soil and silt geochemistry
Status.	Lapsed
References.	British Columbia Ministry of Energy, Mines and Petroleum Resources, Annual Report, 1972, p. 476

CARBONATITE-RELATED OCCURRENCES

Introduction

Cominco geologists discovered the Aley carbonatite occurrence in 1982 (Fig. 79). The following description is based on discussions with K.R. Pride (Cominco Ltd. geologist), a company report, and a description by Pell (in press). A thesis study is being carried out by U.K. Mader at the University of British Columbia under the direction of H. Greenwood.



REFERENCE TO WELLS

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2. Sundance N. Blueberry 94-B-16 a-73-H
3. Can. Hunter et al. Gundy 94-B-16 a-63-G
4. Can. Hunter et al. S. Julienne b-82-L
5. Can. Hunter S. Julienne 94-B-16 d-33-L
6. Blair "A" No. 1 94-B-16 85-E
7. HB Cypress a-86-c
8. Coseka et al. Gundy 94-B-16 b-13-B
9. Pacific Townsend a-20-H
10. (Phillips) Diaber "A" 94-B-16 a-45-D
11. Texaco NFA Townsend a-8-J
12. Can. Hunter W. Kobes c-74-B
13. Mesa et al. Aikman a-92-F
14. Lamar Hunt Soc. Aikman b-2-C
15. Soc. et al. Aikman b-21-D
16. Sinclair Can. Atlantic
17. Dome Altares d-75-H
18. Imperial Calvan a-83-A
19. Mesa et al. Altares c-61-G
20. CPOG et al. Altares a-81-B
21. Gulf POOC Butler a-65-C
22. Elf et al. Horseshoe c-45-B
23. Mesa CPOPG Altares c-72-G
24. Placid Federal a-29-E
25. Shell Hudsons Bay Schooler 94-B-11 b-77-B 1967
26. HB Headstone Creek d-87-I

Figure 79. *Location of gas fields (producing and shut-in; open squares and rectangles, respectively), mining properties (active and expired; hachured squares), and wells (circles with radiating lines) used in structure section preparation. See text for details on gas fields and mineral exploration properties.*

Geological framework

The Aley carbonatite intrudes the Kechika Group. The intrusion is circular in plan and about 2500 m across. Amphibolite rims a dolomitic (rauhaugite) core, and the surrounding stratigraphy is altered (fenitized) up to 1000 m from the intrusion.

The occurrence was displaced eastward by the Burden thrust system; the host thrust sheet forms an overturned anticline.

Local geology

A generalized geological map (prepared by Cominco geologists) shows important stratigraphic and structural contacts, and the distribution of rock types within the carbonatite complex (Fig. 80). The carbonatite core consists of rauhaugite with minor sövite. The dominant mineral constituents are: dolomite, apatite, pyrite, phlogopite and calcite; accessory minerals include zircon, sphene, magnetite, monzonite and fersmite.

The amphibolite ring comprises two textural types: massive and brecciated. The massive amphibolite is made of riebeckite and arfvedsonite (sodic amphiboles) and acmite (sodic pyroxene); accessory minerals are albite, phlogopite, calcite, pyrrhotite, magnetite and chalcopyrite. The breccias consist of angular to rounded fragments of quartzite (presumably Gog), siltstone, dolomite and albitite or trondhjemite. Fragments vary in size from 1 cm to 0.5 m across but most are in the 5 to 10 cm range. Thin reaction rims of riebeckite, acmite and phlogopite accompany the rounded "milled" fragments. The breccia matrix consists of riebeckite, arfvedsonite, acmite, albite and calcite.

The fenitized country rock appears to have been mylonitized, and weathers bright orange-red. Contact with the amphibolite is sharp. Argillaceous limestones have been dolomitized; accessory alteration minerals include apatite, pyrite and magnetite. The fenite halo is cut by numerous iron-carbonatite dikes containing fluorite, barite, siderite, ankerite and pyrite.

Niobium-bearing minerals present are fersmite, pyroclore and columbite; fersmite is most common.

Age of occurrence

Age dating analyses have produced two dates: 349 ± 12 Ma on biotite from a lamprophyre dike; and 339 ± 12 Ma on coarse biotite from brecciated amphibolite. Samples were collected and mineral separates prepared by P.C. LeCouter of Cominco. Argon analyses were done at the University of British Columbia by J. Harakal.

HYDROCARBONS

There are 11 producing gas fields within the Halfway River map area (1:250 000 geological map), all located east of the Foothills structural subprovince (Fig. 80). Three exploratory wells have been drilled within the Foothills structural subprovince, none of them successful. No wells have been drilled within the Rocky Mountains structural subprovince.

Foothills exploration is severely hampered by poor quality data retrieval and high costs. The combination of steep dipping, complex, subsurface structures, and thick shale successions like the Stoddart Group, make interpretation of reflection seismic records difficult to impossible. As techniques improve, interest in Foothills exploration will increase.

Potential for gas exists along the eastern margin of the Rocky Mountains subprovince (see discussion of blind thrusts), but it is unlikely to be tested before the 1990's.

Gas fields

Gas production is from stratigraphic and/or structural traps within the Cretaceous Bluesky Formation (Fort St. John Group), Gething and Dunlevy formations (Minnes Group); the Triassic Halfway formations (Baldonnel, Charlie Lake and Liard formations); the Mississippian Debolt and Shunda formations (Prophet Formation); and the Middle Devonian Slave Point Formation (Dunedin Formation). A summary for each field follows[1].

Altares

Location.	94 B/8
Year of discovery.	1959
Reservoir.	Bluesky Formation (lower part of Fort St. John Group)[2] in sandstone
Trap type.	Structural/stratigraphic
Initial reserves.	109.9 $10^6 m^3$ raw gas

Blueberry West

Location.	94 B/9
Year of discovery.	1956, 1957
Reservoir.	Dunlevy (Minnes Group) in sandstone
Trap type.	Structural
Initial reserves.	42.3 $10^6 m^3$ raw gas

Cypress

Location.	94 B/15
Year of discovery.	1961
Reservoir.	Baldonnel Formation in carbonate
Trap type.	Structural
Initial reserves.	1104.4 $10^6 m^3$ raw gas

[1]This data is summarized from: British Columbia Ministry of Energy, Mines and Petroleum Resources, Engineering and Geological Reference Book – Oil and Gas Pools in British Columbia, 4 vol., 1979, 81, 82, 83.
[2]Equivalent stratigraphic unit using Foothills and Rocky Mountains surface terminology is given in parenthesis.

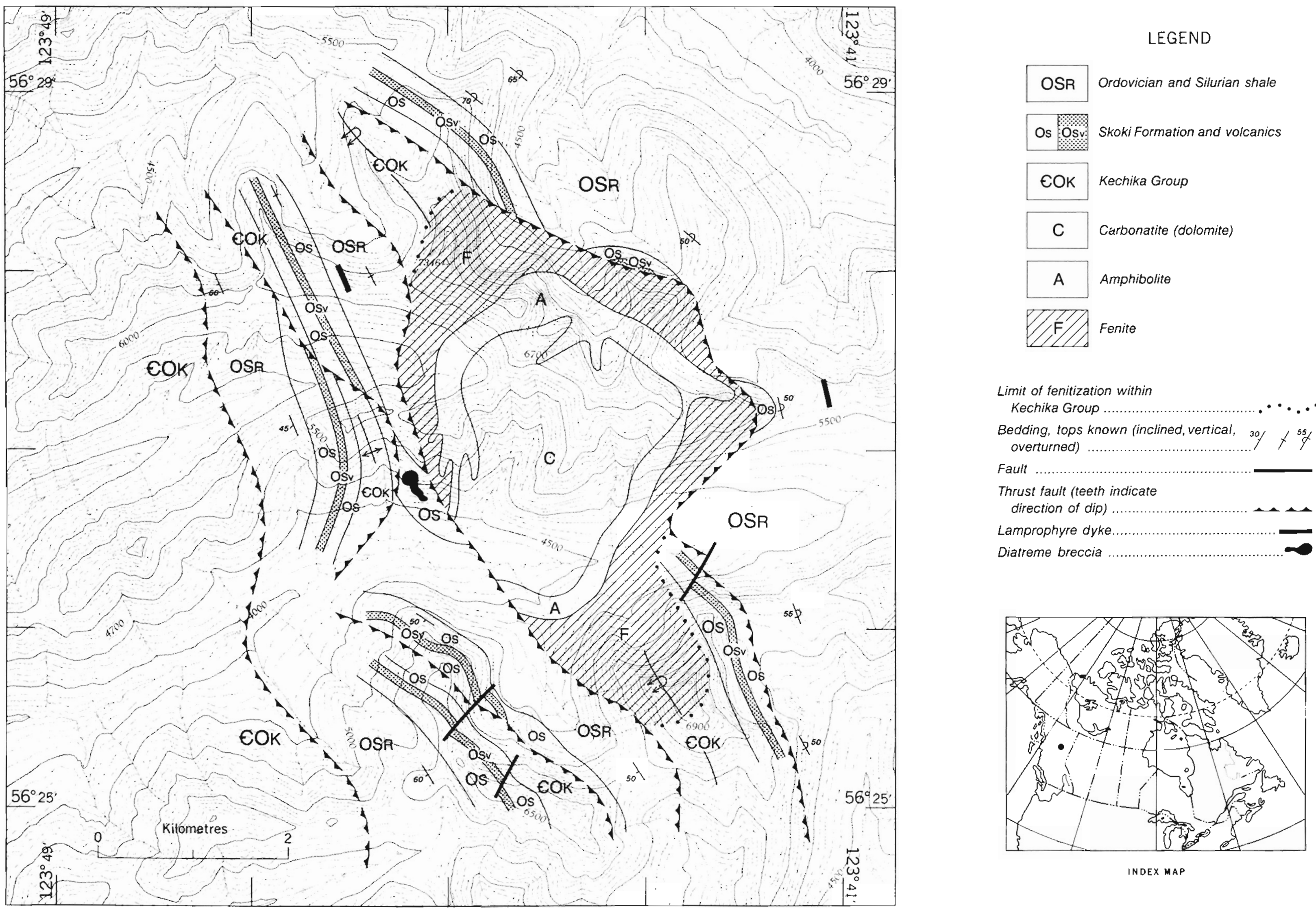

Figure 80. *Generalized geological map showing distribution of rock types within the carbonatite complex.*

Gundy Creek

Location.	94 B/16
Year of discovery.	1954, 1957, 1977
Reservoir.	Baldonnel Formation in carbonate Charlie Lake Formation, Blueberry Member in sandstone Dunlevy Formation (Minnes Group) in sandstone
Trap type.	Structural
Initial reserves.	597.8 $10^6 m^3$ raw gas

Graham

Location.	94 B/9
Year of discovery.	1973, 1977
Reservoir.	Debolt Formation (Prophet Formation), Member A in carbonate Debolt Formation (Prophet Formation), Member B in carbonate Dunlevy Formation (Minnes Group) in sandstone Gething Formation in sandstone
Trap type.	Structural/stratigraphic
Initial reserves.	930.6 $10^6 m^3$ raw gas

Gundy Creek West

Location.	94 B/16
Year of discovery.	1977
Reservoir.	Baldonnel Formation (2 pools) in carbonate Dunlevy Formation (2 pools) (Minnes Group) in carbonate
Trap type.	Structural/stratigraphic
Initial reserves.	1183.9 $10^6 m^3$ raw gas

Highway

Location.	94 B/16
Year of discovery.	1955, 1956
Reservoir.	Baldonnel Formation in carbonate Debolt Formation (Prophet Formation) in sandstone Dunlevy Formation (Minnes Group) in sandstone
Trap type.	Structural
Initial reserves.	374.9 $10^6 m^3$ raw gas

Julienne Creek South

Location.	94 B/16
Reservoir.	Debolt Formation (Prophet Formation) in carbonate Shunda Formation (Prophet Formation) in carbonate Slave Point Formation (Dunedin Formation) in carbonate
Trap type.	Stratigraphic
Initial reserves.	1149.6 $10^6 m^3$ raw gas

Kobes-Townsend

Location.	94 B/8, B/9
Year of discovery.	1956, 1957, 1958, 1977
Reservoir.	Belly Formation (Baldonnel Formation) in carbonate Charlie Lake Formation (3 pools) in sandstone Debolt Formation (Prophet Formation) in carbonate Dunlevy Formation (Minnes Group) in sandstone Halfway Formation (Liard Formation; 2 pools) in sandstone
Trap type.	Structural
Initial reserves.	4920.0 $10^6 m^3$ raw gas

Town

Location.	94 B/16
Year of discovery.	1975, 1976
Reservoir.	Baldonnel Formation in carbonate Halfway Formation, 2 pools (Liard Formation) in sandstone
Trap type.	Structural/stratigraphic
Initial reserves.	491.9 $10^6 m^3$ raw gas

Exploration potential

Anticlines and folded thrust panels of Mississippian Prophet carbonate strata are potential gas reservoirs beneath the central and eastern part of the Foothills structural subprovince. The size and shape of such structures should correspond roughly to the size of surface anticlines involving Triassic strata.

Significant potential for Middle Devonian gas reservoirs exists along a 10 km wide strip at the western margin of the

Foothills subprovince (Structure Cross-sections 1-9). Stratigraphic and structural analysis predicts the presence of folded thrust sheets containing Dunedin barrier facies at depths of 4000 m or less.

Gas potential within the Rocky Mountains structural subprovince has not been evaluated. Paleotemperatures from bitumen analyses (Macqueen and Thompson, 1978) suggest that Stone and older strata were too deeply buried (too hot) to retain gas. Primary reservoir characteristics of the Paleozoic shelf carbonates exposed at the surface are not good, and there is no reason to expect them to be significantly better within the subsurface.

COAL

Major coal resources of northeastern British Columbia occur within the Bullhead and Fort St. John groups. Most potential for commercially viable deposits exists between Smoky and Peace rivers.

Coal potential of the Halfway River map area is restricted to thin seams within the Gething Formation. Many seams were exposed[1] along the Peace River Canyon, but few exceeded 1.5 m in thickness; low- to medium-volatile bituminous coal was mined at Peace River (McLearn and Kindle, 1950). North of the Canyon, exposures are poor and potential for thick seams low. "A structurally thickened seam lies above conglomerate at the south end of Pink Mountain and one 4 foot (1.3 m) seam and coaly shale occurs about 400 feet (130 m) below the top of the Gething (Formation) on the west flank (of Pink Mountain)" (Stott, 1972, p. 161).

[1]Flooding of the Peace River Canyon occurred in 1974 with completion of the W.A.C. Bennett hydroelectric dam.

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APPENDIX 1-1

KECHIKA GROUP
(Map unit ЄOk)

C-56102[1] I.O.[2] 56°14'N; 123°46'W BSN[3]	Large, inarticulate brachiopod trilobite fragments Age: possibly Ordovician

Remarks:
Fossils are extremely rare in the Kechika Group, but collections from GSC locs. 39955, 39961, 51262 and 51297 (in Irish, 1970, p. 23, 24) allow suggestion of several different horizons within the Canadian. The presence of both *Ceratopea* and *Tritoechia* in 39961 indicates latest Canadian (Zone J or *Hesperonomia* Zone).

APPENDIX 1-2

SKOKI FORMATION
(Map unit Os)

C-60903 S.S.[4] 2; 413-413.5 m[5] 56°57'N; 123°50'W BSN	*Blastoidocrinus* sp. *Maclurites* sp. Age: early Middle Ordovician, Whiterock
C-60904 S.S. 2; 225-227 m BSN	*Maclurites* sp. undetermined gastropod (possibly a new genus) Age: Middle or Late Ordovician; but by stratigraphic position early Middle Ordovician (Whiterock)
C-60905 S.S. 2; 130.5- 130.6 m BSN	fragments of gastropods and trilobites indeterminate, small, straight cephalopod undetermined orthid brachiopod *Desmorthis* sp. Age: early Middle Ordovician, probably late Whiterock (*Anomalorthis* Zone) or slightly younger
C-60906 S.S. 2;	indeterminate gastropods and small straight, cephalopod
18.5-19.5 m BSN	*Diparelasma*? sp. *Hesperonomia*? sp. Age: Ordovician, probably latest Canadian, *Hesperonomia* Zone, Zone J.
C-44946 I.O. 56°57'N; 123°50W BSN	*Receptaculites* sp. *Palliseria*? sp. Age: probably early Middle Ordovician (Whiterock)
C-45050 I.O. 56°53'17"N; 123°49'28"W BSN	*Palliseria robusta* Wilson Age: early Middle Ordovician, Whiterock, probably *Anomalorthis* Zone.
C-60913 S.S. 9; 34.5-35 m 56°38'N 123°47'W BSN	echinoderm debris *Diparelasma* sp. *Orthambonites marshalli* (Wilson) *Orthidiella*? sp. *Petroria rugosa* Wilson "*Plectorthis*?" *sinuatis* Wilson Age: earliest Middle Ordovician, Whiterock, *Orthidiella* Zone
C-51614 S.S. 9; 219 m BSN	undetermined brachiopod
C-60912 S.S. 9; 581-585 m BSN	echinoderm debris indeterminate, small gastropod and small, straight cephalopod *Ceratopea* sp. *Maclurites* sp.

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[1]Geological Survey of Canada collection number.
[2]I.O. = Collected from isolated outcrop.
[3]Identification and report by B.S. Norford.
[4]S.S. = Stratigraphic Section; see Figure 14 for section location.
[5]Interval, measured from base of stratigraphic section, in which fossil collection occurs.

Age: late Early to Middle Ordovician, almost certainly early Middle Ordovician.

C-51633
S.S. 9;
641 m
BSN

sponge
indeterminate fragments of orthid brachiopod and gastropod

C-51645
S.S. 9;
763 m
BSN

indeterminate gastropod and straight cephalopod
echinoderm debris
possible sponge

C-51688
S.S. 9;
1058 m
BSN

straight cephalopod(?)
Ceratopea sp.

Age: late Early or early Middle Ordovician

C-51689
S.S. 9;
1064 m
BSN

stromatoporid(?)

Age: not diagnostic

C-51691
S.S. 9;
1091 m
BSN

indeterminate tabulate coral

Age: Middle Ordovician to Permian

C-51704
S.S. 9;
1295 m
BSN

Nyctopora (*Billingsarea*) sp.

Age: Middle Ordovician, late Llanvirn to early Caradoc.

C-45012
I.O.
56°47'30"N;
123°46'W
BSN

indeterminate gastropod and straight cephalopod
possible sponge
indeterminate tabulate coral (?)

Age: probably Middle Ordovician or younger

C-45035
I.O.
56°53'35"N;
123°47'50"W

echinoderm and brachiopod debris
inarticulate and orthid brachiopods
pliomerid trilobite
Receptaculites sp.

Age: Ordovician

C-44966
S.S. 75TW-4
(not shown in Fig. 14)
309 m above base of Skoki Formation
56°57'N;
123°50'W
see C-60903-C-60906
BSN

indeterminate straight cephalopod

Remarks pertaining to Skoki fossils:

The Skoki Formation of the Peace River region has gross lithological similarities to, but is much thicker than, the Skoki Formation of the southern Rocky Mountains (Norford, 1969). Collections C-60903 to C-60906 indicate a latest Canadian to Whiterock age for the formation at their locality, ages analogous to those of the Skoki Formation in its type area. C-44946, C-45050, C-60912 and C-60913 would be of similar age. Davies (1966, p. 65, Pl. 16, fig. 4a, b) reported *Bimuria buttsi* Cooper from what may be the same formation near South Gavreau Creek. This species would indicate post-Whiterock Middle Ordovician age (Porterfield or Wilderness). A roughly similar age can be suggested for C-51704. According to Flower (1961, p. 63) all the North American species assigned to *Nyctopora* (*Billingsarea*) are Chazyan or early Mohawkian (Marmor to Wilderness of current terminology). Thus, C-51704 would appear to be middle Middle Ordovician, consistent with the early Middle Ordovician age (C-60913), 1260 m lower in the same stratigraphic section.

APPENDIX I-3

Unnamed quartzite-dolomite unit
(Map unit Oqd)

C-56101
I.O.
56°08'N;
123°27'W
BSN[1]

echinoderm fragments
bryozoan, gastropods
Grewingkia sp.
indeterminate brachiopods
Diceromyonia cf. *D. tersa* (Sardeson)
Glyptorthis cf. *G. pulchra* Wang
Oepikina sp.
Plaesiomys sp.
Platystrophia sp.
Rhynchotrema sp.
Strophomena sp.
Thaerodonta cf. *T. saxea* (Sardeson)
Anazyga sp.
indeterminate trilobite
Calliops sp.

Age: late Middle Ordovician, late Caradoc, coeval with C-60924

[1]Identification and report by B.S. Norford.

C-56106 I.O. 56°01'N; 123°32'W BSN	echinoderm and trilobite fragments bryozoan *Diceromyonia* cf. *D. tersa* (Sardeson) *Glyptorthis* cf. *G. pulchra* Wang *Lodorthis*? sp. *Oepikina* sp. *Thaerodonta*? sp. Age: late Middle Ordovician, late Caradoc

C-60924 S.S. 15; 217-218 m 56°11'N; 123°27'W BSN	echinoderm fragments bryozoans favositid coral *Grewingkia* sp. *Diceromyonia* cf. *D. tersa* (Sardeson) *Glyptorthis* cf. *G. pulchra* Wang *Oepikina* sp. *Plaesiomys* sp. *Platystrophia* sp. *Rhynchotrema* sp. *Thaerodonta* sp. harpid trilobite(?) Age: late Middle Ordovician, late Caradoc, coeval with C-56101

Remarks:

The unnamed unit yields a diverse fauna that can be correlated with faunas from low in the Maquoketa Formation of Iowa (Wang, 1949) and in the Saturday Mountain Formation of Idaho (Ross, 1959). The age is approximately late Caradoc and coeval with some of the zones in the unnamed quartzite and shale unit, thus indicating a facies change between the two units.

None of the present material indicates the Late Ordovician (Ashgill) fauna that is known from the Beaverfoot Formation to the south, but some of the collections made by Irish (1970, p. 24, 25) do indicate this fauna within the map area. It is approximately coeval with the *D. complanatus ornatus* Zone.

C-56085 S.S. 16; 4675 m BSN	echinoderm debris bryozan streptelasmid coral *Diceromyonia* cf. *D. tersa* (Sardeson) *Plaesiomys*? sp. *Platystrophia* sp. *Rhynchotrema*? sp. *Thaerodonta*? sp. Age: late Middle or Late Ordovician, late Caradoc or Ashgill.

C-56086 S.S. 16; 488 m 56°08'N 123°36'W BSN	echinoderm debris indeterminate solitary coral *Clorinda* sp. indeterminate rhynchonellid and other brachiopods Age: Silurian to Middle Devonian

APPENDIX 1-4

Nonda Formation
(Map unit SN)

C-60925 S.S. 15; 21.0-21.4 m 56°11'N 123°27'W BSN[1]	echinoderm debris indeterminate gastropod and brachiopod indeterminate solitary and tabulate corals *Alispira*? sp. Age: probably Early Silurian

C-60926 S.S. 15; 26-26.5 m BSN	*Catenipora* sp. *Favosites*? sp. *Streptelasma*? sp. undetermined solitary coral Age: Late Ordovician to Late Silurian, probably Silurian.

C-60927 S.S. 15; 117-117.5 m BSN	echinoderm debris *Streptelasma* sp. Age: Late Middle Ordovician to Early Silurian, Caradoc to Llandovery.

C-60928 S.S. 15; 304-304.5 m BSN	*Cystihalysites* sp. *Favosites*? sp. *Halysites* sp. Age: Silurian

C-60930 S.S. 11 unassigned platy carbonates 20 m below base of Nonda Formation 56°34'N 123°35'W BSN	echinoderm debris *Cystihalysites* sp. *Favosites*? sp.

[1]Identification and report by B.S. Norford.

64489
S.S. 11
unassigned
platy
carbonates
0-9 m
below base
of Nonda
Formation
BSN

echinoderm debris
gastropod
solitary coral
Alispira sp.
Favosites? sp.
Halysites sp.
Palaeofavosites? sp.

Age: C-64489 and C-60930 Early Silurian, probably late or middle Llandovery.

64488
S.S. 11;
0.5-6 m
BSN

stromatoporoid
solitary coral
Catenipora? sp.
Cystihalysites sp.
Favosites? sp.

64487
S.S. 11;
11-14 m
BSN

solitary coral
Catenipora sp.
Halysites? sp.

64486
S.S. 11;
26.5-29 m
BSN

solitary coral
Catenipora sp.
Favosites? sp.

64485
S.S. 11;
54-57 m
BSN

Catenipora sp.
Favosites sp.
Palaeofavosites sp.
"*Syringopora*" sp.

64484
S.S. 11;
64 m
BSN

Catenipora sp.
Favosites sp.
Palaeofavosites sp.

Age: 64484 – 64488 Early Silurian, probably late or middle Llandovery, Nonda assemblage 1.

64483
S.S. 11;
134-138 m
BSN

sponge (?), chiton, clam
gastropods, stromatoporoids
undetermined phaceloid and solitary corals
Columnaria columbia Norford
Cystiphyllum sp.
Catenipora sp.
Coenites laminatus (Hall)
Coenites? sp.
Favosites, 2 spp.
Multisolenia sp.
Palaeofavosites sp.
Alispira sp.
Hesperorthis sp.
Idiospira sp.
Platystrophia sp.
Stricklandia cf. *S. lens ultima* Williams

64482
S.S. 11;
172-175 m
BSN

solitary corals
Columnaria columbia Norford
Favosites sp.
halysitid coral
Alispira sp.

Age: 64482 and 64483 late Early Silurian, late Llandovery, Nonda assemblage 2.

64481
S.S. 11;
208 m
BSN

Pentamerus? sp.

64480
S.S. 11;
253-262 m
BSN

stromatoporoid
solitary coral
Cystihalysites sp.
Favosites sp.
Halysites sp.

Age: C-64480 and C-64481 late Early Silurian, late Llandovery, Nonda assemblage 3.

64479
S.S. 11;
273-286 m
BSN

bryozoan
undetermined tabulate and solitary corals
Coenites laminatus (Hall)
Coenites rectilineatus (Simpson)
Cystihalysites sp.
Favosites, 2 spp.
Favosites? sp.
Thamnopora sp.
halysitid corals
undetermined brachiopod
Atrypa cf. *A. parva* Hume

64478
S.S. 11;
292-304 m

solitary coral
Cystihalysites sp.
Favosites?, 2 sp.
Fletcheria sp.
Halysites sp.
Halysites cf. *H. sandpilensis* Norford
Syringopora verticillata (Goldfuss)

64477
S.S. 11;
312-318 m
BSN

solitary corals
Cystihalysites magnitubus (Buehler)
Favosites, 2 spp.
Syringopora verticillata (Goldfuss)

64476 and
C-60919
S.S. 11;
325-328 m
BSN

indeterminate small gastropod
Catenipora sp.
Cystihalysites sp.
Favosites sp.
Favosites? sp.
Halysites occidens Norford
Pentamerus sp.

64475
S.S. 11;
424-425 m
BSN

stromatoporoid
solitary corals
Favosites sp.
Favosites? sp.

64474
S.S. 11;
463-471 m
BSN

solitary corals
Cystihalysites sp.
Favosites?, 3 spp.
Halysites or *Cystihalysites*, 2 spp.
Heliolites cf. *H. megastoma* (M'Coy)
Heliolites sp.
Syringopora? sp.

MEMOIR 425 – R.I. Thompson

STRATIGRAPHY, TECTONIC EVOLUTION AND STRUCTURAL ANALYSIS OF THE HALFWAY RIVER MAP AREA (94B), NORTHERN ROCKY MOUNTAINS, BRITISH COLUMBIA.

Page 103 inadvertently left off during printing, please affix to page 103.

64473 S.S. 11; 478-486 m BSN	echinoderm fragments stromatoporoid solitary corals *Cystihalysites magnitubus* (Buehler) *Cystihalysites* sp. *Favosites*, 2 spp. *Halysites* sp. *Heliolites* cf. *H. megastoma* (M'Coy) *Paleofavosites* sp. undetermined brachiopod *Pentamerus* sp.
C-60909 S.S. 11; 533.5- 534 m	echinoderm fragments and plates stromatoporid algae? indeterminate rugose and tabulate corals *Catenipora simplex* (Lambe) *Coenites* sp. *Cystihalysites* sp. *Cystiphyllum*? sp. *Favosites* sp. *Goniophyllum*? sp. (including operculum) *Halysites sandpilensis* Norford indeterminate halysitid corals *Heliolites* sp. *Romingeria* cf. *R. vannula* Davis *Thamnopora* sp. undetermined brachiopods *Clorinda*? sp. *Cryptatrypa* sp. *Hesperorthis* cf. *H. laurentia* (Billings) *Howellella* sp.
64472 S.S. 11; 544 m BSN	*Thamnopora* aff. "*Coenites*" *laqueata* Rominger
64471 S.S. 11; 558-576 m BSN	solitary corals *Cystihalysites* sp. *Cystiphyllum* sp. *Favosites*? sp.
	Fletcheria aff. *F. tubifera* Milne Edwards and Haime *Halysites* cf. *H. sandpilensis* Norford *Syringopora verticillata* Goldfuss
64470 S.S. 11; 594-595 m BSN	*Coenites* sp. *Halysites* sp. *Halysites* cf. *H. sandpilensis* Norford *Striatopora*? sp. *Thamnopora* sp. Age: 64470 – 64479, C-60909, C-60919 late Early Silurian, late Llandovery, probably Telychian, Nonda assemblage 4.
C-56087 S.S. 16; 556 m 56°08'N; 123°36'W BSN	*Cystihalysites* sp. *Favosites* sp. Age: Silurian
C-44980 S.S. 75TW-4, 5; 36 m (not shown in Fig. 14) 56°36'N; 123°33'W BSN	*Coenites* sp. Age: Silurian or Devonian
C-44981 S.S. 75TW-4, 5; 44 m (not shown in Fig. 14) BSN	indeterminate solitary and halysitid corals *Favosites*, 3 spp. *Halysites* sp. Age: Silurian

APPENDIX 1-5

Dunedin Formation
(Map unit DD)

C-46842 S.S. 3 56°56'N; 123°41'31"W MJC[1]	*Moelleritia canadensis* Copeland *Moelleritia canadensis insignis* Copeland Age: Early Devonian – Emsian Remarks: The range of *M. canadensis* in northwestern mainland and Arctic Canada is aligned with the conodont *dehiscens*, *gronbergi*, *inversus*, *serotinus* and the lower part of the *patulus* Zones, that is, Emsian of the Early Devonian. These ostracodes occur typically in upper strata of the 'Bear Rock Formation' and correlatives.

[1]Identification and report by M.J. Copeland.

C-56105 I.O. 55°59'N; 123°32'W AWN[1]	*Coenites* sp. *Spinatrypa* (*Spinatrypa*) sp. cf. *S.* (*S.*) *andersonensis* (Warren) *Spinatrypa* sp. – finely costate form cf. *Emanuella* sp. 1 of Caldwell (1968) ostracode planispiral gastropod fragments *Styliolina* sp. *Tentaculites* sp. echinoderm ossicle with single axial canal fish tooth fragment Age: Middle Devonian Remarks: The presence of *Coenites* sp., *Spinatrypa* sp. cf. *S.* (*S.*) *andersonensis* and cf. *Emanuella* sp. indicates a Middle Devonian age. These are long ranging forms that occur in both the Eifelian and Givetian of the Middle Devonian of northwestern Canada.
C-56105 (as above) AEHP[2]	*Favosites* sp. *Thamnopora*, 2 spp. Age: Early or Middle Devonian
C-44884 S.S. 12 56°25'N; 123°28'W AEHP	*Striatopora* sp. Age: Silurian to Middle Devonian
C-45015 I.O. 56°50'N; 123°39'30"W AEHP	*Syringopora* sp. undet. Age: Silurian to Permian
C-52031 I.O. 56°41'24"N; 123°42'06"W AEHP	*Squameofavosites* sp. *Mackenziephyllum* sp. nov. *Cyttaroplasma* sp. nov. *Xystriphyllum* sp. *Sociophyllum* sp. *Warrenella* sp. Age: Middle Devonian, early Eifelian Remarks: *Mackenziephyllum* is known from the Hume and Harrogate formations (both Eifelian) of Western Canada and from Eifelian strata in the Urals. *Cyttaroplasma* occurs as early as Zlichovian at Royal Creek. This suggests a Lower Devonian age, but *Sociophyllum* and *Mackenziephyllum* are confined to Eifelian strata in western North America.
C-52031 (as above) TTU[3]	*Icriodus culicellus* (Bultynck) (small form) *Polygnathus parawebbi* Chatterton *Belodella* spp. Age: early Couvinian (Eifelian) Remarks: A similar fauna has been found previously in the *pedderi-parawebbi* faunal unit, from about the middle part of the Hume Formation at Powell Creek, western District of Mackenzie (Uyeno, 1979, p. 236). Klapper and Ziegler (1979, Fig. 4) have equated this fauna with the *australis* Zone.

APPENDIX 1-6

Road River strata undivided
(Map unit OSDR)

C-56103 I.O. 56°14'N; 123°46'W AEHP	stromatoporid, not studied *Cladopora* (s.s.) sp. undet. Age: Silurian to Middle Devonian; probably Late Silurian or Early Devonian.
C-56104 I.O. 56°25'N; 123°45'30"W BSN[4]	graptolite fragments *Orthograptus*? sp. Age: Middle Ordovician to Early Silurian

[1]Identification and report by A.W. Norris
[2]Identification and report by A.E.H. Pedder.
[3]Identification and report by T.T. Uyeno.
[4]Identification and report by B.S. Norford.

APPENDIX 1-7

Graptolitic shale-quartzite unit
(Map unit Osq)

C-45600
S.S. 7;
126.5 m
56°47'30"N;
123°44'W
BSN[1]

Climacograptus sp.
C. cf. *C. caudatus* Lapworth
Dicellograptus ex gr. *D. morrisi*; Hopkinson
Glyptograptus? sp.
Leptograptus sp.
Orthograptus? sp.

Age: late Middle Ordovician, late Caradoc, probably *Orthograptus quadrimucronatus* Zone

C-45599
S.S. 7;
246.5 m
BSN

Dicellograptus sp.
Orthograptus? sp.

Age: Middle or Late Ordovician, Caradoc to Ashgill

C-60918
I.O.
56°48'N;
123°45'W
BSN

Climacograptus sp.
Dicellograptus cf. *D. complanatus* Lapworth
Diplograptus? sp.
Orthograptus sp.

Age: Late Ordovician, Ashgill, *Dicellograptus complanatus ornatus* Zone

C-51808
I.O.
56°48'30"N;
123°45'W
BSN

Diplograptus? sp.
Glossograptus? sp.

Age: Ordovician

C-60915
I.O.
56°48'30"N;
123°45'W
BSN

Dicellograptus sp.
Diplograptus? sp.
Glossograptus? sp.
Orthograptus sp.

Age: late Middle Ordovician, Caradoc

C-51689
S.S. 9;
192 m
56°38'N;
123°47'W
BSN

stromatoporid(?)

C-51691
S.S. 9;
219 m
BSN

indeterminate tabulate coral

Age: Middle Ordovician to Permian

C-60911
S.S. 9;
0-10 m float
BSN
56°38'N;
123°46'30"W
BSN

inarticulate brachiopod
trilobite fragment
Orthograptus? sp.

Age: Middle Ordovician to Early Silurian

C-56005
S.S. 6;
162-163 m
56°50'N;
123°46'W
BSN

graptolite fragments
Climacograptus? sp.
Orthograptus? sp.

Age: Middle Ordovician to Early Silurian

C-56006
S.S. 6;
169.5-170 m
BSN

Climacograptus sp.
Dicellograptus? sp.
Dicranograptus? sp.

Age: Middle or Late Ordovician

C-56007
S.S. 6;
172-173 m
BSN

Climacograptus? sp.

Age: Early Ordovician to Early Silurian

C-56008
S.S. 6;
162-163 m
BSN

diplograptid graptolite

Age: Middle Ordovician to Early Silurian

C-56009
S.S. 6;
162-163 m
BSN

diplograptid graptolite

Age: Middle Ordovician to Early Silurian

C-56010
S.S. 6;
182-183 m
BSN

Climacograptus ex gr. *C. bicornis* (Hall)
Orthograptus? sp.

Age: late Middle Ordovician, Caradoc, probably *Climacograptus bicornis* Zone to *Orthograptus quadrimucronatus* Zone.

Remarks:
Beds within the graptolitic shale-quartzite unit are dated as probably *O. quadrimucronatus* Zone (C-45600), and *D. complanatus ornatus* Zone (C-60918). Davies (1966, p. 21, 54) did not recognize the latter zone in the unit within the Halfway River map area, but did document the *quadrimucronatus* Zone. Demonstrated ages of the unit thus include both late Caradoc and Ashgill horizons.

[1]Identification and report by B.S. Norford.

APPENDIX 1-8

Carbonaceous limestone unit
(Map unit Scl)

C-52016
I.O.
56°43'30"N;
123°44'50"W
BSN[1]

indeterminate gastropod and solitary coral
Catenipora? sp.
Favosites? sp.

Age: probably Silurian

APPENDIX 1-9

Brown siltstone unit
(Map unit SDbs)

C-52005
I.O.
56°43'10"N;
123°42'30"W
AWN[2]

Anastrophia sp.

Age: Silurian to Early Devonian

Remarks:
Samples from Geological Survey of Canada localities C-52005 and C-51777 contain a pentamerid brachiopod suggestive of the genus *Anastrophia* which ranges from the Silurian to Early Devonian. Another form suggestive of the genus *Spirigerina* in the sample from Geological Survey of Canada locality C-51777 also has approximately the same range.

C-51777
as for
C-52005
AWN

cf. *Anastrophia* sp.
cf. *Spirigerina* sp.

Age: Silurian to Early Devonian

Remarks:
See C-52005 above.

C-51902,
C-45011
I.O.
56°54'06"N;
123°56'40"W
AWN

inarticulate brachiopod(?) fragments
echinoderm ossicle with single axial canal
Coenites? sp.
Favosites? sp.
echinoderm ossicle with five-point star shaped axial canal

Age: Silurian to Devonian

Remarks:
The genus *Coenites*, questionably present in the sample from C-45011 ranges from the Silurian to the Devonian.

C-45019,
C-45038
I.O.
56°48'20"N;
AWN

Coenites? sp.
Favosites? sp.
Gasterocoma? *bicaula* Johnson and Lane
echinoderm ossicle with single axial canal
undetermined digitate stromatoporoid
very large echinoderm ossicle with single axial canal
echinoderm ossicle with single axial canal
echinoderm ossicle with five-point star shaped axial canal

Age: late Early Devonian, probably Emsian

Remarks:
The distinctive and widely distributed 'two-holer' echinoderm ossicle (*Gasterocoma*? *bicaula* Johnson and Lane), present in the sample from Geological Survey of Canada locality C-45019, is known to range throughout the Emsian (late Early Devonian) to early Eifelian (early Middle Devonian). Where dated by associated conodonts the acme of its occurrence seems to be in beds of late Emsian (late Early Devonian) age. In northeastern British Columbia, this form is known from the lower part of the Dunedin Formation of Taylor and MacKenzie (1970).

C-45018
I.O.
56°43'N;
123°41'W
AEHP[3]

Carlinastraea sp. undet.

Age: Early Devonian, Lochkovian or Pragian

Remarks:
In the well known areas of Royal Creek, Yukon Territory and Roberts Mountains, Nevada, *Carlinastraea* is apparently confined to Lochkovian strata, but in Alaska, Europe, Asia and Australia it is known to range upward into Pragian beds.

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[1]Identification and report by B.S. Norford.
[2]Identification and report by A.W. Norris.
[3]Identification and report by A.E.H. Pedder.

C-51791
I.O.
56° 43'N;
123° 42'W
AEHP

Exilifrons sp. nov. (Geological Survey of Canada figured specimen 53073).

Age: Early Devonian, probably late Zlichovian

Remarks:
This specimen of *Exilifrons* has been figured and discussed by Pedder (1977, Geological Survey of Canada Paper 77-1B, Figs. 34.13-34.16); it is believed to indicate a probable late Zlichovian age.

C-51789,
C-51790
I.O.
56° 43'N;
123° 42'W
AEHP

ptenophyllid coral, indet.
biaxial crinoid columnals
favositid coral, poorly preserved
tentaculitid (s.l.) poorly preserved

Age: Early Devonian, Zlichovian or Dalejan

C-45016
I.O.
56° 45'30"N;
123° 39'30"W
AEHP

Favosites sp. undet.

Age: Silurian to Middle Devonian

C-45021
I.O.
56° 49'30"N;
123° 41'30"W
AEHP

Favosites sp. undet.

Age: Silurian to Middle Devonian

C-51904
I.O.
56° 43'30"N;
123° 44'W
BSN

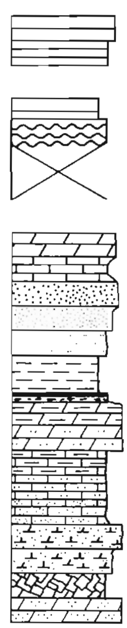
Cyrtograptus sp.
Monograptus ex gr. *M. priodon* (Bronn)

Age: latest Llandovery to Wenlock, probably latest Llandovery

APPENDIX 2

LEGEND TO ACCOMPANY APPENDIX 2

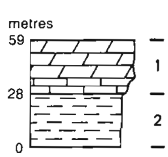
Explanation of graphic symbols



Thick bedded to massive
Very thin to medium bedded

Planar bedded
Wavy bedded
Covered interval

Dolostone
Limestone
Quartz sandstone; medium to coarse grained
Quartz sandstone; very fine to fine grained
Siltstone
Shale, argillite
Chert; massive, lenses, nodules
Argillaceous dolostone
Silty dolostone
Argillaceous limestone
Silty limestone
Sandy limestone
Dolomitic sandstone or calcareous sandstone
Dolomitic siltstone or calcareous siltstone
Intricate veining and/or crackle brecciation
Sandy dolostone



Descriptive units bounded by bold lines, and numbered on right. Cumulative unit thickness shown on left (Sections 9, 10 and 11, show cumulative thicknesses for individual formations or groups of formations.)

Unconformity ....

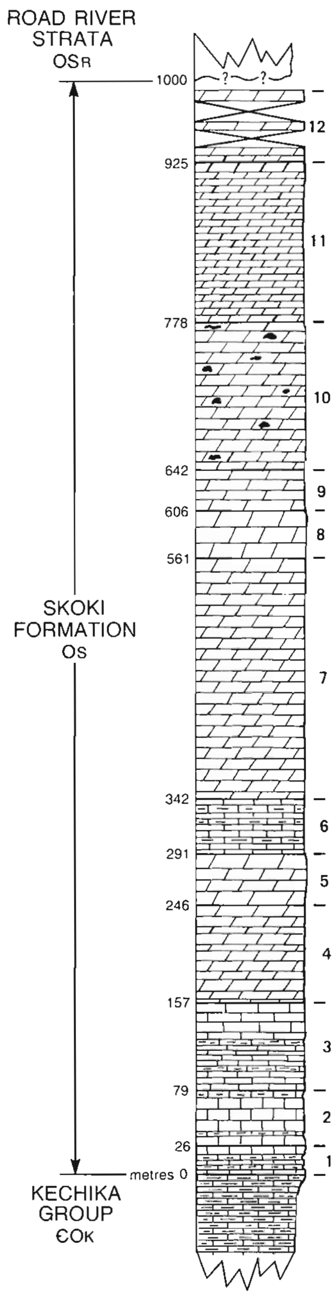
Break in Section ....

Remainder of section removed by erosion ....

GSC locality number (see Appendix 1 for description of specimen) C-60924

Location of section (see Fig. 14) 6

APPENDIX 2-1
SECTION 13
SKOKI FORMATION
56°18′N, 123°43′W



12. **Dolostone:** fine crystalline; ribs showing in covered interval; section in thrust fault contact with overlying succession

11. **Dolostone:** fine crystalline, siliceous; planar, thin bedded; oncolites; grey, weathers light grey to pink

10. **Dolostone:** fine crystalline; planar, medium bedded, irregular chert masses, oncolites; mottled; dark grey, weathers grey

9. **Dolostone:** fine crystalline; planar, medium bedded; dark grey, weathers greenish grey

8. **Dolostone:** fine crystalline, siliceous; planar, thick bedded; mottled; dark grey, weathers grey and dark grey

7. **Dolostone:** fine crystalline, siliceous; planar, thick bedded; oncolites in some beds; mottled; grey to dark grey, weathers grey and light grey

6. **Limestone** 60%: fine crystalline; planar, thin bedded, dolomite partings; dark grey, weathers grey and tan
   Argillaceous limestone 40%: planar, thin bedded; dark grey, weathers dark grey

5. **Dolostone:** fine crystalline; selective silicification; planar, thick bedded, oncolites; silicified fossil debris; dark grey, weathers grey

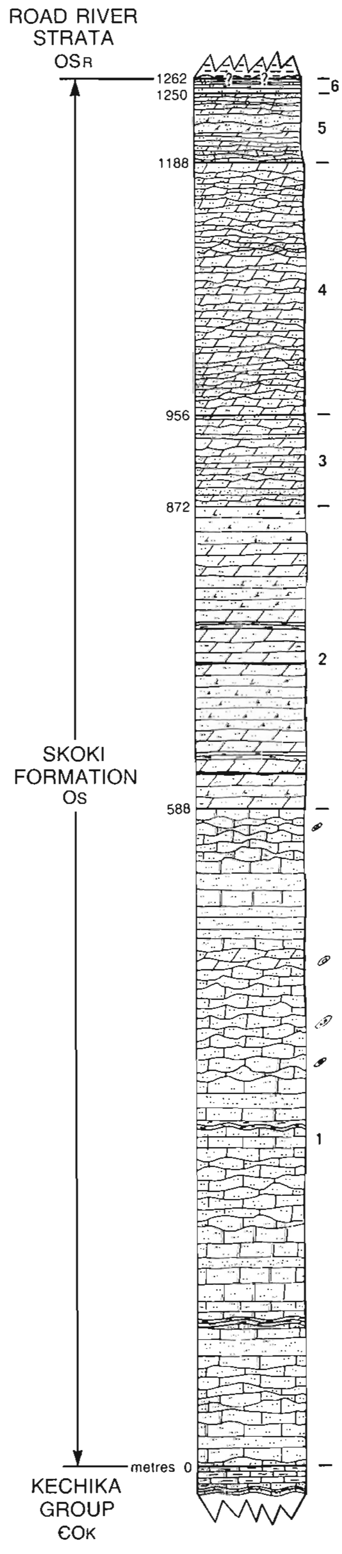
4. **Dolostone** 50%: fine crystalline; selective silicification; planar, thick bedded; dark grey, weathers light grey and tan
   Dolostone 50%: fine crystalline; planar, medium bedded; light grey, weathers tan to brown

3. **Limestone** 95%: fine crystalline; dolomite partings, planar laminated; grey, weathers grey
   Argillaceous limestone 5%: fine crystalline; cleaved; grey, weathers dark grey

2. **Limestone** 90%: fine crystalline; planar, thin bedded; grey, weathers dark grey
   Argillaceous limestone 10%: planar laminated; grey to tan, weathers brown

1. **Limestone** 50%, **calcilutite** 50%: fine crystalline; planar, thin bedded; grey, weathers dark grey
   Argillaceous limestone 50%: cleaved, tan to grey, weathers brown
   Argillaceous limestone: penetratively cleaved; tan to grey, weathers brown (Kechika Group)

APPENDIX 2-2
SECTION 9
SKOKI FORMATION
56°38′N, 123°46′W



6. **Siltstone:** sorting excellent; quartz <90%; planar laminated; tan to grey
Dolostone 80%: silty, fine crystalline; nodular, thin bedded, breccia zones; dark grey; fossiliferous, bioturbated (C-51704)

5. **Dolomitic siltstone** 15%: sorting excellent; quartz >90%; nodular, medium bedded; dark grey; oolites grading upward to oncolites
Limestone 5%: silty, fine crystalline; wavy, very thin bedded; fossiliferous, bioturbated; light grey with red zones

4. **Dolostone** 50%: silty, fine crystalline; nodular, medium bedded; dark grey; bioturbated
Dolostone 50%; silty, fine crystalline; nodular, medium bedded; light grey; occasional algal laminae and stromatolites
Dolostone < 1%: silty, fine crystalline; nodular, medium bedded; brecciated; light and dark grey interbedded (C-51691, C-51689, C-51688)

3. **Dolostone** 90%: silty, fine crystalline; nodular, very thick bedded; grey to dark grey
Dolostone 5%: silty, fine crystalline; planar, thin bedded; light grey to white
Dolomitic sandstone 3%: very fine grained; planar, medium bedded; low angle cross laminae; quartz <90%, sorting excellent; tan to grey
Dolomitic siltstone 2%: planar, medium bedded; brownish grey (C-51641)

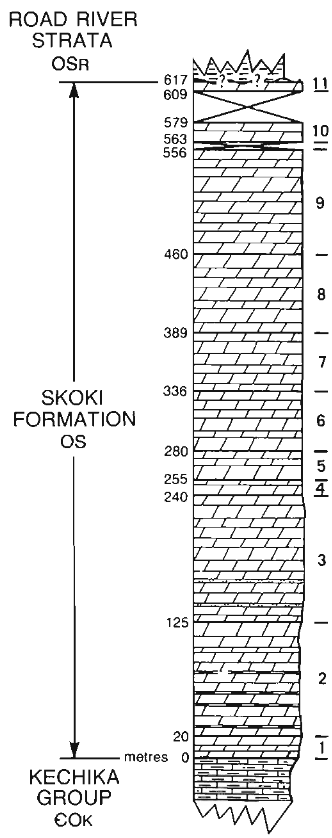
2. **Dolomitic siltstone** 45%: sorting excellent; quartz >90%; bioturbated oncolites rare; dark grey
Dolostone 25%: silty, very fine crystalline; planar, thick bedded; grey
Dolomitic siltstone 10%: sorting excellent; quartz >90%; planar, medium bedded; greenish grey
Limestone 7%: silty; chert beds, planar, thick bedded; light grey
Argillite 7%: planar, medium bedded; quartz <90%; light grey
Chert 3%: light grey to white
Dolomitic siltstone 3%: quartz <90%; sorting excellent; dark brown (C-51633)

1. **Limestone** 55%: silty, selective silicification; nodular, medium bedded; oncolites; grey, reddish brown
Limestone 20%: silty; planar, thick bedded; light grey
Siltstone 20%: sorting excellent; quartz <90%; planar, thick bedded; grey to green
Dolostone 3%: silty, selective silicification; nodular, medium bedded; grey
Limestone 1%: silty; planar, medium bedded; dark grey
Argillite 1%: wavy, very thin bedded; black

Argillaceous limestone 50%: fine crystalline; bioturbated; wavy, very thin bedded; silvery grey
Silty limestone 50%: fine crystalline; bioturbated; wavy, very thin bedded; tan, grey and orange (Kechika Group) (C-51614)

APPENDIX 2-3
SECTION 2
SKOKI FORMATION
56°57′N, 123°50′W

(measured by B.S. Norford)



11. **Dolostone:** very fine crystalline, siliceous; thin to thick bedded; dark grey, weathers light grey, blocky (C-60901)

10. **Dolostone:** very fine crystalline, siliceous; thin to thick bedded; dark grey, weathers dull olive grey (C-60902)

9. **Dolostone:** very fine crystalline, some siliceous beds; many beds with large oncolites; some pellet beds; medium to very thick bedded; dark grey, weathers grey; some mottling; large gastropods

8. **Dolostone:** very fine crystalline, some siliceous beds; mottled, oncolites, argillaceous partings; dark grey, weathers grey (C-60903)

7. **Dolostone:** very fine crystalline, siliceous; thin to thick bedded; light grey, weathers light grey, pale orange

6. **Dolostone:** microcrystalline, siliceous; thin to thick bedded; light grey, weathers light grey

5. **Dolostone:** very fine crystalline, siliceous; thin to thick bedded; grey, weathers light grey

4. **Dolostone:** very fine crystalline, siliceous; thin to thick bedded; dark grey; weathers dark grey

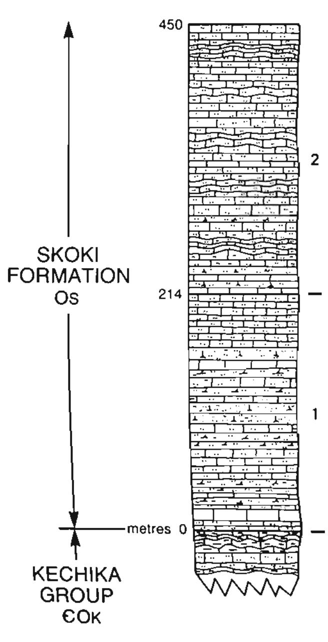
3. **Dolostone:** very fine crystalline, irregular silicification; thin to very thick bedded; quartz silt layers near base; oncolites, rare chert nodules; dark grey, weathers dark grey (C-60904)

2. **Dolostone:** fine crystalline; thin to thick bedded; nodular and rubbly dolomite with argillaceous partings over 40% of unit; dark grey, weathers dark grey (C-60905)

1. **Dolostone:** very fine crystalline, siliceous; medium to thick bedded; dark grey, weathers grey; basal contact not exposed (C-60906)

Limestone: argillaceous, very fine crystalline; bedding indistinct; dark grey, weathers light greenish grey (Kechika Group)

APPENDIX 2-4
SECTION 1
SKOKI FORMATION
56°57′N, 123°58′W



2. **Silty limestone** 40%: very fine crystalline; wavy, thin bedded, bioturbated; black with red and brown patches

Silty limestone 40%: very fine crystalline; platy, planar, thin bedded; dark grey
Calcareous siltstone 12%: planar, very thin bedded; light brown
Siltstone 5%: planar, very thin bedded; tan
Limestone 3%: very fine crystalline; planar, medium bedded; grey

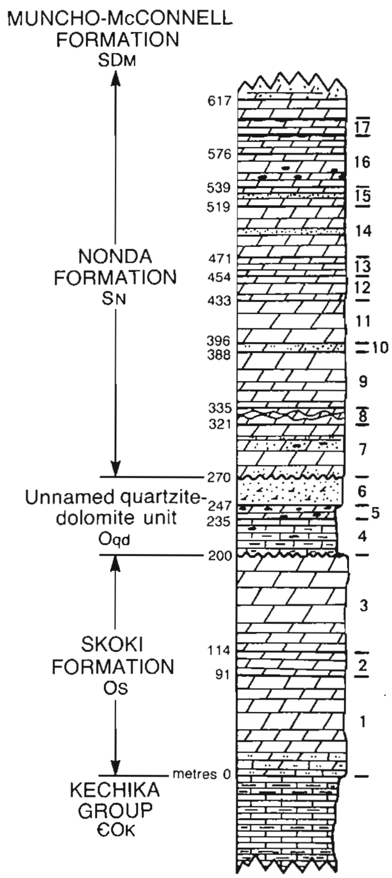
1. **Silty limestone** 50%: very fine crystalline; platy, planar, thin bedded; tan to dark grey

Dolomitic siltstone 17%: sorting excellent; quartz >90%; planar, thin bedded; grey
Calcareous siltstone 17%: sorting excellent; quartz >90%; planar, medium bedded; light grey
Limestone 1%: fine crystalline; planar, medium bedded; medium grey

Argillaceous limestone: fine crystalline; wavy, very thin bedded; silvery grey
Silty limestone: medium crystalline; wavy, thick bedded; dark grey to black (Kechika Group)

APPENDIX 2-5
SECTION 15
SKOKI FORMATION, ORDOVICIAN QUARTZITE-DOLOMITE UNIT,
NONDA FORMATION
56°11′N, 123°27′W

(measured by B.S. Norford)



17. **Dolostone:** very fine crystalline, siliceous; medium to very thick bedded; chert beds; tabulate corals; grey, weathers light grey

16. **Dolostone:** microcrystalline, siliceous; medium to very thick bedded; chert nodules; dark grey, weathers grey (C-60928)

15. **Dolostone:** very fine crystalline, siliceous; thin to very thick bedded; light grey, weathers pale orange, pale yellowish brown
Quartzite: crosslaminated

14. **Dolostone:** fine crystalline, siliceous; medium to very thick bedded; dark grey, weathers dark grey
Minor quartzite

13. **Dolostone:** very fine crystalline, siliceous; thin to very thick bedded; mottling; brownish grey, weathers light grey

12. **Dolostone:** microcrystalline, siliceous; thin to thick bedded; dark grey, weathers grey

11. **Dolostone:** microcrystalline, siliceous; medium to very thick bedded; mottled; olive grey, weathers light grey

10. **Dolostone:** fine crystalline, quartz sand and/or silt; thin to thick bedded; dark grey, weathers greyish orange

9. **Dolostone:** fine crystalline, siliceous; low to moderate porosity; thin to very thick bedded, crosslaminae; light grey, weathers light grey (C-60927)

8. **Dolostone:** microcrystalline; siliceous; thin to thick bedded; dark grey, weathers olive grey

7. **Dolostone:** microcrystalline; thin to very thick bedded; nodular chert, siliceous partings; dark grey, weathers grey, pale orange, pale brown
Sandy dolostone (basal 10 m): grey, weathers olive grey
Shale interbeds < 1%: (C-60925, C-60926)

6. **Quartzite** and **dolomitic quartz sandstone:** fine to coarse crystalline; medium to thick bedded, minor shale interbeds; grey and light brown grey, weathers brownish grey

5. **Dolostone:** microcrystalline, siliceous; thin bedded, very minor shale interbeds; black chert nodules; dark grey, weathers olive grey; *Receptaculites* sp.

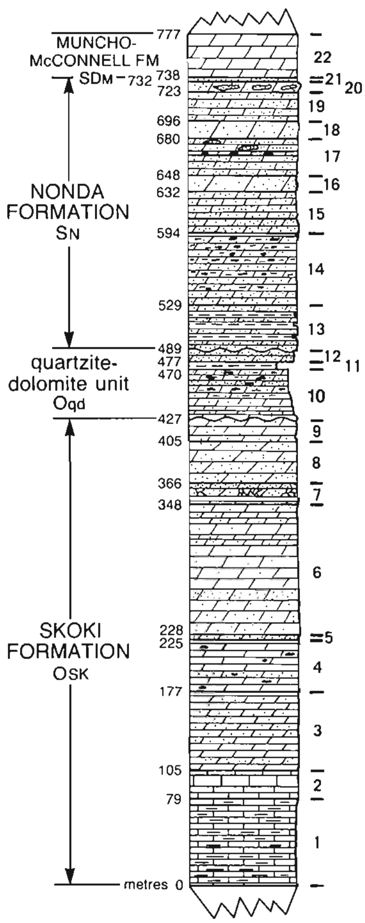
4. **Limestone:** dolomitic, fine crystalline; thin to medium bedded; nodular; dark grey, weathers yellowish grey; chert nodules (C-60924)
Calcareous shale near base: abrupt contact

3. **Dolostone:** fine crystalline, siliceous; mottled; oncolites; grey, weathers light grey

2. **Dolostone:** fine crystalline; thin to very thick bedded; dark grey, weathers olive grey

1. **Dolostone:** fine crystalline; thin to thick bedded, some beds with quartz silt layers, basal beds calcareous; oncolites; gradational lower contact; grey, weathers olive grey, pale yellowish brown (C-60922)
Limestone: fine crystalline, argillaceous; grey, weathers olive grey
Calcareous shale: cleaved; greyish black (Kechika Group)

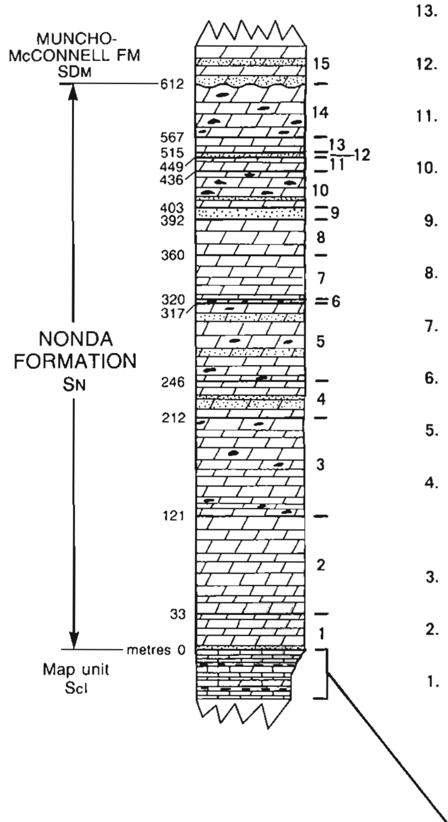
APPENDIX 2-6
SECTION 16
SKOKI FORMATION, ORDOVICIAN QUARTZITE-DOLOMITE UNIT,
NONDA FORMATION
56°08′N, 123°36′W



22. **Dolostone:** fine to medium crystalline; planar, thick bedded; light grey, weathers light grey
21. **Dolomitic sandstone:** planar, thin bedded; brown, weathers grey
20. **Sandy dolostone:** planar, very thick bedded, sandstone lenses; light grey, weathers grey
19. **Sandy dolostone:** planar, medium bedded; silicified fossil fragments; grey, weathers dark grey
18. **Sandy dolostone:** planar, very thick bedded; mottled; light grey, weathers dark grey
17. **Sandy dolostone:** planar, thin to medium bedded; chert nodules, dolomitic sandstone lenses; grey, weathers grey
16. **Sandy dolostone:** planar, very thick bedded; brown, weathers brown to dark grey
15. **Dolostone:** fine crystalline; sandy; planar, medium bedded; grey to brown, weathers dark grey
14. **Sandy dolostone** 60%: planar, medium bedded; silicified fossil fragments; grey, weathers dark grey (C 56101)
    Dolostone 40%: argillaceous, fine crystalline; planar, thin bedded; chert nodules; grey, weathers grey
13. **Sandy dolostone** 85%: black chert nodules and interbeds; planar laminated; grey, weathers dark grey (C 56086)
    Dolomitic shale 15%: planar laminated; black, weathers black
12. **Sandy dolostone:** planar, thick bedded; red, argillaceous partings; abundant brachiopod fragments; grey, weathers dark grey
11. **Shale:** fissile; black, weathers black
10. **Dolostone, sandy dolostone:** planar, thin bedded (platy); chert nodules; dark brown to grey, weathers dark grey (C-56085)
9. **Dolostone:** fine crystalline; alternating light and dark grey beds; planar, thick bedded
8. **Sandy dolostone:** fine crystalline; planar, thick bedded; light grey, weathers dark grey
7. **Sandy dolostone** 80%: fine crystalline; planar, thin bedded; veined and brecciated; grey, weathers dark grey
    Sandy dolostone 20%: planar, thin bedded, small crosslaminae; grey, weathers dark grey
6. **Sandy dolostone** 85%: fine crystalline; thin to thick, planar bedded; grey, weathers dark grey; oncolites
    Dolostone 15%: medium crystalline; thin bedded; light grey, weathers dark grey; oncolites, gastropods
5. **Dolostone:** oncolitic; sandy; medium bedded; dark grey, weathers dark grey
4. **Sandy dolostone:** medium bedded; oncolites, chert nodules; grey, weathers dark grey
3. **Dolostone:** fine to medium crystalline; medium bedded; quartz sand impurity; minor mottling; oncolites, gastropods, crinoids; grey to brown, weathers grey
2. **Limestone** 50%: medium crystalline; planar, thin to medium bedded; light grey, weathers grey
    Limestone 50%: fine crystalline; thick bedded; light grey, weathers grey
1. **Argillaceous limestone (calcilutite):** highly fractured, highly cleaved; wavy, thin to medium bedded; grey, weathers tan to grey; basal contact gradational

APPENDIX 2-7
SECTION 11
NONDA FORMATION
56°34′N, 123°35′W

(measured by B.S. Norford)



15. Dolostone, sandy dolostone: light grey, weathers light grey

14. Dolostone: fine crystalline, siliceous; very thick bedded; chert nodules, quartz sand laminae; dark grey, weathers dark grey (C-64470)

13. Dolostone: very fine crystalline; argillaceous partings, thick to very thick bedded; dark grey, weathers dark grey (C-64471, C-64472, C-60909)

12. Dolostone: fine crystalline, siliceous; sandstone layer; dark grey, weathers dark grey (C-64473, C-64474)

11. Dolostone: fine to coarse crystalline; some siliceous content; thin to very thick bedded; dark grey, weathers light grey

10. Dolostone: fine to coarse crystalline, sparse silicification; thin to very thick bedded; chert nodules, quartz sand laminae; dark grey, weathers dark grey (C-64475)

9. Quartzite: fine to coarse grained; poor sorting; very thick bedded; some crossbedding; light grey, weathers light grey

8. Dolostone: fine crystalline, siliceous; thick bedded; dark grey, weathers dark grey
Dolostone: light grey interbeds

7. Dolostone: very fine crystalline, siliceous; thick bedded; dark grey, weathers dark grey (C-64476, C-60910)

6. Dolostone: fine to medium crystalline, siliceous; medium to very thick bedded; chert nodules; dark grey, weathers grey

5. Dolostone: very fine crystalline, siliceous; medium to very thick bedded; chert nodules; some sandy beds; dark grey, weathers grey (C-64477 to C-64480)

4. Dolostone: very fine crystalline, siliceous; thin to very thick bedded; grey, weathers light grey, dark grey, olive grey
Silty and sandy dolostone: crosslaminated
Quartzite <5%: erosion surface at 246 m

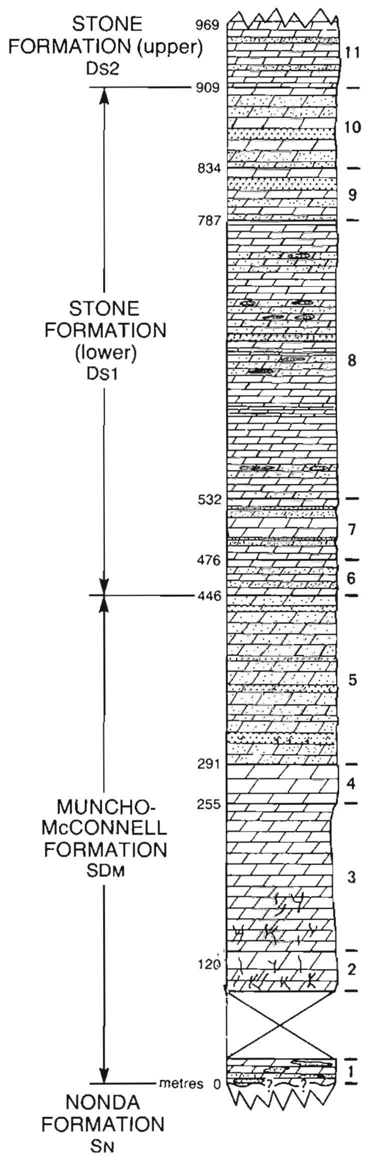
3. Dolostone: fine crystalline, siliceous; chert nodules; dark grey, weathers dark grey, olive grey; (C-64481 to C-64483)

2. Dolostone: fine crystalline; thick to very thick bedded; dark grey, weathers dark grey, olive grey (C-64484, C-64485)

1. Dolostone: fine crystalline; thin to very thick bedded; chert nodules; grey, weathers olive grey and light grey (C-64486 to C-64488)
Quartzite: very fine grained (basal 15 cm)

Dolomitic limestone, dolomite: very fine crystalline, argillaceous; very thin to thin bedded; dark grey, weathers olive grey, brown, dark yellowish orange
Calcareous shale interbeds: well cleaved (C-64489, C-60930)

APPENDIX 2-8
SECTION 14
MUNCHO-McCONNELL FORMATION, STONE FORMATION
56°21′N, 123°25′W



11. **Dolomite:** fine crystalline; planar, thin bedded, some sandy interbeds; some mottling; grey, weathers light grey

10. **Sandy dolomite** 65%: planar, medium bedded; crosslaminated with ripple marks; grey, weathers grey and tan
Dolomite 30%: fine crystalline; planar, medium bedded; dark grey, weathers grey
Sandstone 5%: fine grained; well rounded and good sphericity, well sorted; tan, weathers tan

9. **Dolostone** 50%: fine crystalline; planar, medium bedded; grey, weathers light grey to buff
Sandy dolomite 45%: planar, medium bedded; grey, weathers grey to tan
Dolomitic sandstone 5%: thick bedded; tan to grey

8. **Dolostone** 80%: fine crystalline; planar, medium bedded, with some thin bedded intervals; sandy interbeds and lenses; light grey to tan, weathers grey and tan
Dolostone 15%: fine crystalline; planar, medium bedded; dark grey, weathers dark grey
Dolomitic sandstone 5%: fine grained, good sorting, well rounded; planar, medium bedded; dark grey, weathers dark grey to brown

7. **Sandy dolostone** 65%: medium bedded; light grey, weathers tan
Dolomitic sandstone 30%: medium bedded; grey, weathers tan
Quartzite 5%: fine grained; well rounded, good sphericity; light grey, weathers light grey

6. **Sandy dolostone** 55%: planar, medium bedded; yellow to tan, weathers greyish tan
Dolostone 30%: fine crystalline; medium bedded; light grey, weathers grey
Dolostone 15%: fine crystalline; thick bedded; mottled; dark grey, weathers dark grey; basal contact abrupt

5. **Sandy dolostone** 90%: planar, thick bedded; approx. 40% quartz sand; bioturbation in some beds; light grey, weathers grey
Dolomitic sandstone 10%: planar, thick bedded, crosslaminae; grey, weathers tan and grey

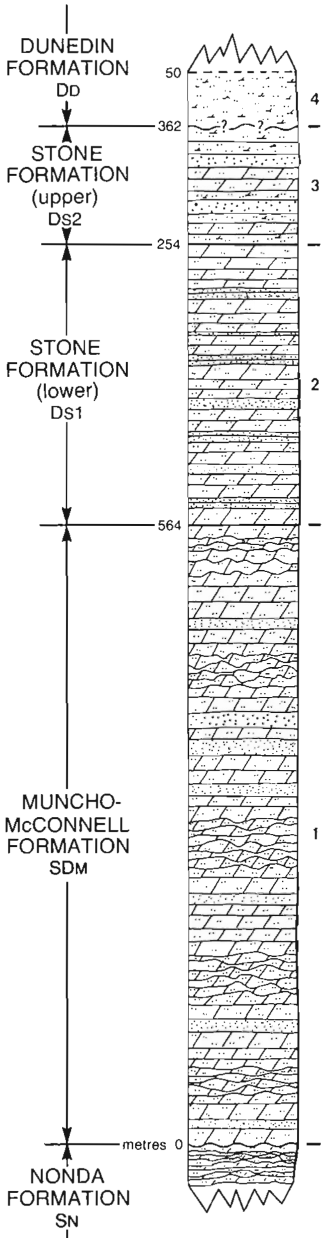
4. **Dolostone:** fine crystalline; planar, very thick bedded; light grey, weathers light grey

3. **Dolostone:** fine crystalline; planar, thick bedded; decreasing abundance of quartz stringers (40%); light grey, weathers light grey

2. **Dolostone:** fine crystalline; planar, very thick bedded; quartz stringers (up to 20% by volume); light grey, weathers grey; some bioturbation

1. **Sandy dolostone** 80%: thin bedded, crosslaminae; sandstone lenses; tan, weathers light grey
Dolostone 20%: fine crystalline; planar, thick bedded; light grey, weathers grey; basal contact abrupt
Dolostone: fine crystalline; planar, thin bedded; dark grey, weathers dark grey (Nonda Formation)

APPENDIX 2-9
SECTION 10
MUNCHO-McCONNELL FORMATION, STONE FORMATION,
DUNEDIN FORMATION
56°34′N, 12°36′W



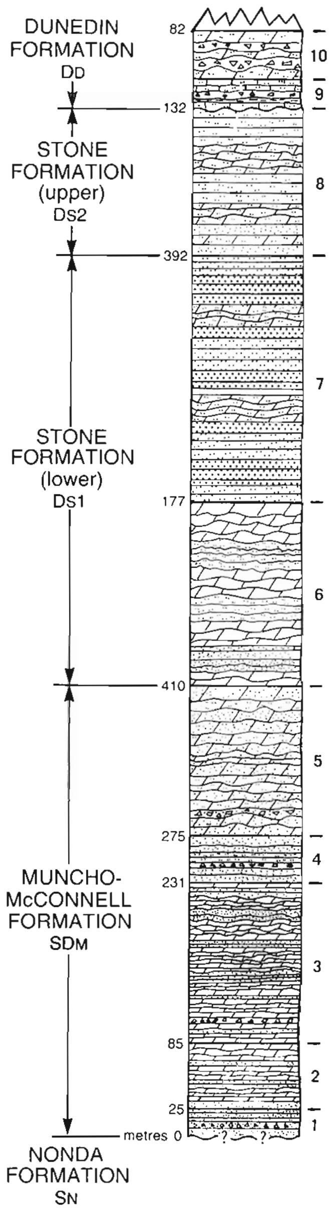
4. **Dolomitic siltstone:** brecciated, unbedded; quartz >90%; dark grey to black; dolomitized brachiopod(?) fragments

3. **Siltstone** 50%: selective silicification, dolomitic quartz >90%; sorting good; light grey to tan
Sandstone 30%: medium grained; well rounded, sphericity good; sorting good; quartz >90%; planar, thick bedded; light grey to tan
Dolostone 20%: silty, selective silicification; planar, thick bedded; light grey

2. **Dolostone** 75%: silty, selective silicification; planar, thick bedded; light grey
Sandstone 15%: very fine grained; well rounded, sphericity good; sorting good; quartz >90%; planar, very thin bedded; tan
Sandstone 10%: fine grained; planar, thin bedded; tan
Siltstone <1%: sorting excellent; planar, very thin bedded; dark grey; matrix <20%
Basal contact mixed gradational

1. **Dolostone** 40%: silty, fine crystalline; selective silicification; nodular, thick bedded; light grey
Dolostone 30%: silty, fine crystalline; selective silicification; planar, thick bedded; light grey to orange
Sandstone 15%: fine grained; well rounded, sphericity good; sorting good; quartz >90%; planar, thick bedded; tan to grey
Dolostone 10%: medium crystalline; planar, thick bedded; light grey
Sandstone 5%: medium grained; well rounded, sphericity good; sorting good; quartz >90%; planar, thick bedded; tan to grey
Dolomitic siltstone: nodular, thin bedded; dark grey; fossiliferous (Nonda Formation)

APPENDIX 2-10
SECTION 12
MUNCHO-McCONNELL FORMATION, STONE FORMATION, DUNEDIN FORMATION
56°25′N, 123°28′W



10. **Dolostone** 60%: silty; nodular, thin bedded; dark grey; fossiliferous
Breccia 40%: diamict, granule to very large pebble; angular; dark grey, white matrix

9. **Limestone** 40%: fine crystalline, silty; nodular, medium bedded; dark grey; fossiliferous
Limestone 20%: fine crystalline, silty; nodular, medium bedded; black; very fossiliferous
Shale 20%: calcareous; laminated; black; concretions
Breccia 15%: diamict; dolomite-quartz matrix; granule to very large pebble; dark grey
Limestone 5%: recrystallized-bioclastic; nodular, thin bedded
Basal contact abrupt

8. **Siltstone** 45%: sorting fair; well rounded, sphericity good; planar, medium bedded; dolomitic; grey
Dolostone 25%: silty; nodular, medium bedded; light grey
Sandstone 25%: medium grained; sorting good; well rounded, sphericity good; quartz >90%; planar, medium bedded; brownish grey
Dolostone 5%: silty; medium bedded; dark grey
Basal contact gradational

7. **Siltstone** 60%: well rounded, sphericity good; sorting good; quartz >90%; planar, medium bedded; light grey to orange
Sandstone 30%: medium grained; well rounded, sphericity good; sorting good; quartz >90%; planar, thin bedded; tan to grey
Dolostone 5%: silty; wavy, thin bedded; light grey
Sandstone 5%: fine grained; quartz >90%; well rounded, sphericity good; sorting good; planar, thin bedded; tan to grey

6. **Siltstone** 45%: well rounded, sphericity good; sorting excellent; quartz >90%; medium bedded, crinkly; light grey to orange
Dolostone 40%: silty; medium bedded, nodular; grey
Sandstone 15%: fine grained; well rounded, sphericity good; sorting good; quartz >90%; planar, thin bedded; grey to tan

5. **Dolostone** 55%: silty, selective silicification; nodular, thick bedded; light grey
Dolostone 40%: sandy, selective silicification; nodular, thick bedded; grey
Dolostone 3%: silty, selective silicification; nodular; thin bedded
Breccia 2%: diamict; poorly rounded, sphericity poor; nonbedded; grey, white matrix

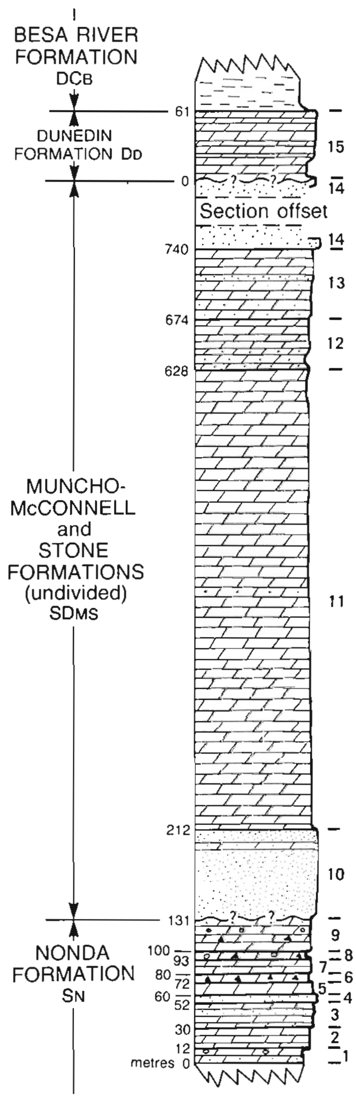
4. **Sandstone** 75%: fine grained; moderately rounded, sphericity fair; sorting good; quartz >90%; thick bedded, nodular; light grey
Siltstone 15%: sorting good; planar, thin bedded; light grey
Dolostone 5%: fine crystalline; laminated, crinkly; light grey
Breccia 5%: clasts up to 150 mm, angular to subangular; dark grey
Basal contact mixed gradational

3. **Sandstone** 35%: very fine grained; poorly rounded, sphericity poor; sorting good; quartz >90%; medium bedded, wavy; weathers grey to tan
Dolostone 35%: silty; nodular, thin bedded; white to light grey; bioturbation
Siltstone 30%: nodular, medium bedded; weathers tan to grey
Sandstone <1%: medium grained; very well rounded, sphericity excellent; sorting good; planar, medium bedded; tan to grey
Breccia <1%: sandstone clasts up to 150 mm; subangular; dolomite <20% of matrix

2. **Dolostone** 65%: silty, selective silicification; nodular, thin bedded; stromatolites; weathers light grey to tan
Sandstone 15%: fine grained; well rounded, sphericity good; sorting good; thin bedded; quartz >90%, grey
Sandstone 10%: medium grained; well rounded, sphericity good; sorting good; thin bedded; quartz >90%; crossbedding; grey, weathers tan
Dolostone 10%: silty, selective silicification; thin bedded; grey
Basal contact mixed gradational

1. **Sandstone** 40%: fine grained; well rounded, sphericity good; sorting good; quartz >90%; thin bedded; weathers light brown
Sandstone 35%: medium grained; well rounded, sphericity good; sorting good; quartz >90%; medium bedded; weathers light brown
Siltstone 25%: very fine grained; sorting good; quartz >90%; weathers light greenish grey to tan
Breccia 5%: angular fragments up to 25 mm; sandstone in dolomite matrix
Sandstone 5%: calcareous; medium grained; moderately rounded, sphericity fair; sorting fair; quartz <90%; matrix>20%; recessive weathering
Basal contact abrupt

APPENDIX 2-11
SECTIONS 3 AND 4
MUNCHO-McCONNELL AND STONE FORMATIONS (UNDIVIDED)
DUNEDIN FORMATION
123°45′N, 56°52′30″W



Shale: very thin planar laminated; black, weathers brown (Besa River Formation)

15. **Dolostone:** fine to medium crystalline; planar, medium to thick bedded; zebra texture; veined, fractured, crackle breccia, dark grey, medium grey weathering; crinoid ossicles, stromatoporoids, ostracodes at 4.5 m, fish plates at 10 m (C-46482)

14. **Orthoquartzite:** thin bedded; medium sized, rounded quartz grains; dolomite matrix ~5%; light grey, buff weathering; *Amphipora*(?) fragments

13. **Dolostone:** sandy 60%; thick bedded, thin interbeds of dolomitic quartzite, 10% rounded, medium sized quartz grains; birdseye fabric; light grey
Dolostone 30%: thick bedded; birdseye fabric; light grey
Orthoquartzite 10%: dolomite cement; thick bedded, crossbedded; very light grey; relict oolites

12. **Dolostone** 70%: thick bedded; birdseye fabric, zebra texture, relict stromatolite-like structures; medium grey, weathers light grey
Dolostone 30%: sandy; 10-80% rounded quartz; thick to medium bedded; cryptalgal laminae, stromatolite-like structures; light grey, weathers buff

11. **Dolostone** 95%: medium to fine crystalline; thick bedded to massive; cryptalgal laminae, birdseye fabric, stromatolite-like structures, zebra texture; crackle breccia, veining, stylolitic fracture surfaces; some vuggy porosity; light grey, weathers light grey; brachiopod-rich beds at 301 m
Dolostone 5%: sandy/silty; 25% quartz sand and silt; thin bedded; light grey, weathers buff

10. **Orthoquartzite** 90%: thin to thick bedded, tabular crossbeds; fine to medium sized rounded quartz grains, dolomite cement (< 10%); intraformational breccia beds 0.5-2 m thick at base; angular to rounded clasts, poorly sorted; light grey, weathers buff
Dolostone 10%: sandy; laminated to thin bedded; birdseye fabric, intraformational conglomerate; rounded, medium sized quartz grains, 10-30%
Breccia: dolostone, sandy dolostone pebbles to boulders; buff weathering proto-quartzite matrix

9. **Dolostone** 10%: silty; thin to thick bedded; 0-40% quartz silt; medium grey weathering
Dolostone 90%: fine crystalline; thin to thick, planar bedded; black chert <5%; dark grey, weathers medium grey; *Favosites*, *Halysites*, crinoid ossicles, corals

8. **Dolostone** 10%: lenticular chert; thin bedded, bioturbated; crinoid ossicles, *Halysites*, *Favosites*, pellets

7. **Dolostone:** fine crystalline; indistinct, thin bedded; medium grey weathering

6. **Dolostone:** >10% lenticular black chert; thin bedded

5. **Dolostone:** fine crystalline; thin bedded; breccia; dark grey, weathers medium grey

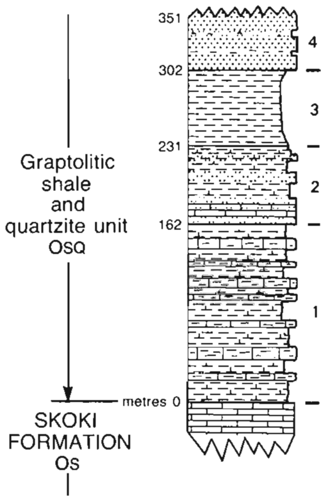
4. **Orthoquartzite:** dolomitic; tabular crossbeds; thin to thick bedded; light grey, weathers buff

3. **Dolostone** 90%: sandy; tabular crossbeds, thin bedded; coarse, rounded quartz grains up to 25%; light grey, weathers buff

2. **Dolostone:** fine crystalline; thin to thick bedded, some bioturbation; *Syringopora*, rugose corals, crinoid ossicles; medium grey

1. **Dolostone:** sandy; thin bedded to laminated, some crossbeds; crinoid ossicles, brachiopods; dark grey, weathers medium grey

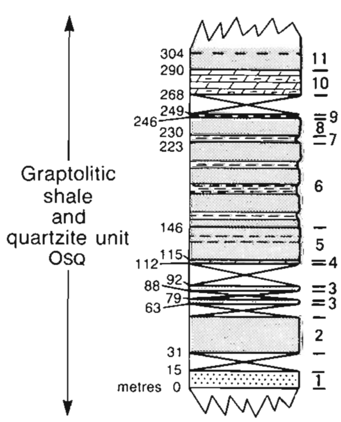
APPENDIX 2-12
SECTION 6
GRAPTOLITIC SHALE AND QUARTZITE UNIT
56°50'N, 123°46'W



4. **Quartz sandstone** 85%: fine grained; planar, thin to medium bedded; well rounded, sorting good; minor graded bedding, convoluted bedding, small-scale crosslaminae; grey, weathers tan
   Calcareous shale 10%: planar, thinly laminated; grey, weathers brown
   Calcareous siltstone 5%: highly cleaved; brown, weathers grey
3. **Shale:** noncalcareous; planar, thinly laminated (fissile); graptolitic; black
2. **Calcareous siltstone** 80%: planar laminated to thin bedded; brownish grey, weathers dark grey
   Shale 15%: slightly calcareous; planar, thinly laminated; graptolitic; black
   Argillaceous limestone 5%: planar, thin bedded; grey, weathers dark grey
1. **Argillaceous limestone** 60%: lenticular laminated; grey, weathers dark grey
   Calcareous shale 40%: planar, thinly laminated; dark grey, weathers dark grey
   Limestone (wackestone): planar, medium bedded; dark grey, weathers grey (Skoki Formation)

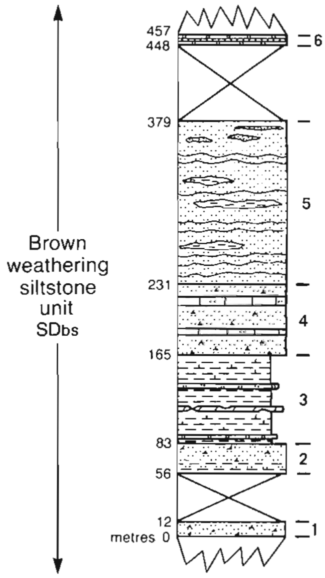
APPENDIX 2-13
SECTION 7
GRAPTOLITIC SHALE AND QUARTZITE UNIT
56°47'30"N, 123°44'W

(measured by B.S. Norford; Note: section is located within hinge of tight anticline, thicknesses may be tectonically increased)



11. **Quartzite:** dolomitic; very fine to medium grained; poorly sorted; thin to very thick bedded; brownish grey, weathers yellowish grey
10. **Dolostone:** microcrystalline, siliceous; very thin to medium bedded, nodular, recessive graptolitic shale interbeds (25%?); grey
9. **Shale:** siliceous, noncalcareous; cleaved; 20% quartzite beds; greyish black, weathers dark grey (C-45599)
8. **Quartzite:** very fine to fine grained; thin to very thick bedded, rare shale partings; brownish grey, weathers dark grey
7. **Shale:** noncalcareous; quartzite interbeds; light brownish grey, weathers dark grey
6. **Quartzite** 70%: very fine to fine grained; thin to very thick bedded; dark grey, weathers light brown; quartz stringers
   Shale 30%: noncalcareous; cleaved; dark grey, weathers dark grey
5. **Quartzite:** very fine to fine grained; sole markings; thin to medium bedded; thin shale interbeds, graptolitic; brownish grey, weathers yellowish brown (C-45600)
4. **Dolostone:** fine crystalline; siliceous; abundant shell debris; medium to thick bedded; grey, weathers light olive grey
3. **Quartzite:** very fine to fine grained; medium to thick bedded; grey, weathers yellowish brown and light olive grey
2. **Quartzite:** very fine to medium grained; medium to thick bedded; grey, weathers light brown
1. **Quartzite:** fine to medium grained; medium to thick bedded; grey, weathers light yellowish brown

APPENDIX 2-14
SECTION 8
BROWN SILTSTONE UNIT
56°48′N, 123°41′W



6. **Siltstone** 60%: planar laminated; dark grey, weathers pink and tan
   Quartz sandstone 40%: dolomitic; medium to very fine grained; chert nodules, silicified, solitary corals; single, thick beds; light grey, weathers light grey; forms mixed gradational contact with overlying, massive, dolomitic, quartz sandstone succession

5. **Siltstone** 95%: wavy, thin bedded; convoluted; tabular clasts; grey, weathers tan
   Calcareous silty shale 5%: lenticular, thin bedded; dark grey, weathers dark grey
   Quartzite partings and lenses near upper contact

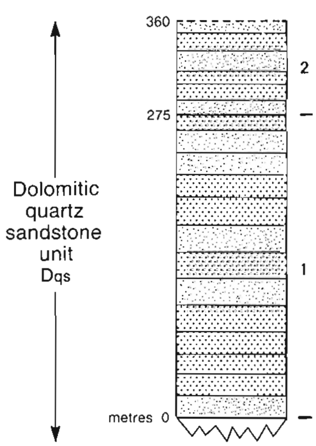
4. **Calcareous siltstone** 80%: planar laminated, occasional convolutions and crossbeds; dark grey, weathers grey
   Limestone 20%: planar laminated, convoluted in part; grey, weathers light grey and tan

3. **Shale** 80%: calcareous; planar, thinly laminated; dark grey, weathers dark grey
   Siltstone 5%: planar, thin bedded; forms resistant ribs within shale unit; grey, weathers grey
   Dolostone 5%: laminated, lenticularly bedded; forms resistant ribs in the shale unit; grey, weathers light grey

2. **Dolomitic siltstone** 80%: wavy, thin bedded; dark grey, weathers tan
   Dolomitic siltstone 15%: wavy, thin bedded; dark grey, weathers dark grey
   Shale 5%: noncalcareous; planar laminated; black, weathers black

1. **Dolomitic siltstone**: planar laminated, thin bedded; dark grey, weathers tan

APPENDIX 2-15
SECTION 5
DOLOMITIC QUARTZ SANDSTONE UNIT
56°54′N, 123°43′W



2. **Sandstone** 50%: fine grained; sorting good; well rounded, sphericity good; quartz >90%; dolomite cement; very thick bedded; light grey
   Sandstone 30%: fine grained; planar, thick bedded; light grey
   Sandstone 20%: medium grained; sorting good; well rounded, sphericity good; quartz >90%; dolomite cement; very thick bedded; light grey

1. **Sandstone** 30%: fine grained; sorting good; well rounded, sphericity good; quartz >90%; dolomite cement; very thick bedded; light grey to grey
   Sandstone 20%: medium grained; sorting good; well rounded, sphericity good; quartz >90%; dolomite cement; very thick bedded; light grey to grey
   Sandstone 20%: medium grained; sorting poor; well rounded, sphericity good; quartz <90%; dolomite cement; very thick bedded; light grey to grey
   Sandstone 20%: medium grained; sorting good; well rounded, sphericity good; quartz >90%; dolomite cement; very thick bedded; light grey to grey
   Sandstone 10%: very fine grained; sorting good; well rounded, sphericity good; quartz <90%; dolomite cement; very thick bedded; light grey to grey