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STRUCTURAL GEOLOGY OF THE CARIBOO GOLD MINING DISTRICT, EAST-CENTRAL BRITISH COLUMBIA

L.C. STRUIK

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PREFACE

Rich placer gold deposits were discovered in the Cariboo district in 1859 after news of the initial discoveries of gold in British Columbia triggered a mass migration of miners from the California goldfields. These placer deposits are now largely exhausted, after having yielded about three million ounces of gold. Although gold ore was obtained from a myriad of quartz veins and bedded pyritic deposits in Paleozoic rocks, the source of the gold in the Cariboo placer deposits remains uncertain. The present study of the structural geology may be an effective aid to locating lode gold deposits that remain to be discovered.

The geological history of the Cordillera is complex. This part of the Cordillera is interpreted as a mosaic of far-travelled terranes that were accreted to each other and to the western margin of North America as a result of the subduction of intervening lithosphere and were juxtaposed as a result of large displacements on transform faults, processes that are still continuing on the western margin of North America today. In this detailed structural analysis of the Cariboo district the late Precambrian to Mesozoic rocks are divided into four structurally and stratigraphically distinct terranes that are separated by thrust faults and strike-slip faults. The complex history of the structural stacking and disruption of the terranes is a central focus of the study.

R.A. Price
Assistant Deputy Minister
Geological Survey of Canada

PRÉFACE

C'est en 1859 que furent trouvés les graviers aurifères de la région de Cariboo, après que l'annonce des premières découvertes d'or en Colombie-Britannique eut déclenché une migration massive de mineurs de la Californie. Les placers de Cariboo sont aujourd'hui presque épuisés, ayant produit jusqu'ici environ trois millions d'onces du précieux métal. Le minerai a été extrait d'une myriade de veines de quartz et de dépôts pyritiques stratifiés qui se sont formés dans de la roche paléozoïque, mais on ne connaît pas avec certitude la source de l'or placérien de Cariboo. Cependant, l'étude structurale dont il est rendu compte ici pourrait aider à découvrir les veines qui n'ont pas encore été retracées.

La Cordillère a une histoire géologique complexe; on pense qu'elle est constituée d'une mosaïque de fragments de l'écorce terrestre ou de terrains allochtones d'origine lointaine; ces terrains se sont soudés les uns par rapport aux autres et aussi du côté ouest de la marge nord-américaine et donnant comme résultat le phénomène de subduction par l'intermédiaire de la lithosphère qui a été juxtaposée à la suite d'amples mouvements de failles transformantes, processus qui existe actuellement sur la marge ouest nord-américaine. L'auteur présente une analyse structurale détaillée des roches de la fin du Précambrien au Mésozoïque de la région aurifère de Cariboo. Il divise la région en quatre terrains structuralement et stratigraphiquement distincts et séparés par des décrochements et des failles de poussée et relate l'histoire complexe de l'assemblage et du remaniement des terrains sous l'action des forces tectoniques.

Le sous-ministre adjoint
Commission géologique du Canada
R.A. Price

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p. 56:	In the left column after the heading "Goose Peak quartzite", in the last sentence of the first paragraph, "Pine conglomerate" should read "Agnes conglomerate".
p. 59:	In the right column, the twenty-eighth line from the top, should read "2 to 150 m thick..."
p. 65:	The patterns indicating six contour intervals at the bottom of Figure 47 should accompany Figure 49.
p. 80:	In the left column, the tenth line from the bottom should read "Crenulations trend at 120°..."
p. 96:	In the last line of the caption to Figure 62, "Figure 65" should read "Figure 63".
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STRUCTURAL GEOLOGY OF THE CARIBOO GOLD MINING DISTRICT, EAST-CENTRAL BRITISH COLUMBIA

Abstract

The Cariboo gold mining district extends from Big Valley Creek in Wells map area southeast to Cariboo Lake and is entirely within the Quesnel Highlands.

Proterozoic to Jurassic rocks are bound by thrust and strike-slip faults into four tectonically and stratigraphically unique terranes. All rocks of the terranes were deposited in an ocean and vary, east to west, from continental shelf clastics and carbonates (Cariboo Terrane) through continental shelf and slope clastics, carbonates and volcaniclastics (Barkerville Terrane), and rift floor pillowed basalt and chert (Slide Mountain Terrane), to island arc volcaniclastics and fine grained clastics (Quesnel Terrane). Mesozoic shallowly-dipping faults have shortened the distance between the terranes and late Mesozoic and Tertiary moderately- to steeply-dipping (probably listric) faults have shortened, translated and extended the stacked terranes. Mesoscopically, the structure is dominated by east- and west-verging multiple folds, but regionally by shear, as expressed mainly by transpressional terrane boundary faults. The structural stacking and disruption of the terranes was accompanied by regional prograde and retrograde metamorphism which subjected most of the area to chlorite grade, and the extreme southeast to kyanite grade.

Résumé

Le district aurifère de Cariboo se trouve dans les hautes-terres de Quesnel et s'étend du ruisseau Big Valley dans la région cartographiée de Wells vers le sud-est jusqu'au lac Cariboo.

Des roches datant du Protérozoïque au Jurassique sont divisées, par des failles chevauchantes et des décrochements, en quatre terrains tectoniquement et stratigraphiquement définis. Ces roches se sont toutes accumulées dans un milieu océanique; d'est en ouest, elles changent, passant de roches clastiques et carbonatées de plate-forme continentale (terrain de Cariboo), à des roches clastiques, carbonatées et volcanoclastiques de plate-forme et de talus continentaux (terrain de Barkerville), à des basaltes en coussins et à des cherts de fond de fosse (terrain de Slide Mountain) pour se terminer par des roches volcanoclastiques et clastiques à grain fin d'arc volcanique (terrain de Quesnel). Des failles mésozoïques à faible pendage ont raccourci la distance séparant les terrains, tandis que des failles (probablement listriques) tardi-mésozoïques et tertiaires, à pendage moyen ou brusque, ont raccourci, déplacé et prolongé les terrains empilés. Du point de vue mésoscopique, la structure est dominée par des plis multiples à déversement est et ouest; du point de vue régional, elle est dominée par un cisaillement qui se traduit principalement par des failles-limites de transpression. L'empilement structural et la rupture des terrains ont été accompagnés d'un métamorphisme prograde et rétrograde régional, la plus grande partie de la région ayant atteint le faciès à chlorite et l'extrémité sud-est, le faciès à cyanite.

SUMMARY

The northern Quesnel Highlands (Fig. 1) are underlain by four geological terranes (Fig. 2), three of which are fault bounded. The terranes are defined by their unique stratigraphic successions. The easternmost is the Cariboo Terrane (Fig. 2, 3) consisting of sedimentary rocks in fault contact with the western margin of the Precambrian North American Craton along the Rocky Mountain Trench. The Barkerville Terrane consists mostly of sedimentary rocks and is west of, and in fault contact with, the Cariboo Terrane. The Barkerville and Cariboo terranes are overthrust by the Slide Mountain Terrane composed of basic volcanics and intrusives and generally fine grained clastic rocks. The root zone of the Slide Mountain Terrane is considered to be serpentinite and sheared mafic rocks that exist locally at the western boundary of the Barkerville Terrane. West of that root zone is the Quesnel Terrane composed of volcanic, volcaniclastic and fine grained clastic rocks. The oldest recognized assemblage that overlaps all of these terranes within the map area is the Pleistocene glacial deposits.

The Cariboo Terrane of the Quesnel Highlands consists of Precambrian to Permo-Triassic mainly clastic rocks. It can be divided into two successions; one Cambrian and older and the other Ordovician to Permo-Triassic. They are separated by an Ordovician unconformity. The older succession consists of grit, limestone, sandstone and shale that are laterally continuous with rocks of the Cariboo Mountains. It is the continuity of Cambrian and Precambrian stratigraphy that is the foundation of the Cariboo Terrane. Because the younger succession of basinal shale, dolostone, wacke, limestone and minor basalt (Fig. 4) is unconformable upon that foundation, it too must be part of the Cariboo Terrane. The lithologies and their ages within the younger succession correlate well with parts of the Cassiar Platform and Selwyn Basin of Yukon and northern British Columbia.

The Barkerville Terrane is dominated by Precambrian and Paleozoic varieties of grit, quartzite, and black and green pelite with lesser amounts of limestone and volcaniclastic rocks (Fig. 4). Although a stratigraphic sequence can be recognized within the terrane, it is not well understood. The Barkerville Terrane is generally more deformed and metamorphosed than any of the others. The contact between the Cariboo and Barkerville terranes is the northwest-trending, east-dipping Pleasant Valley Thrust. The Barkerville Terrane correlates well with parts of the Eagle Bay Formation of south-central British Columbia (which is included in the Kootenay Terrane with the Lardeau Group; Monger and Berg, 1984), and possibly the Horsethief Creek Group of southeast British Columbia.

SOMMAIRE

La partie nord des hautes-terres de Quesnel (fig. 1) repose sur quatre terrains géologiques (fig. 2), dont trois sont limités par des failles. Ces terrains se caractérisent par leurs successions stratigraphiques exceptionnelles. Le terrain de Cariboo (fig. 2, 3), le plus à l'est, comprend des roches sédimentaires qu'une faille a mises en contact avec la marge ouest du craton précambrien de l'Amérique du Nord le long du sillon des Rocheuses. À l'ouest du terrain de Cariboo, le terrain de Barkerville comprend surtout des roches sédimentaires; ces deux terrains ont été mis en contact par une faille, puis chevauchés par le terrain de Slide Mountain, qui comprend des roches volcaniques et intrusives et des roches clastiques généralement à grain fin. La zone d'enracinement du terrain de Slide Mountain se composerait de serpentinite et de roches mafiques cisaillées que l'on retrouve par endroits le long de la limite ouest du terrain de Barkerville. Le terrain de Quesnel se trouve à l'ouest de cette zone d'enracinement; il se compose de roches volcaniques, de roches volcanoclastiques et de roches clastiques a grain fin. Les sédiments glaciaires du Pléistocène représentent la plus ancienne association connue qui recouvre tous ces terrains dans la région cartographique.

Le terrain de Cariboo des hautes-terres de Quesnel se compose surtout de roches clastiques qui datent du Précambrien au Permo-Triassique. Il comporte deux successions, l'une datant au moins du Cambrien et l'autre de l'Ordovicien au Permo-Triassique, séparées l'une de l'autre par une discordance ordovicienne. La plus ancienne de ces successions se compose de sable grossier, de calcaire, de grès et de schiste argileux qui représentent une continuation latérale des roches des monts Cariboo. La continuité de la stratigraphie cambrienne et précambrienne forme la base du terrain de Cariboo. Puisque la succession plus récente de schiste argileux, de dolomie, de wacke, de calcaire et de petites quantités de basalte (fig. 4), accumulés dans un bassin, repose en discordance sur cette base, elle doit également faire partie du terrain de Cariboo. L'âge et la nature des lithologies de la succession plus récente peuvent être mis en corrélation avec certaines parties de la plateforme des Cassiars et du bassin de Selwyn du Yukon et du nord de la Colombie-Britannique.

Le terrain de Barkerville se compose surtout de variétés précambriennes et paléozoïques de sable grossier, de quartzite et de pélite noire et verte avec de plus petites quantités de calcaire et de roches volcanoclastiques (fig. 4). Bien qu'on reconnaisse une séquence stratigraphique dans ce terrain, elle n'est pas bien connue. Le terrain de Barkerville est généralement le plus déformé et métamorphisé des quatre terrains. Le chevauchement de Pleasant Valley, à orientation nord-ouest et à pendage est, forme le contact entre le terrain de Cariboo et celui de Barkerville. Le terrain de Barkerville peut être mis en corrélation avec certaines parties de la formation d'Eagle Bay dans la partie centrale sud de la Colombie-Britannique (cette formation fait partie du groupe de Lardeau du terrain de Kootenay; Monger, sous presse) et peut-être le groupe de Horsethief Creek dans le sud-est de la province.

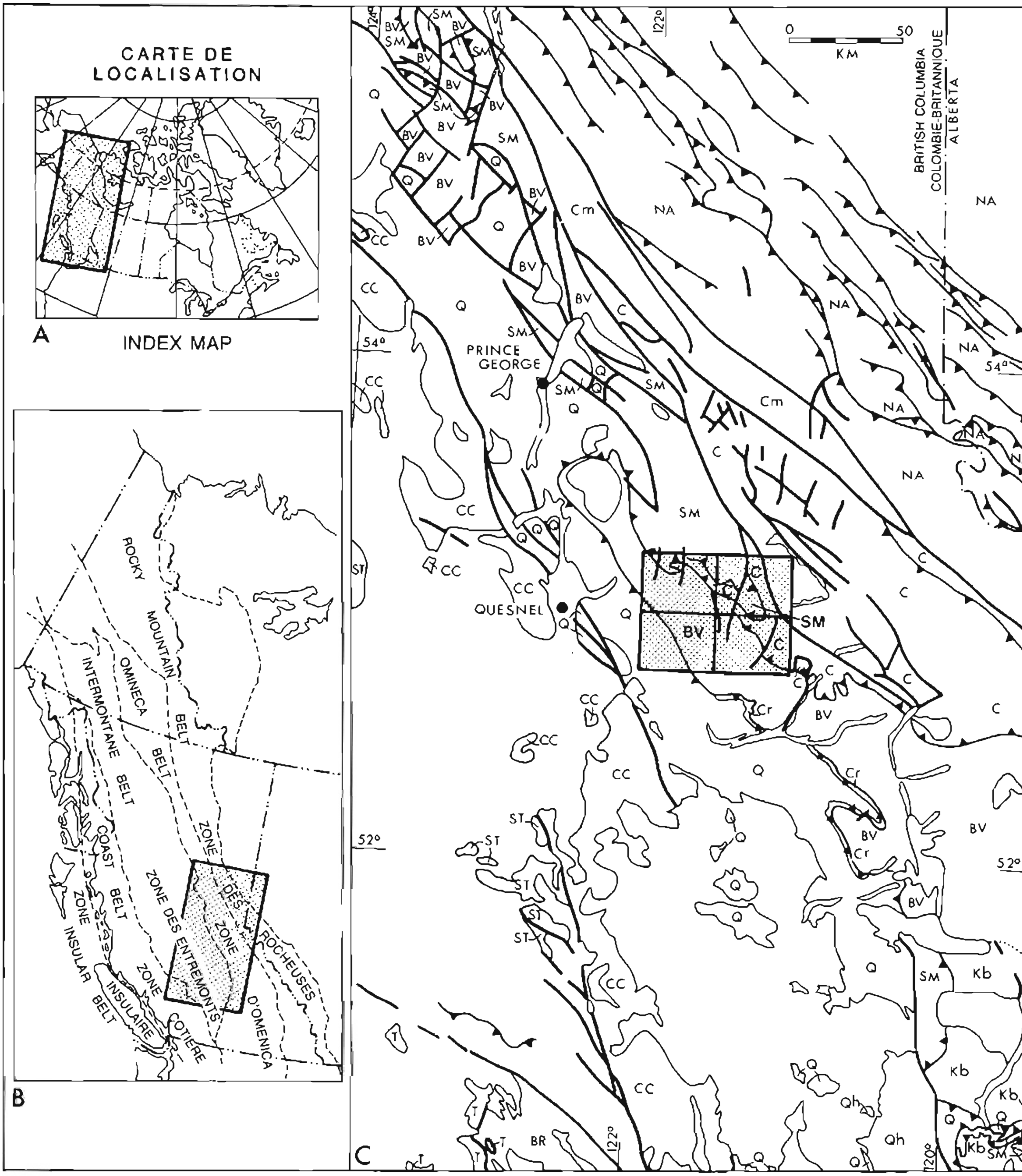

Figure 1. Regional maps telescoping the location of the map area from A) within Canada, through B) western Canada, to C) the four 1:50 000 scale maps that accompany this report shown on a regional map of Paleozoic and lower Mesozoic terranes. BR-Bridge River, BV-Barkerville, C-Cariboo, CC-Cache Creek, Cm-McGregor (subdivision of Cariboo Terrane), Kb-Kootenay (southern extension of BV), NA-North American, Q-Quesnel, Qh-Harper Ranch (subdivision of Quesnel), SM-Slide Mountain, ST-Stikinia, T-Tyaughton.

Figure 1. Cartes régionales montrant l'emplacement de la région cartographique A) à l'intérieur du Canada, B) dans l'ouest du pays et C) dans les quatre cartes à 1/50 000 qui accompagnent le présent rapport et qui paraissent sur une carte régionale des terrains du Paléozoïque et du Mésozoïque inférieur: BR-Bridge River, BV-Barkerville, C-Cariboo, CC-Cache Creek, Cm-McGregor (subdivision du terrain de Cariboo), Kb-Kootenay (prolongation méridionale du BV), NA-nord-américain, Q-Quesnel, Qh-Harper Ranch (subdivision du terrain de Quesnel), SM-Slide Mountain, ST-Stikinia, T-Tyaughton.

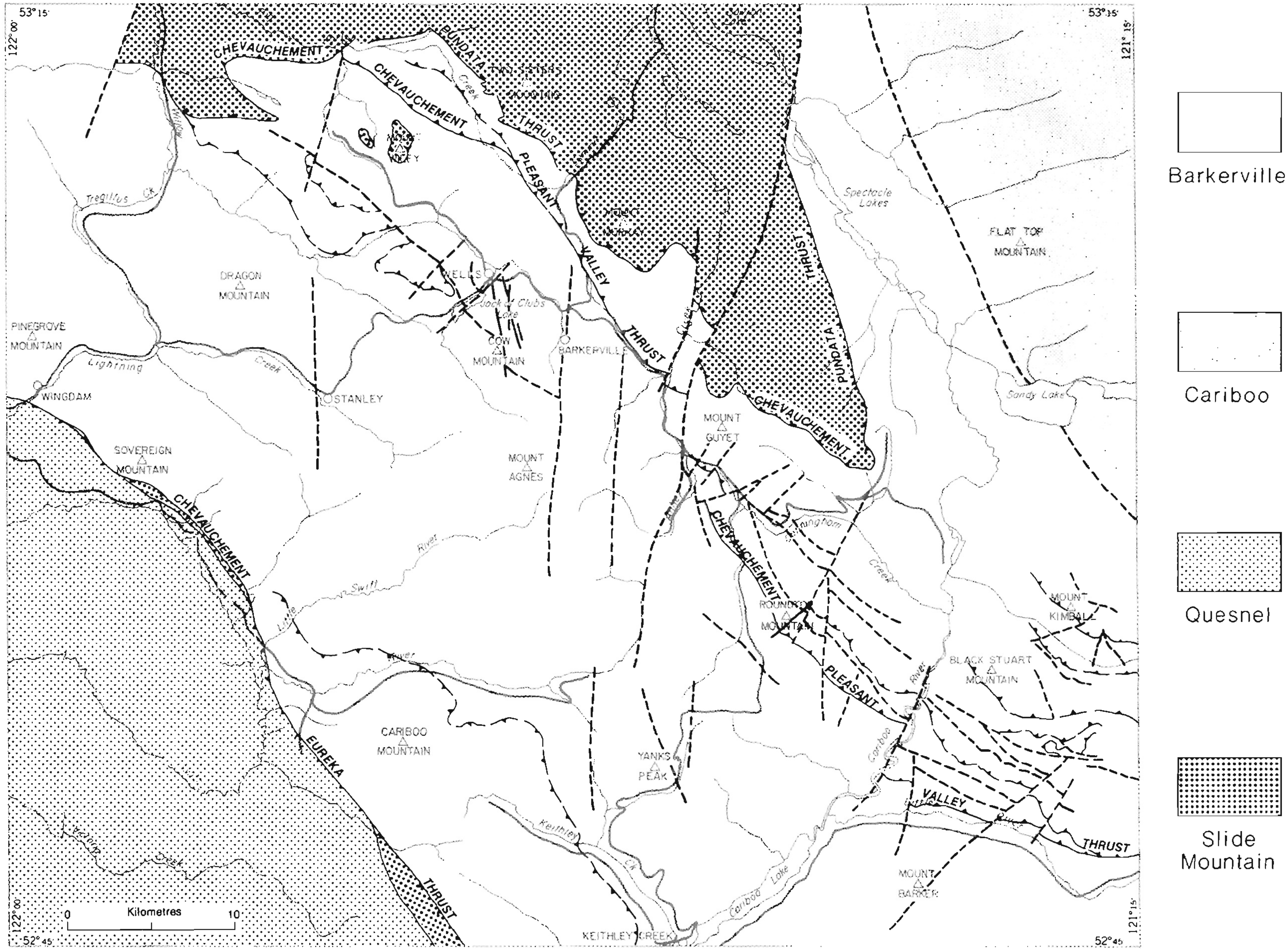

Figure 2. A map of the report area that shows the distribution of the four terranes and their bounding faults. From lightest to darkest tone the terranes are Barkerville, Cariboo, Quesnel and Slide Mountain. All contacts shown are faults.

Figure 2. Carte de la région étudiée montrant la répartition des quatre terrains et de leurs failles-limites. Les terrains sont les suivants, du moins foncé au plus foncé: Barkerville, Cariboo, Quesnel et Slide Mountain. Tous les contacts indiqués sont des failles.

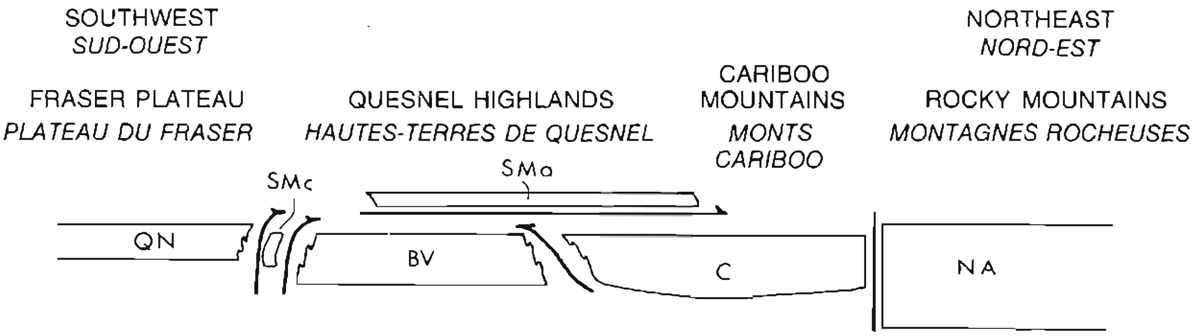

Figure 3. A cartoon structural cross-section from southwest to northeast across the map area of Figure 2 showing the relative structural position of the terranes. The terrane symbols are BV-Barkerville, C-Cariboo, SMa-Slide Mountain (Antler Formation), SMc-Slide Mountain (Crooked Amphibolite), QN-Quesnel and NA-North American.

Figure 3. Coupe transversale schématisée du sud-ouest au nord-est de la région cartographique indiquée à la figure 2, qui montre la position structurale relative des terrains. Les symboles utilisés sont les suivants: BV-Barkerville, C-Cariboo, SMa-Slide Mountain (formation d'Antler), SMc-Slide Mountain (Amphibolite Crooked), QN-Quesnel et NA-nord-américain.

The Slide Mountain Terrane consists of Mississippian to Permian basalt, in part pillowed, and chert-pelite sequences intruded by diorite, gabbro and minor ultramafic rock (Fig. 4). There are scattered localities of greywacke, grit, conglomerate and serpentinite. The terrane is internally imbricated by thrust faults. It overlies the Barkerville and Cariboo terranes on the Pundata Thrust. Included in the terrane are serpentinite, amphibolite and other mafic rocks exposed along the western margin of the Barkerville Terrane. The Slide Mountain Terrane contains correlative sequences the length of the Cordillera (Struik and Orchard, 1985). The correlatives include Anvil Range Group of Yukon, Sylvester Group of northern British Columbia and Kaslo Group and Fennell Formation of southern British Columbia. There are many similarities between the Slide Mountain Terrane and the Havallah sequence of Nevada.

The Quesnel Terrane lies west of the root zone of the Slide Mountain Terrane (Fig. 2, 3). Within the map area, it consists of Upper Triassic and Lower Jurassic mainly black slate and volcaniclastic greenstone. Lesser amounts of micritic limestone, conglomerate and sandstone are associated with the black slate (Fig. 4). The lower contact with more easterly terranes may be either tectonic or stratigraphic; no unequivocal evidence has been found to support either relationship. The Quesnel Terrane encompasses a broad area of Upper Triassic and Jurassic rocks throughout the Intermontane Belt of the Canadian Cordillera.

The time of emplacement of both the Barkerville and Slide Mountain against the Cariboo Terrane is constrained between the Early Permian and Late Cretaceous (Fig. 5). However, from various regional arguments the time interval can be

Le terrain de Slide Mountain se compose de basalte mississippien à permien, trouvé par endroits en coussins, et de séquences de chert et de pélite coupées par des intrusions de diorite, de gabbro et de petites quantités de roches ultramafiques (fig. 4). On y trouve par endroits du grauwacke, du sable grossier, du conglomérat et de la serpentine. Ce terrain recouvre les terrains de Barkerville et de Cariboo le long du chevauchement de Pundata et sa partie intérieure a été imbriquée par des failles chevauchantes. On y trouve également de la serpentinite, de l'amphibolite et d'autres roches mafiques qui affleurent le long de la marge ouest du terrain de Barkerville. Le terrain de Slide Mountain contient des séquences corrélatives le long de la Cordillère (Corey et al., 1982), ces séquences comprennent le groupe d'Anvil Range du Yukon, le groupe de Sylvester du nord de la Colombie-Britannique ainsi que le groupe de Kaslo et la formation de Fennell du sud de la Colombie-Britannique. Il existe de nombreuses similarités entre le terrain de Slide Mountain et la séquence d'Havallah du Nevada.

Le terrain de Quesnel se trouve à l'ouest de la zone d'enracinement du terrain de Slide Mountain (fig. 2, 3). À l'intérieur de la région cartographique, le terrain de Quesnel comprend surtout des schistes noirs et des roches vertes volcanoclastiques du Trias supérieur et du Jurassique inférieur. Des quantités inférieures de calcaire micritique, de conglomérat et de grès sont associées au schiste noir (fig. 4). Le contact inférieur entre les terrains plus à l'est peut être de nature tectonique ou stratigraphique; aucun indice non équivoque n'a été découvert jusqu'à présent pour appuyer l'une ou l'autre de ces possibilités. Le terrain de Quesnel englobe une vaste zone de roches du Trias supérieur et du Jurassique dans la zone intermontagneuse de la Cordillère du Canada.

La mise en contact des terrains de Barkerville et de Slide Mountain avec le terrain de Cariboo aurait eu lieu entre le Permien ancien et le Crétacé tardif (fig. 5). Toutefois, diverses indications régionales permettent de raccourcir cet intervalle.

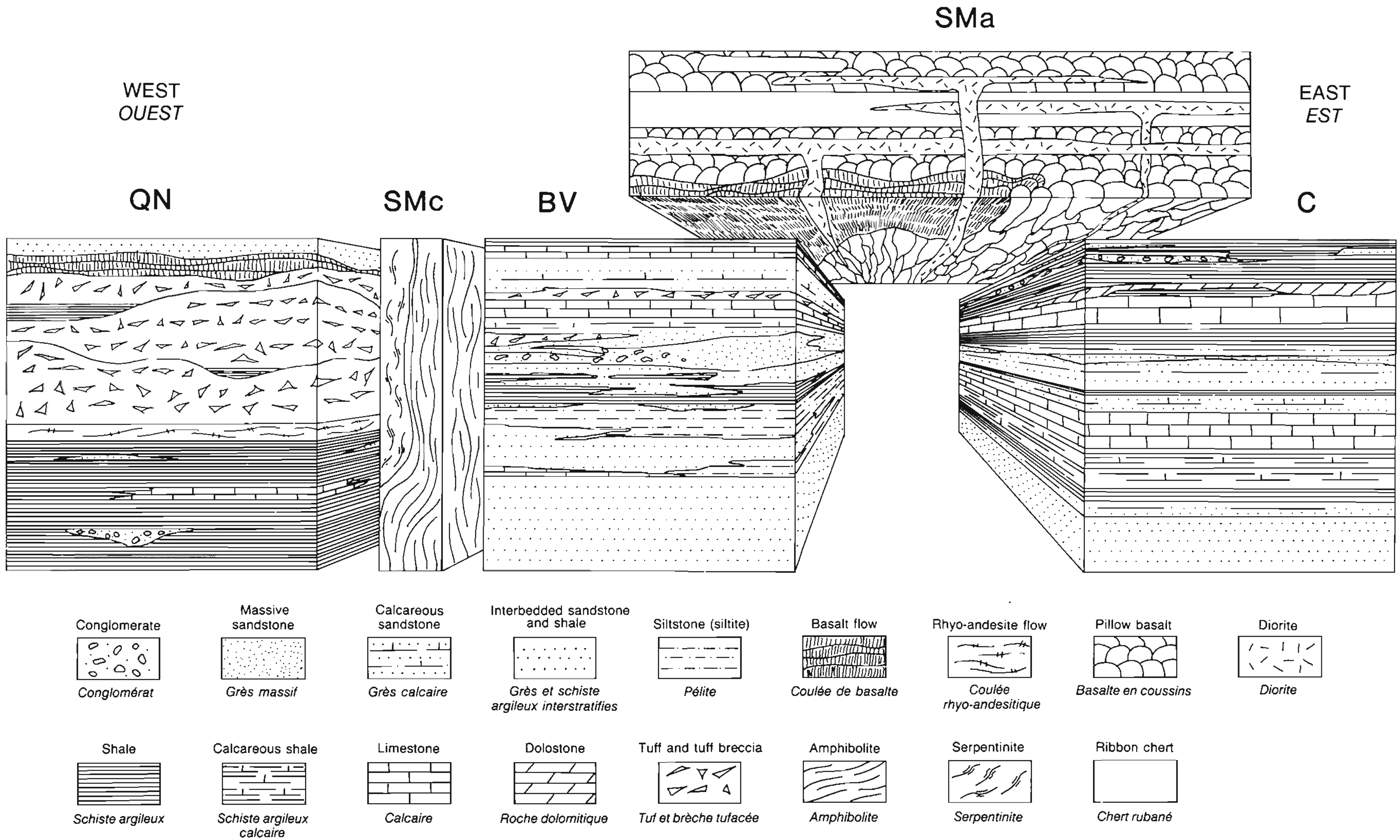

Figure 4. Simplified stratigraphy of the four terranes (symbols as in Fig. 3) of the Cariboo gold belt shown in their relative structural positions.

Figure 4. Stratigraphie simplifiée des quatre terrains (les symboles étant les mêmes qu'à la fig. 3) de la zone aurifère de Cariboo, présentés dans leur position structurale relative.

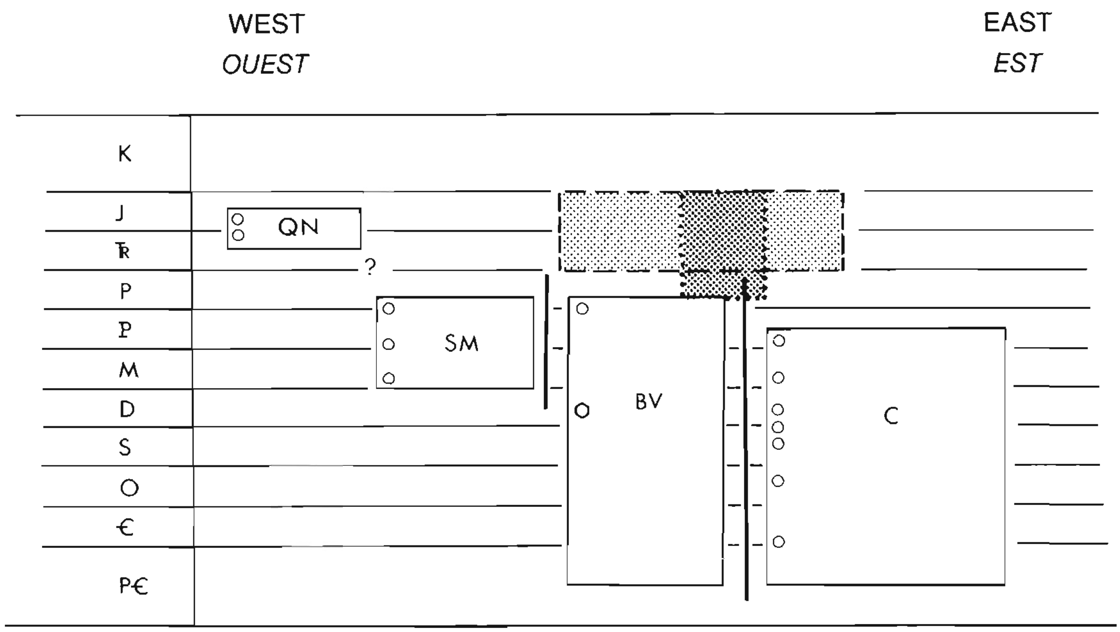

Figure 5. A time-space diagram of the four terranes (symbols as in Fig. 3) of the Cariboo gold belt showing their age ranges and their relative geographic position. Open circles are fossil ages confined to an epoch. The open hexagon is a fossil age spanning the Paleozoic. The finely patterned block gives the maximum age range during which the Cariboo Terrane was thrust on the Barkerville Terrane. The coarsely patterned block defines the maximum age range during which the Slide Mountain Terrane was thrust onto Barkerville and Cariboo terranes.

Figure 5. Diagramme spatio-temporel des quatre terrains (les symboles étant les mêmes qu'à la fig. 3) de la zone aurifère de Cariboo montrant leurs tranches d'âge et leur position géographique relative. Les cercles vides représentent des âges paléontologiques limités à une seule époque. L'hexagone vide est un âge paléontologique embrassant l'ensemble du Paléozoïque. Le bloc finement ligné donne la tranche d'âge maximale pendant laquelle le terrain de Cariboo a chevauché le terrain de Barkerville. Le bloc grossièrement ligné définit la tranche d'âge maximale pendant laquelle le terrain de Slide Mountain a chevauché les terrains de Barkerville et de Cariboo.

narrowed. The youngest age from the Fennell Formation is Late Permian (M.J. Orchard *in* Schiarizza and Preto, 1984) and the equivalent Antler Formation is assumed to include rocks at least as young, hence, the Slide Mountain Terrane (Antler, Fennell etc.) was emplaced after the Late Permian. The terrane is involved in regional folding that is dated to the southeast as primarily Middle Jurassic (Campbell, 1971) and therefore would have been in place prior to, or during, the Middle Jurassic. Emplacement of the Slide Mountain Terrane from these arguments is during the Triassic or Early Jurassic (Fig. 5). The Barkerville and Cariboo terranes were juxtaposed prior to emplacement of the Slide Mountain because the Slide Mountain Terrane was thrust over both of them. Therefore, the Barkerville was in fault contact with the Cariboo Terrane prior to, or during, the Early Jurassic and probably after the deposition of Pennsylvanian strata of the Cariboo Terrane (Fig. 5).

The terranes are defined by their different stratigraphies; they may also have different structural

Dans la formation de Fennell, les roches les plus récentes datent du Permien tardif (M.J. Orchard *dans*: Schiarizza et Preto, 1984) et la formation équivalente d'Antler comprendrait des roches au moins aussi récentes; le terrain de Slide Mountain (Antler, Fennell, etc.) a donc été mis en place après le Permien tardif. Le terrain est touché par des plis régionaux qui datent surtout du Jurassique moyen (Campbell, 1971) au sud-est et qui auraient donc déjà existé avant ou pendant le Jurassique moyen. La mise en place du terrain de Slide Mountain daterait donc du Trias ou du Jurassique ancien (fig. 5). La juxtaposition des terrains de Barkerville et de Cariboo a eu lieu avant la mise en place du terrain de Slide Mountain puisque celui-ci chevauche les deux premiers. Ainsi, une faille aurait poussé le terrain de Barkerville contre le terrain de Cariboo avant ou pendant le Jurassique ancien et probablement après l'accumulation des couches pennsylvaniennes du terrain de Cariboo (fig. 5).

histories. Therefore they are described independently of each other in all geological aspects. Because the faults between the terranes are referred to throughout the text, their definition and description follows.

Terrane boundary faults

Pleasant Valley Thrust

The Pleasant Valley Thrust is the boundary between the Cariboo and Barkerville terranes. It trends southeast from Big Valley Creek to the southern edge of the map area. It dips from 40 to 75° to the east. It is marked by cataclastic rocks north of Cariboo Lake and by mylonite south of the lake. North of Cariboo Lake it is exposed where it crosses Antler Creek, and near the mouth of Pleasant Valley Creek. To the south mylonitic rocks of the fault zone are exposed along the northeast bank of Little River.

Pundata Thrust

The Pundata Thrust was named by Struik (1981). It is defined as the fault that separates the Antler Formation of the Slide Mountain Terrane from the underlying rocks of the Barkerville and Cariboo terranes. The fault traverses the area from Big Valley Creek southeastward to Cunningham Creek. It is generally shallowly dipping although folding in the area of Yuzkli Creek has produced steep dips. The fault is exposed in Pundata Creek and in drill core from the area near Prince George (viewing courtesy of Vestor Exploration and partners). In both cases the width of the fault zone ranges up to several metres. It consists of phyllonite and/or disorganized breccia, primarily of broken and sheared clastic sedimentary rocks. Characteristic of the fault zone is the lack of mylonitic rock.

Eureka Thrust

The Eureka Thrust forms the boundary between the western edge of the Barkerville Terrane and the Slide Mountain and Quesnel terranes in the southwestern part of the area mapped. It is named from Eureka Peak in MacKay River (93A/7) map area to the southeast where the fault is folded in a regional anticline-syncline pair. The fault was mapped in detail by Campbell (1971), Rees (1981), Struik (1983) and J. Carye (personal communication, 1983), trends northwest, and is apparently nearly layer parallel. It is folded and marked by amphibolite and metamorphosed mafic rocks correlated with the Antler Formation of the Slide Mountain Terrane. The fault can be traced north of the map area to Stone Creek near Prince George and south to Mahood Lake.

Chacun de ces terrains est défini en fonction de sa stratigraphie unique; en outre, chacun peut avoir une histoire structurale différente de celles des autres. Il est donc possible de décrire ces terrains indépendamment les uns des autres sous tous les aspects de la géologie. Puisque les failles qui séparent ces terrains sont mentionnées partout dans le texte, elles sont définies et décrites dans les paragraphes suivants.

Failles-limites

Chevauchement de Pleasant Valley

Le chevauchement de Pleasant Valley sépare le terrain de Cariboo de celui de Barkerville. Il a une orientation nord-ouest-sud-est et s'étend du ruisseau de Big Valley jusqu'à la marge sud de la région cartographique. Son pendage plonge vers l'est de 40 a 75°. Il se caractérise par la présence de roches cataclastiques au nord du lac Cariboo et de mylonite au sud du lac. Il est visible au nord du lac Cariboo à l'endroit ou il traverse le ruisseau Antler et près de l'embouchure du ruisseau Pleasant Valley. Au sud, les roches mylonitiques de la zone de failles affleurent le long de la rive nord-est de la rivière Little.

Chevauchement de Pundata

Le chevauchement de Pundata, baptisé par Struik (1981), est défini comme étant la faille qui sépare la formation d'Antler du terrain de Slide Mountain des roches sous-jacentes des terrains de Barkerville et de Cariboo. Il traverse la région à partir du ruisseau de Big Valley vers le sud-est jusqu'au ruisseau Cunningham. Son pendage est généralement faible, bien que la formation de plis aux environs du ruisseau Pundata ait produit des pendages abrupts. Il est visible dans le ruisseau Pundata et dans des carottes provenant des environs de Prince George (ces carottes ont été examinées avec la permission de Vestor Exploration et associés). Dans les deux cas, la largeur de la zone de failles varie et peut atteindre plusieurs mètres. Le chevauchement de Pundata se compose de phyllonite ou de brèche dérangée composée principalement de roches sédimentaires clastiques brisées et cisaillées, ou des deux. L'absence de roches mylonitiques est caractéristique de la zone de failles.

Chevauchement d'Eureka

Le chevauchement d'Eureka représente la limite entre la marge ouest du terrain de Barkerville et les terrains de Slide Mountain et de Quesnel dans la partie sud-ouest de la région cartographiée. Il tire son nom du mont Eureka dans la région cartographique de la rivière Mackay (93A/7) au sud-est, à l'endroit ou la faille a été plissée pour former un groupe anticlinal-synclinal régional. Cartographiée en détail par Campbell (1971), Rees (1981, 1983), Struik (1983) et J. Carye (communication personnelle, 1983), ce chevauchement a une orientation nord-ouest et serait presque parallèle aux couches. Il est plissé et marqué par la présence d'amphibolite et de roches mafiques métamorphosées qui ont été mises en corrélation avec la formation d'Antler du terrain de Slide Mountain. La faille peut être tracée vers le nord au-delà de la région cartographique jusqu'au ruisseau Stone près de Prince George, et au sud jusqu'au lac Mahood.

Within the map area the Eureka Thrust is marked by sheared ultramafic rocks and serpentinite. Black phyllite, siltite and minor sandstone and limestone of the Triassic and Lower Jurassic of the Quesnel Terrane are locally in fault contact with the grit and pelite of the Barkerville Terrane where recognizable Slide Mountain Terrane rocks are absent.

Dans la région cartographique, le chevauchement d'Eureka se caractérise par la présence de roches ultramafiques cisaillées et de serpentinite. Par endroits, de la phyllade noire, de l'aleurolite et de faibles quantités de grès et de calcaire du Trias et du Jurassique inférieur du terrain de Quesnel ont été mis en contact avec le sable grossier et la pélite du terrain de Barkerville, là ou l'on ne reconnaît aucune roche du terrain de Slide Mountain.

ROCKS OF THE CARIBOO TERRANE

The Cariboo Terrane, bound on the west by the Pleasant Valley Thrust and the Barkerville Terrane, is underlain by an unknown basement and is terminated upwards by the Pundata Thrust and the tectonically emplaced Slide Mountain Terrane. The Cariboo Terrane extends south of the map area to Azure Lake, east to the Rocky Mountain Trench and north to Macleod Lake. Its strata correlate with rocks of the Cassiar Mountains (Mansy and Gabrielse, 1978).

Stratigraphic nomenclature for rocks of this terrane is retained from earlier work with the addition of the Alex Allan Formation. The status and definitions of some of the units are changed. The Black Stuart is elevated to group status and the Greenberry limestone member of the Guyet Formation is defined as a formation (Greenberry Formation). The type areas of the Cunningham, Yankee Belle, Yanks Peak and Midas formations of the Cariboo Group are abandoned and suggested instead to be the reference sections described by Campbell et al. (1973).

Stratified rocks

Introduction

The sequence of rocks ranges from Upper Hadrynian to Middle Pennsylvanian and possibly Permo-Triassic (Table 1). The lowest member of the section is the Kaza Group (Sutherland Brown, 1957,1963; Campbell et al., 1973). Overlying the Kaza Group with gradational conformity is the Cariboo Group, which ranges from Upper Hadrynian to Cambrian.

The Ordovician to Mississippian Black Stuart Group unconformably overlies the Kaza and Cariboo groups. The Mississippian Greenberry Formation overlies the Waverly and Guyet formations of the Black Stuart Group. Since Johnston and Uglow (1926) first described them the rocks of the Guyet, Greenberry and Antler formations have been included in the Slide Mountain Series or Group (see Table 2). This report proposes to remove the Guyet and Greenberry formations from the Slide Mountain Group and to include them in the Black Stuart Group with which they intertongue and overlie.

Kaza Group

The type section of the Kaza Group was described by Sutherland Brown (1963, p. 13-19) from Bowron Lake Park where he estimated the thickness to be approximately 3940 m. The base is not exposed. Predominant rock types are brown weathering feldspathic and micaceous quartzite, silver-green phyllite and schist and schistose granule conglomerate. The quartzite is thick bedded (3 m), forming 60 to 90 m thick units relatively free of phyllite.

Within the field area, the Kaza Group coarse grained, poorly sorted feldspathic, micaceous quartzite (grit) is mapped in one locality; near the mouth of the first large south-flowing tributary of Little River.[1] The exposure is too poor to warrant significant lithological or stratigraphical descriptions, for which reference can be made to Sutherland Brown (1963), Murphy and Rees (1983) and Pell and Simony (1984).

Cariboo Group

Introduction

The Cariboo Group, which conformably overlies the Kaza Group on a gradational contact, has a long history of re-interpretation after being initially introduced by Bowman (1889) as the Cariboo Series. Within this paper, the evolution of Cariboo Group stratigraphy follows the interpretation of Young (1969), Mansy (1970) and Campbell et al. (1973), and its historical evolution is outlined in Table 2 (in pocket).

The Cariboo Group is subdivided into seven formations which are from oldest to youngest: Isaac, Cunningham, Yankee Belle, Yanks Peak, Midas, Mural and Dome Creek. Within the described area the group is thickest at Kimball Mountain where it attains at least 2500 m. Elsewhere it is thinner, but thickness variations are poorly understood due to complexities of structure, and erosion below an Ordovician unconformity.

Contacts within the group are gradational or conformable. For descriptions of the group's formations in

[1] This creek will be informally called Separation Creek throughout this report.

Table 1. Table of formations for the Cariboo Terrane.

Era	Period	Group	Formation/unit	Lithology
PALEOZOIC	PENNSYLVANIAN		*(0-8 m)*	*Grey crinoidal, fusulinid limestone*
			Disconformity	
PALEOZOIC	MIDDLE PENNSYLVANIAN		*ALEX ALLAN FORMATION (0-5 m)*	*Dark grey micritic limestone, minor slate*
			Disconformity	
PALEOZOIC	LOWER MISSISSIPPIAN	BLACK STUART GROUP	*GREENBERRY FORMATION (0-30 m)*	*Grey crinoidal limestone*
			Conformity	
PALEOZOIC	LOWER MISSISSIPPIAN AND UPPER DEVONIAN	BLACK STUART GROUP	*GUYET FORMATION (0-300 m)*	*Conglomerate, orthoquartzite, greywacke*
			Disconformity?	
PALEOZOIC	MIDDLE DEVONIAN OR UPPER DEVONIAN	BLACK STUART GROUP	*WAVERLY FORMATION (0-50 m)*	*Agglomerate, pyroclastic, pillow basalt, minor chloritic siltstone*
			Interdigitating contact	
PALEOZOIC	MISSISSIPPIAN OR YOUNGER	BLACK STUART GROUP	*Sandstone unit (0?-50 m)*	*Olive grey micaceous and white quartzite, black and pink chert*
			Conformity	
PALEOZOIC	DEVONIAN AND (?) YOUNGER	BLACK STUART GROUP	*Black pelite unit (300-400 m)*	*Dark grey and black slate, phyllite, argillite, siltite, dolostone and limestone*
			Conformity?	
PALEOZOIC	LOWER DEVONIAN AND UPPER SILURIAN	BLACK STUART GROUP	*Chert-carbonate unit (0-60 m)*	*Mottled chert breccia, grey dolostone breccia, light grey dolostone chert*
			Disconformity?	
PALEOZOIC	UPPER ORDOVICIAN	BLACK STUART GROUP	*Black pelite unit (0-50 m?)*	*Dark grey slate and minor siltstone*
			Unconformity	
PALEOZOIC	LOWER TO (?) UPPER CAMBRIAN	CARIBOO GROUP	*DOME CREEK FORMATION (0-50 m)*	*Dark grey slate, shale and minor grey limestone*
			Conformity	
PALEOZOIC	LOWER CAMBRIAN	CARIBOO GROUP	*MURAL FORMATION (50-500 m)*	*Grey limestone, dolostone, fine marble*
			Conformity	
PALEOZOIC AND PROTEROZOIC	LOWER CAMBRIAN AND HADRYNIAN	CARIBOO GROUP	*MIDAS FORMATION (40-250 m)*	*Grey shale, slate, phyllite and micaceous quartzite, dark grey siltite*
			Conformity	
PALEOZOIC AND PROTEROZOIC	LOWER CAMBRIAN AND HADRYNIAN	CARIBOO GROUP	*YANKS PEAK FORMATION (0-290 m)*	*Dark grey to white quartzite, minor shale and granule quartzite*
			Gradational contact	
PROTEROZOIC	HADRYNIAN (WINDERMERE)	CARIBOO GROUP	*YANKEE BELLE FORMATION (170-1000 m)*	*Green-grey micaceous quartzite, siltite, grey-green shale, slate and phyllite, limestone and sandy limestone*
			Gradational contact	
PROTEROZOIC	HADRYNIAN (WINDERMERE)	CARIBOO GROUP	*CUNNINGHAM FORMATION (400-650 m)*	*Limestone, dolostone, fine grained marble*
			Gradational contact	
PROTEROZOIC	HADRYNIAN (WINDERMERE)	CARIBOO GROUP	*ISAAC FORMATION (0-1200 m)*	*Dark grey to black phyllite, slate, limestone and minor calcareous sandstone*
			Gradational contact	
PROTEROZOIC	HADRYNIAN	KAZA GROUP	*BASE NOT EXPOSED*	*Micaceous poorly sorted feldspathic quartzite, grey-green and grey phyllite, limestone*

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their type areas refer to Sutherland Brown (1963) for the Isaac Formation and to Campbell et al. (1973) for the entire Cariboo Group. Lithological descriptions of the formations in the following discussions are compiled from incomplete sections within the map area.

Isaac Formation

The Isaac Formation consists of phyllite, calcareous phyllite, limestone and minor quartzite. The thin-bedded marly layering and carbonate content of the phyllite is characteristic of the unit. The calcareous phyllite distinguishes it from the Dome Creek Formation which also has interbedded limestone and phyllite. Isaac Formation pelite is more calcareous than the Black Stuart Group and the pelite is generally less siliceous.

Isaac Formation occurs on the east slopes of Kimball Ridge extending into the valley of Matthew River, in the valley of Lostway Creek and north of Separation Creek. The type area of the Isaac Formation is along the east side of Isaac Lake in the Cariboo Mountains. It was first described by Sutherland Brown (1963) and later by Young (Young, 1969; *in* Campbell et al., 1973). Within the map area, Mansy (1970) described Isaac Formation at Kimball Ridge. The formation is thickest (approximately 1000 m) in the Matthew River area and may thin westward.

Although the base of the Isaac Formation near Mount Kimball is not seen, the upper contact with the Cunningham Formation is gradational. The gradational interval involves thinning of the phyllite beds and thickening and lightening in colour of the limestone. The top of the Isaac Formation is placed at the dark brown weathering dark grey micritic limestone which is interbedded with black phyllite in near equal proportions beneath the massive thick carbonate of the Cunningham Formation. The base of the Cunningham Formation has 3 to 5 cm beds of grey micritic limestone separated by lamellae of black phyllite (0.5 cm or less). Approximately 8 m of the base of the Cunningham Formation and some 120 m of the upper part of the Isaac Formation define the gradational interval.

For another 250 m downsection steel-blue weathering black to dark grey phyllite is interbedded with 2 to 6 cm calcareous brown weathering phyllite and minor brown weathering dark grey micritic limestone. The phyllite contains porphyroblasts (2 to 4 cm) of pyrite and ankerite. Scattered outcrops downsection from Kimball Ridge and extending into Matthew River valley consist of the same lithology but with occasional beds of grey limestone (3 to 10 m thick). In the valley calcareous phyllite with marly laminae (1 to 5 cm) is typical of the type section on Isaac Lake. Associated with the phyllite and calcareous phyllite are rare 1 to 2 m beds of moderate to coarse grained greenish-grey micaceous quartzite. The section on Kimball Ridge is more carbonate-rich than that at Isaac Lake.

Campbell et al. (1973, p. 34) reported that the Isaac Formation becomes more calcareous westward from the Rocky Mountain Trench where it consists of phyllite, siltstone and minor sandstone directly below the Cunningham Formation. The increase in limestone content westward is confirmed by the Matthew River-Kimball Ridge exposures.

Age and correlation. Young (1969 and in Campbell et al., 1973) correlated the Isaac Formation with the upper Miette Group of the Rocky Mountains because of similar lithology and thickness and both units overlie lithologically similar rocks, the Kaza Group and Middle Miette Group, respectively. The Isaac Formation, like the Upper Miette, is considered upper Proterozoic (Young, 1969). No evidence has been found to either substantiate or refute this correlation. The calcareous phyllite of the Matthew River-Mount Kimball area is most certainly Isaac Formation and can therefore be correlated with occurrences of that formation in the Cariboo Mountains to the east.

Cunningham Formation

The Cunningham Formation consists of limestone, dolostone marble and minor phyllite. In solitary outcrops the carbonate is easily mistaken for parts of the Isaac or Mural formations. In more continuous exposures the Cunningham can be distinguished from the Mural Formation by more common interleaving of black carbonaceous phyllite in its lower sections and by being gradational with both the underlying Isaac and overlying Yankee Belle formations. The Mural Formation is in sharp contact with bounding units.

Cunningham Formation underlies much of the area east of Pleasant Valley Thrust. However, it is everywhere incompletely exposed due either to overburden or structural disruption.

The type section for the Cunningham Formation is defined as the reference section described by Campbell et al. (1973) to the northeast at Haggen Creek, Cariboo Mountains. That section is complete and represents the present concept of the Cunningham Formation. The type section originally defined by Holland (1954) near Roundtop Mountain is incomplete and contains rocks other than Cunningham Formation.

At the reference locality the Cunningham Formation consists entirely of limestone except for minor grey shale near its lower contact with the Isaac Formation. Young (*in* Campbell et al., 1973) pointed out that west and east of this section the Cunningham Formation includes shale, siltstone and sandstone whereas to the east near the Rocky Mountain Trench it is primarily dolostone. He reported the Cunningham Formation to be thickest (550 m) at Haggen Creek. Thicknesses of the Cunningham Formation in the map area are:

360 m;	Kimball Ridge
640 m;	south of Separation Creek and just north of Little River
420 m;	below the thrust fault that puts Cunningham

on Cunningham Formation in the north bank of Separation Creek
500 m; Shepherd Creek, north-northeast of Wells.

The Cunningham Formation limestone is predominantly grey weathering, grey to cream or greenish grey, and medium to finely crystalline. It is best preserved at the base of the formation on the east end of Kimball Ridge where it retains a micritic texture and is thin bedded as defined by carbonaceous pelitic lamellae. Limestones throughout the remainder of the section display the profound changes in texture usually found throughout the map area. Micritic limestone grades to very finely crystalline limestone to a coarsely crystalline marble. Colour changes follow most variations in crystal size. The darkest limestone is micrite and with increasing crystal size it varies from dark grey to creamy white. Marbles and moderately crystalline limestone commonly are tinged with green, a colour rarely found in carbonates of the Mural Formation. Increased recrystallization of the micritic limestone generally correlates directly with increased stratigraphic and structural depth but is also very much controlled by tectonism. Cunningham Formation is nearly completely marble in its southeasternmost outcrops where it is structurally the lowest. Marble also occurs throughout the map area but is usually associated with faulting and tight folding. Commonly found within the limestone and marble are knots of dolomite crystals. They are usually from 0.5 to 2 mm across and account for 1 to 10% of the limestone, but locally may be more abundant. The dolomite is usually ferroan, imparting a pinkish speckling where the iron has become oxidized. A good example of this type of carbonate is found in the cliff banks of Sixbee Creek, where the flow changes from northerly to southerly.

Young (*in* Campbell et al., 1973) used ferroan dolomite content as a method of distinguishing Cunningham Formation from Mural Formation limestones in the Cariboo Mountains. This criterion has not proven useful because it cannot be demonstrated to be confined to the Cunningham Formation. For example, on the south slope of Kimball Mountain Mural Formation limestone is spotted with dolomite near a small fault. The dolomite is not ferroan at this locality, but ankerite and ferroan dolomite are recorded from every other formation of the Cariboo Group.

Finely crystalline limestone commonly displays moderate grey and light grey to white banding (0.5 to 2 cm) which is found in all the limestone of the area and is not restricted to the Cunningham Formation. The bands commonly are layer-parallel and isoclinally folded, yet are parallel to regional bedding trends and most can be assumed to mimic original bedding.

Dolostone is best developed along Roundtop Creek and its junction with Cunningham Creek. There very coarse dolostone is associated with quartz veins with minor amounts of galena. It varies from creamy white to light grey and moderately crystalline, sugary textured to very coarsely crystalline. Where a small percentage of the dolomite is ferroan the dolostone is buff or light brown. Dolostone in the cliffs overlooking Lostway Creek occurs along faults. The Lostway Creek thrust, which puts Cunningham Formation onto Mural Formation, has a glide zone some 1 to 3 m thick composed of sugary textured white dolostone.

The irregular distribution of dolostone and limestone is best explained by local controls of dolomitization and is not a regional phenomenon resulting from paleoenvironmental factors. Pore spaces that formed during initial dolomitization were later infilled with anhedral dolomite indicating redistribution of carbonate material. The dolomitization process can be seen to cut across bedding, forming irregular pods of dolostone in limestone. The best example of this is where Mural Formation is faulted against Yankee Belle Formation at the head of Roundtop Creek.

Young (*in* Campbell et al., 1973, p. 37) reported "pisolites, oolites, pellets, algal-coated complex grains, intraclasts and unidentified ovoid grains" from the Cunningham Formation limestone of the Cariboo Mountains. Record of these features is scanty. The dolomite knots may be dolomitized remnants of some type of pelletoid (*see* Sutherland Brown, 1957, 1963).

Table 3. Reference section of Cunningham Formation on Kimball Ridge (location; 52°55'10"N, 121°00'10"W).

Unit	Lithology	Thickness in metres Unit	Above base
4	Limestone to fine marble, grey to light greenish grey, moderately crystalline	9	360
	Covered	18	351
3	Limestone, grey and buff mottled weathering grey and light orange brown, colours are streaked and patched, upper section is spotted with 3% orange ferroan dolomite	4	333
	Covered, probably limestone	90	329
2	Limestone, moderately light grey weathering grey, moderately fine to finely crystalline	7	239
	Covered, probably limestone	100	232
1	Limestone, grey weathering grey to dark grey, fine grained; interleaved with films of silvery grey phyllite in 3 to 5 cm beds, 1%, 1 to 2 mm pyrite cubes disseminated throughout	12	132
	Covered	120	120
	Underlying beds; interbedded brown-weathering dark grey limestone, and black phyllite of Isaac Formation.		

The section of Cunningham Formation on Kimball Ridge (location; 52°55'10"N, 121°00'10"W) provides a local reference and is not meant to replace the type section (Table 3).

Age and correlation. No new information has been gathered on the dating of the Cunningham Formation. An upper Proterozoic age was assigned by Young (*in* Campbell et al., 1973) and is accepted in this report. It is correlated with the Espee Formation of northern British Columbia (Mansy and Gabrielse, 1978).

Yankee Belle Formation

The Yankee Belle Formation consists of slate or phyllite, quartzite, siltite, limestone and sandy limestone. Isolated exposures of pelite, siltite or fine grained quartzite can be confused with similar rocks of the Midas Formation or parts of the Keithley succession of the Barkerville Terrane. Yankee Belle Formation limestone has most of the characteristics of Cunningham Formation limestone. In more continuous exposures the gradation upsection from interbedded limestone, quartzite, and pelite to dominant pelite and siltite with lesser olive quartzite to white quartzite and, near the top, to olive pelite is typical of the Yankee Belle Formation.

The formation underlies much of the area southeast of Cunningham Creek and east of the Pleasant Valley Thrust. A small outcrop of the unit is exposed on Summit Creek.

The type section of the Yankee Belle Formation is defined as the reference section (908 m thick) described by Young (*in* Campbell et al., 1973) in northern Cariboo Mountains at the headwaters of Dome Creek. The original type section from Yanks Peak described by Holland (1954) is abandoned because it is not Yankee Belle Formation as presently used. Those rocks are part of the Barkerville Terrane and have only remote, if any, affinities to Yankee Belle of the Cariboo Terrane. A composite section of Yankee Belle Formation on Kimball Ridge has a thickness of approximately 1085 m, although Mansy (1970) measured a similar composite section on Kimball Ridge as being 600 m thick. Thicknesses in other parts of the area are: approximately 150 m, 5 km southwest of Black Stuart Mountain and 170 to 500 m in Roundtop Mountain area. The relationship of thickness variation to geographic position is complicated by east to west thrusting (Fig. 6).

The lower contact of the Yankee Belle Formation is conformable and gradational with the Cunningham Formation. It is defined as the first occurrence of quartzite above the massive limestone of the Cunningham Formation.

Limestone of the Yankee Belle Formation is grey to greenish grey and may be white or light greenish cream due to recrystallization to marble. It forms 1 to 25 m members near the base, decreasing in frequency upsection from the Cunningham Formation contact. Though

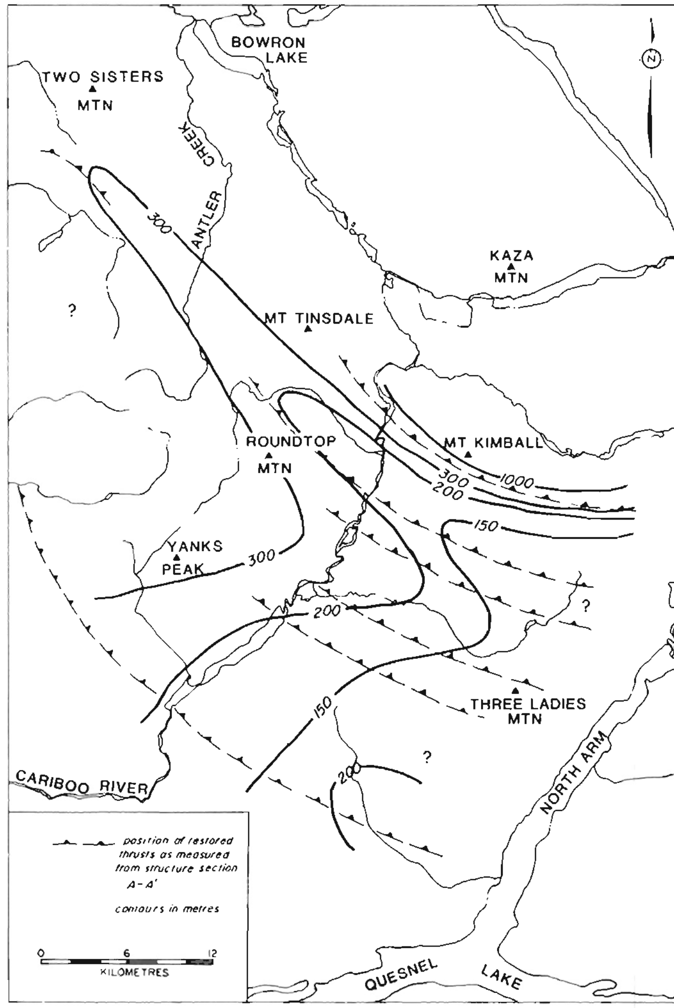


Figure 6. Palinspastic isopach map of the Yankee Belle Formation.

the thickness of the Yankee Belle Formation varies considerably, south of Cariboo River, everywhere its lower half has interbedded limestone and phyllite or slate. North of Cariboo River, limestone is not found in the Yankee Belle Formation, but Summit Creek is the only place where the base of the Yankee Belle Formation is known with certainty. On the south slope of Roundtop Mountain, Yankee Belle Formation quartzite and slate overlie grey limestone, but whether the limestone is a basal member of the Yankee Belle Formation or part of the Cunningham Formation is not known. Sandy limestone is found only on Kimball Ridge and 5 km southeast of Black Stuart Mountain.

Pelletoidal limestone on Kimball Ridge has undergone irregular neomorphism, producing patches of finely and moderately fine crystalline calcite matrix surrounding 0.5 to 3 mm pellets. The pellets have very fine to finely crystalline calcite in their cores. In thin section concentric outlines of grey opaque material ring the centres. In some cases radial crystal growth occurs between the opaque rings and in one example cuts across several rings. The radial growth is possibly secondary, and may be superimposed over an algal laminate structure. A sample examined in detail contains approximately 60% pellets, 4% silt-size quartz, 3% muscovite (silt-size) and 33% matrix of calcite and fine opaques.

Grey to olive siltite is found in all occurrences of Yankee Belle Formation but is more prominent in western exposures and north of Cariboo River. Grey, grey-olive and lesser amounts of white quartzite are found in most exposures except at Summit Creek. It is fine- to medium-grained, micaceous and sorted. It occurs throughout the section but coarsens and increases in percentage stratigraphically upward everywhere except at Summit Creek. Gradations within the sequence occur as discrete but small clast-size changes between beds, not as internally graded beds. Bedding thickness ranges from 3 to 75 cm and is commonly 10 to 30 cm. The siltstone-quartzite sequences develop a characteristically 0.5 to 1.5 cm spaced cleavage which anastomoses irregularly around lenses of quartzite. Light grey to white fine grained quartzite beds commonly occur near the top of the Yankee Belle Formation below the massive white quartzites of the Yanks Peak Formation. South of Cariboo River, in western exposures, quartzite does not occur with the basal limestones. In eastern exposures there is more quartzite throughout the section and it is coarser grained.

Both Mansy (1970) and Young (1969, *in* Campbell et al., 1973) described three cycles of sedimentation within the Yankee Belle Formation. The lowest cycle consists of a sequence of shale beds which coarsen upsection to a cap of sandstone, the middle one of repeating limestone and pelite beds and the upper one of upward coarsening shale to sandstone sequences which at the top coarsen into the overlying quartzite of the Yanks Peak Formation. The middle and upper cycles, but not the lower, are found south of Cariboo River and are well developed on Kimball Ridge. North of Cariboo River only the upper cycle of shale-sandstone was observed.

A reference section of Yankee Belle Formation (Table 4) is on eastern Kimball Ridge (52°55'N,121°01''W).

Age and correlation. Campbell et al. (1973) mapped the Yankee Belle Formation as Hadrynian, because they correlated it with the uppermost part of the Hadrynian upper Miette Group of the Rocky Mountains. A summary diagram showing recent correlations is displayed by Mansy and Gabrielse (1978, p. 2). No new information was gathered to re-evaluate these correlations.

Yanks Peak Formation

The Yanks Peak Formation consists of quartzite, siltite, slate, phyllite and minor calcareous sandstone. White coarse grained orthoquartzite is characteristic of the Yanks Peak Formation in the Cariboo Terrane. It may be confused with a similar, but better sorted and massive quartzite, of the Keithley succession of the Barkerville Terrane. The Yanks Peak Formation orthoquartzite, however, has a larger variation in grain size and colour, is slightly richer in muscovite and shows bedding and locally crossbedding.

The Yanks Peak Formation mainly underlies the area south of Cunningham Creek Pass and east of the Pleasant Valley Thrust. An isolated exposure occurs at Summit Creek northeast of Wells. The type section of the Yanks Peak Formation is defined as the reference section described by Young (*in* Campbell et al., 1973) from an area at the headwaters of Dome Creek in northern Cariboo Mountains. The original type section on Yanks Peak was described by Holland (1954), but is no longer representative of the Yanks Peak Formation as viewed today. It does not contain rocks reliably equivalent to those called Yanks Peak Formation in the Cariboo Terrane and should be abandoned. Young reported the Yanks Peak Formation to have a maximum thickness of 580 m near Betty Wendle Creek. It thins to the north and northeast and at the type section is 403 m thick. In the map area the thickness ranges from 1 to approximately 300 m or more. On Kimball Ridge the Yankee Belle Formation is capped by 280 m of Yanks Peak Formation quartzite below a zone where structural complications disrupt the sequence.

The lower contact of the Yanks Peak Formation is conformable with the Yankee Belle Formation. In most places it is gradational but locally sharp (Fig. 7). Gradational relationships are best seen on Roundtop and Kimball mountains. There the Yankee Belle Formation has increasing numbers and thickness of white quartzite beds within a 30 m thick interval approaching its upper contact. The contact is defined at the base of the first thick bed (1.5 m or more) of medium- to coarse-grained white orthoquartzite.

Quartzite of the Yanks Peak Formation is generally white and can be light grey to black, brown or pink. In western exposures the colour changes upsection from white to dark grey. The quartzite is mostly medium- to coarse-grained and can be very coarse grained to granule conglomerate. The grains are subrounded and sorted to well sorted. The matrix is mainly quartz with minor sericite and muscovite and locally is calcite.

Siltite, slate and phyllite are grey and olive grey and exist as minor thin interbeds to the quartzite locally and generally in the eastern exposures. They are much like rocks of the Yankee Belle and Midas formations.

Bedding characteristically is thick and indistinct, although in areas it is well displayed by interbeds of fine grained rock and by compositional variations in the matrix. Grading and crossbedding occur locally.

The thickness of the Yanks Peak Formation varies considerably throughout the area, generally thinning to the southwest (Fig. 8). It is more consistent within fault bounded sequences. The formation is greater than 300 m thick and is most diverse and coarsest grained on Kimball Ridge east of the Kimball Fault. West of the fault it is

Table 4. Yankee Belle Formation stratigraphic section on eastern Kimball Ridge (location; 52°55'N, 121°01"W).

Unit	Lithology	Thickness in metres Unit	Above base
38	Quartzite, grey, fine grained; 3 to 10 cm beds with minor interbeds of grey slate	24	693
37	Quartzite, light green, moderately fine; 1 to 1.5 m beds; thin interbeds to dull olive-brown siltite	18	669
36	Interbedded slate, light green to olive grey, and siltite, light green grey	38	651
35	Interbedded quartzite, light green, fine grained, slate, olive to olive brown, and minor olive siltite; 6 to 25 cm beds	25	613
34	Limestone, dark grey; approximately 1%, 1 to 2 mm dolomite crystals	3	588
33	Slate, olive, coarsening up section to olive-grey siltite	23	585
32	Limestone, grey, finely crystalline	2	562
31	Interbedded slate, grey, limestone, grey, and quartzite, brownish grey	2.5	560
30	Slate, green grey	3.5	557.5
29	Interbedded quartzite, brown weathering light green grey, fine grained, slate, light olive and brown weathering olive grey, and siltite, grey weathering grey; 1 to 35 cm interbeds with 2 to 4 cm beds common; quartzite is lensoid near top	18	554
28	Slate, grey	7	536
27	Interbedded limestone, grey and sandy limestone brown-grey weathering grey, finely crystalline	2	529
26	Quartzite, light grey, moderately fine; sharp contact with overlying limestone	5	527
25	Interbedded quartzite, light olive, fine grained, minor moderately fine grained, siltite, olive grey, and slate, olive grey and dark grey	45	522
24	Sandstone, brown weathering, grey, calcareous	2.5	477
23	Interbedded quartzite, green-grey, fine to moderately fine grained, siltite, grey, and slate, grey; finely laminated	49	474.5
22	Quartzite, white, moderately fine grained	3	425.5
21	Gradational sandy limestone to calc careous sandstone passing to quartzite, grey, fine grained up-section	8	422.5
20	Interbedded slate, dark grey, and quartzites, light green, fine grained; graded 2 to 15 cm beds	10	414.5
19	Interbedded slate, grey, and siltite, grey; 4 to 15 cm beds	20	404.5
18	Limestone, buff-grey weathering dark grey, finely crystalline, with 3 mm lamellae of sandy limestone to calcareous sandstone spaced 1 to 3 cm apart, top 6 m is relatively void of sand; 1 to 3 mm stylolite peaks perpendicular to bedding	15	384.5
17	Interbedded quartzite, light grey, slate, grey, and minor 10 to 20 cm beds of sandy limestone	58	369.5
16	Limestone, grey to orange weathering slightly greenish grey, in part sandy, finely crystalline	18	311.5
15	Interbedded quartzite, brown-grey, slate, dark grey, and sandy limestone, orange weathering grey	5	293.5
14	Limestone, buff weathering grey, finely crystalline minor interbeds of sandstone and slate, grey	28	288.5
13	Interbedded quartzite greater than slate much greater than sandy limestone, grey, minor quartzite, white moderately fine grained	46	260.5
12	Interbedded slate and quartzite, grey with approximately 20% sandy limestone 3 to 8 cm beds gradational up section after 5 m to limestone and sandy limestone, grey finely crystalline much greater than slate, grey	10	214.5
11	Poorly exposed interbedded quartzite light grey, fine grained, siltite, olive grey and slate, dark grey	14	204.5
10	Limestone, grey, finely crystalline, partially sandy	4	190.5
9	Poorly exposed, primarily interbedded quartzite grey, moderately fine; 3% 1 to 3 mm ankerite porphyroblasts; and slate, grey	54	186.5
8	Limestone, grey weathering moderately dark grey, finely crystalline, interbedded with calcareous sandstone and sandy limestone in 2 cm beds near top	5	132.5
7	Poorly exposed interbedded quartzite and slate, grey, changing up section to slate greater than quartzite	53	127.5
6	Limestone, grey, finely crystalline	11	74.5
5	Partially covered, interbedded slate and siltite, light green weathering olive grey to grey, coarsening up section to quartzite, light green-grey, moderately fine grained greater than slate, grey, minor white quartzite near top	32	63.5
4	Limestone, grey, finely crystalline, sandy, thin lamellae of sandstone as lenses in limestone	2.5	31.5
3	Interbedded siltstone and slate, olive grey; 3 to 6 cm beds	7	29
2	Limestone, grey, finely crystalline	3	22
1	Silty slate to slate, grey	19	19
	Underlying beds: Cunningham Formation		

Figure 7. Yanks Peak Formation conformably overlying Yankee Belle Formation on the hill northeast of Little River. (GSC 191004)

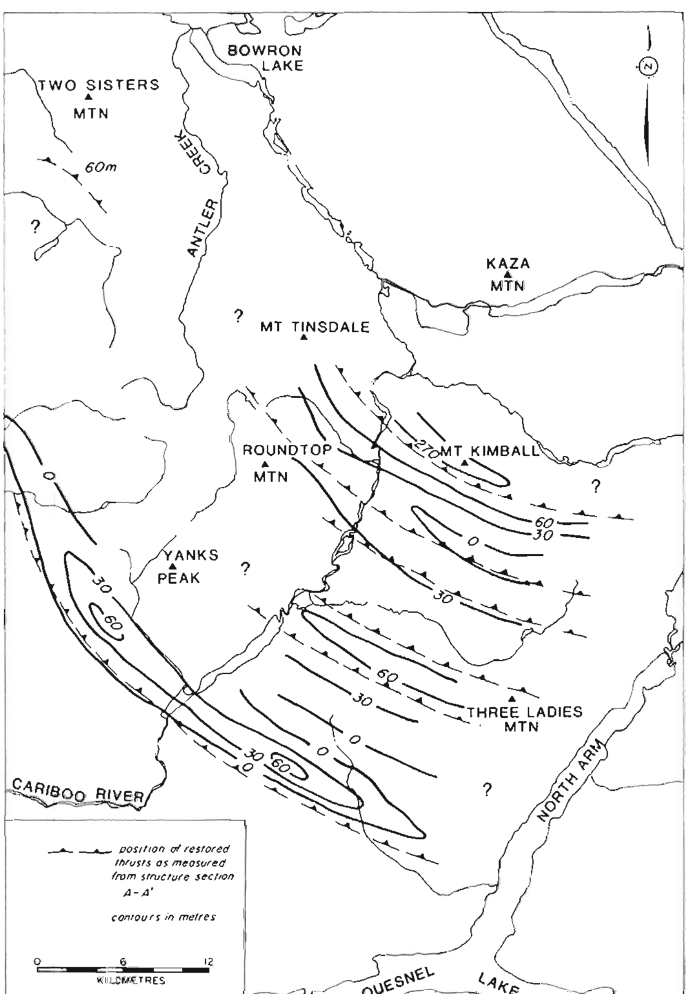


Figure 8. Palinspastic isopach map of the Yanks Peak Formation.

only 60 m thick, more homogeneous and finer grained. Stratigraphic differences across fault zones with the magnitude presented here are usually used as indicators of large displacement along the fault. The geometry of the Kimball Fault does not justify large dip-slip displacement. This implies rapid thickness and grain size changes of the Yanks Peak Formation at Kimball Ridge or strike-slip displacement on the Kimball Fault. The palinspastic isopach map of Figure 8 assumes dip-slip movement on the Kimball Fault for purposes of simplification. Westward through the Black Stuart Synclinorium the unit is mainly less than 40 m thick. Its average thickness increases slightly in the westernmost faulted sequence of the Cariboo Terrane passing through Roundtop Mountain southward to the ridge northeast of Little River. Holland (1954) and Sutherland Brown (1957,1963) suggested that the formation is as thin as 1 m in this area, however, the thinness is attributed to attenuation on fold limbs and in thrust faults.

For reference, a partial section on Mount Kimball (52°55'40''N, 121°3'02''W) of Yanks Peak Formation continuous with Yankee Belle Formation, can be used (Table 5).

Sutherland Brown (1963, p. 32) suggested from considerations of composition, thickness and sorting that the Yanks Peak Formation was deposited in a shallow marine basin. Young (*in* Campbell et al., 1973, p. 47-48) postulated that the quartzites are littoral deposits, separated by shallow marine fine clastics deposited during periods of transgression. Mansy (1970, p. 56) from comparisons of statistical grain size curves derived for the Yanks Peak Formation quartzite, concluded the Yanks Peak Formation Quartzites are fluviatile.

This study presents no data different from that presented by these three workers. Young's hypothesis comes from work covering more geographic area than either Mansy or Sutherland Brown and may therefore be more realistic. It is accepted here.

From geographic distribution of facies, Young postulated a south to southwest dispersal trend. The paleogeographic distribution, as approximately pictured in Figure 8, does not dispute Young's hypothesis. However, Young (1969, p. 105) argued that the Yanks Peak Formation quartzite of Summit Creek is a sandstone tongue representing "the westernmost progradation of the Yanks Peak Formation". With a south-southwest dispersal trend, quartzite of the Yanks Peak Formation farther south should be thinner and finer grained than at Summit Creek. The quartzite of the Roundtop Mountain area is as thick and thicker and as coarse and locally coarser than at Summit Creek. It is associated with white quartzite of the upper Yankee Belle Formation which is absent at Summit Creek. The quartzite of the Roundtop Mountain area could be a more proximal facies than Summit Creek, but this contradicts southerly dispersal trends, but not southwesterly dispersion or complex trends involving various current patterns. A complex pattern is required to explain the absence of quartzite between areas where thicknesses can be up to 60 m. A simple southwest dispersion would require gradual diminishing of quartzite to the southwest. This gradual diminishing is not the picture presented in Figure 8.

Table 5. Yanks Peak Formation incomplete stratigraphic section on Mount Kimball (location; 52°55'40"N, 121°3'02"W).

Unit	Lithology	Thickness in metres Unit	Above base
13	Quartzite, white, coarse grained to fine granule conglomerate, coarsens upsection, crossbeds in lower part of unit	45	280
12	Interbedded micaceous quartzite, grey, and well sorted white quartzite, moderately coarse grained	2	235
11	Quartzite, white, moderately coarse grained	10	233
10	Interbedded slate, olive-green-brown, silty, micaceous quartzite, brown weathering grey, moderately fine grained and well sorted quartzite, light pink, medium grained; 1 to 7 cm beds	4	223
9	Quartzite, olive and brown weathering grey, micaceous, moderately fine grained	4	219
8	Quartzite, white, well sorted, coarse- to granule-grained	28	215
7	Quartzite, grey, moderately fine grained; slightly sericitic	3	187
6	Sandstone, brown limonite cemented	.02	184
5	Interbedded quartzites, light green, well sorted, moderately fine and olive grey, micaceous medium grained	10	184
4	Conglomerate, white, granules of rounded to subrounded clear quartz with minor blue quartz	1	174
3	Interbedded quartzite, dark brown weathering grey-brown, micaceous, moderate to moderately fine grained, poorly sorted, limonite, slate, olive and minor quartzite, white, medium grained; gradational 1 to 30 cm beds with 3 to 10 cm beds common	90	173
2	Quartzite, grey, micaceous, medium grained; 15 to 50 cm beds	30	83
1	Quartzite, white, medium grained	53	53
	Underlying beds: Yankee Belle Formation		

Age and correlation. Campbell et al. (1973) considered the Yanks Peak Formation to be of lowermost Cambrian age and correlated it with the McNaughton Formation of the Rocky Mountains directly to the east. No further evidence on the age of the Yanks Peak Formation has been obtained and the conclusion of Campbell et al. is accepted here.

Midas Formation

The Midas Formation consists of slate, phyllite, siltite and quartzite. It can be differentiated from the Yankee Belle Formation in four ways: 1) the Midas Formation is not known to have limestone members in its basal part as does the Yankee Belle Formation; 2) the pelites of the Midas Formation are generally greyer and more platy where not metamorphosed to phyllite; 3) if present, the crossbedded Vic Sandstone with its lamellae of heavy minerals is distinct from any lithology in the Yankee Belle Formation; and 4) the Yankee Belle Formation coarsens upsection to the Yanks Peak Formation whereas the Midas Formation fines upwards to the Mural Formation. Where the Yanks Peak Formation quartzite is not present the contact between the Yankee Belle Formation and Midas Formation is difficult to define; however, in almost all cases there is at least a lens of white fine- to medium-grained quartzite at the horizon where the Yanks Peak Formation is expected. In isolated small outcrops distinction between Midas and Yankee Belle formations is uncertain.

Midas Formation has not been recognized north of lower Pleasant Valley Creek though it underlies much of the area to the south. The type locality of the Midas Formation is defined as the reference section described by Campbell et al. (1973) from the Dome Creek area (56°36'N, 121°01'W). The original type section defined by Holland (1954) at Yanks Peak is recommended to be abandoned. It does not contain the same unit of rocks as those called Midas Formation east of the Pleasant Valley Thrust. At the Dome Creek section the unit is 148 m thick and the thickness varies from 91 to 305 m throughout the Cariboo Mountains (Campbell et al., 1973). Thickness of the formation throughout the map area varies from 40 to 250 m. North of Cariboo River, the thickness is determined at two localities. The formation is approximately 40 m thick with the base unexposed at the junction of Pleasant Valley and Antler creeks, and about 45 m thick at Loskey Creek. Thicknesses south of Cariboo River can be determined for three localities. On the hill 5 km southeast of Black Stuart Mountain, it is approximately 50 m thick, bounded on top by archaeocyathid-bearing Mural Formation limestone and at the base by a thin quartzite lens assigned to the Yanks Peak Formation. On Anderson Ridge,[1] poorly exposed Midas Formation may attain 250 m in thickness. On the south slopes of Kimball Ridge a partly covered succession is approximately 130 m thick.

[1]Anderson Ridge is the east-trending ridge south of upper Kimball Creek and southwest of upper Connection Creek. This name was used by Lang (1938).

The lower contact of the Midas Formation is conformable and gradational over about a metre interval with the Yanks Peak Formation quartzite.

On Roundtop-Middle Ridge the base of the Midas Formation is a dark grey limonitic siltite resting sharply on medium grained dark grey quartzite of the Yanks Peak Formation. South of Cariboo River, in fault packages east of those which include the Roundtop-Middle Ridge, the base of the Midas Formation is usually an impure quartzite or siltite similar to beds found in the underlying Yankee Belle Formation.

Slate and phyllite of the Midas Formation are grey and grey olive, much like those of the Yankee Belle Formation; generally not as olive. There are some localities where they are dark weathering and the primary one is described in the measured section from 4.5 km southeast of Black Stuart Mountain. Outcrops on Anderson Ridge, continuous with this section, are lighter weathering and equivalent strata north of Kimball Creek contain only minor dark grey phyllite.

Siltite and fine grained quartzite are light grey to dark grey, flecked with less than 1% hematite grains and weather light olive grey to rusty brown. They are sorted to poorly sorted with a micaceous matrix, interbedded on a 3 to 30 cm scale with grey pelite, and may grade normally upward into that pelite. They are not confined to any particular part of the formation, however, they are more common near the base and top.

Vic Sandstone. It is proposed that the Vic Sandstone be a formal member of the Midas Formation. It consists of olive-grey to light grey quartzite and is distinguished from other quartzite of the Midas Formation by being coarser grained and containing crossbedding. It occurs in most areas underlain by the formation. A reference locality for the Vic Sandstone is approximately 3.5 km upstream along Cunningham Creek from its junction with Cunningham Pass Creek (52°57'10"N, 121°21'40"W). This locality was described by Holland (1954, p. 21, Plate IVA) and Sutherland Brown (1957, p. 27, Plate IVA). Geoff Hodgson (personal communication, 1978) named the unit after the Vic claim block in which it lies. The thickness at the reference locality is approximately 5-8 m and varies to 12 m throughout the area.

The quartzite is medium grained, sorted to poorly sorted and micaceous. Its most distinguishing characteristic is its 10 to 50 cm trough and planar crossbeds marked by concentrations of ilmenite and other heavy minerals (Fig. 9). The quartzite is commonly graded normally from medium grained olive grey to fine grained or silty, grey to dark grey. The mineralogy and grain size as determined with a microscope are listed in Table 6, with those of several other rocks of the Midas Formation.

Table 6. Mineralogical compositions (volume per cent) and quartz grain size (microns) of samples from the Midas Formation (from point-count data)

	1	2	3	4	5	6	7	8	9	10
Quartz	22.2	40	56	10.2	54.5	52.4	5.8	51	81.7	68.5
Potassium feldspar	.5		.1		.8			4.2	.9	8.6
Plagioclase		.8			.7			1.8	1.4	.3
Muscovite	.6	1.4	.2		.6	.3		2.4	.1	.2
Biotite					.1			.2		
Chlorite				.7	.2					
Opaques	29.8	5.4	3.9	3.3	10	14.1	1.8	6.2	3.4	4.3
Zircon	1.3	.3	.6		1	.1		.6	.3	
Epidote			.1						.1	
Sphene								.1		
Tourmaline	.5									
Apatite										.2
Secondary minerals										
Sericite and clays	38.9	7.5	39.1		32.2	33.1	92.4	22.4	5.1	17.9
Chlorite		44.5						5.7	3.7	
Biotite								5.4		
Matrix				85.8					3.3	
Average quartz size ($x10^{-3}$ mm)	70	40	70	40	90	80	40	55	70	80

1 Siltite from junction of Antler and Pleasant Valley creeks.
2 Siltite from Loskey Creek area.
3 Fine grained quartzite (Vic) 3.5 km upstream along Cunningham Creek from junction with Cunningham Pass Creek.
4 Phyllite from headwaters of Roundtop Creek.
5 Fine grained quartzite (Vic) from saddle between Roundtop and Middle mountains.
6 Very fine grained quartzite from ridge south of Middle Mountain.
7 Slate from same locality as 6.
8 Very fine grained quartzite (Vic) from ridge northeast of Middle Mountain.
9 Fine grained quartzite from 4.5 km southeast of Black Stuart Mountain.
10 Fine grained quartzite from top of section 4.5 km southeast of Black Stuart Mountain.

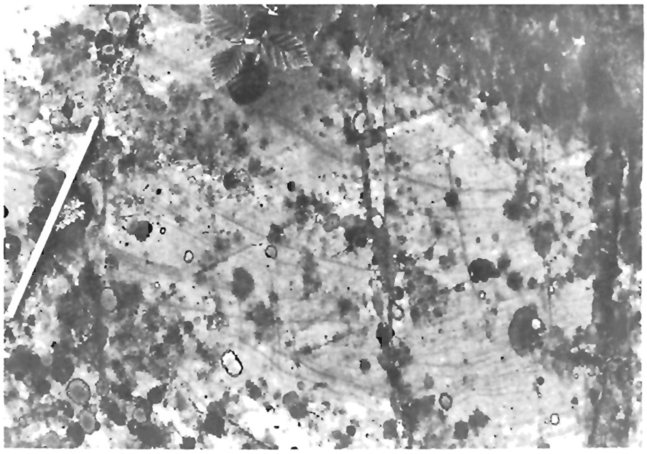


Figure 9. Trough crossbedding in Vic Sandstone member of the Midas Formation. (GSC 191005)

The Vic Sandstone can be found as far north as the junction of Antler and Pleasant Valley creeks where it is approximately 10 m thick and was mapped as Yanks Peak Formation by Campbell et al. (1973). There it shows very little crossbedding but is graded in much the same way as at the reference locality. As Sutherland Brown (1957, p. 27) reported, the Vic Sandstone member can be traced southeastward some 10 km. Approximately 8 m of Vic Sandstone are exposed on the northeast-trending ridge from Middle Mountain. South of Cariboo River, 4 m or less, approximately 20 m from the top of the Midas Formation and 25 m from the base, are found in the section 5 km southeast of Black Stuart Mountain. The Vic is also found in the section on the southern slope of Kimball Ridge where it is 8 m thick and occurs 70 m above the base and 60 m from the top of the Midas Formation. Large scale crossbedded fine quartzites are also found on Anderson Ridge and also beneath the dark grey to black phyllite and siltite reported in the section 4.5 km southeast of Black Stuart Mountain.

Several sections of Midas Formation from south of Cariboo River are described below. They are presented to give an indication of Midas Formation stratigraphy in this area. Two of these sections have been described by Mansy (1970), but these descriptions differ somewhat from his interpretation. The section 5 km southeast of Black Stuart Mountain is different because the stratigraphy and structure of this area have been considerably changed from that of Mansy. His interpretation of the overlying Mural Formation is agreed upon and is now supported by fossil evidence, however, the lower part of the section involves a transition through Yankee Belle Formation which is thrust onto Yanks Peak Formation. Mansy (1970, p. 49-50) provided mineral compositions for quartzites of his sections.

The Midas Formation section 5 km southeast of Black Stuart Mountain (52°51'40"N, 121°05'W) provides a good reference locality (Table 7).

Table 7. Midas Formation stratigraphic section 5 km southeast of Black Stuart Mountain (location, 52°51'40"N, 121°05'W).

Unit	Lithology	Thickness in metres Unit	Above base
	Base of Mural Formation light grey moderate to finely crystalline limestone.		
	Covered-probably green phyllite with minor limestone in 1 cm thick lenses	5	48.85
12	Interbedded quartzite, grey green, fine grained, and slate, green grey and grey green	3.5	43.85
11	Quartzite, pinkish weathering-greenish grey, fine grained parts in 1 cm thick layers	.5	40.35
10	Slate, light brown and green grey, weathering green grey; grades into overlying unit	6	39.85
9	Siltite, grey, minor green grey slate grades into overlying unit	1	33.85
8	Quartzite, green to light green, slightly micaceous, fine grained with coarser fraction near top	3.5	32.85
7	Interbedded quartzite, green and light green, fine grained; slightly sericitic, 1 cm partings, trough crossbedding and slate, grey and green, 40 to 30%, grades into overlying unit, irregular contact with underlying unit	4	29.35
6	Slate, green grey, green, moderately dark grey and grey, minor quartzite, green grey, very fine grained; sericitic	6	25.35
5	Phyllitic slate, greenish-grey and brown weathering dark grey	12.7	19.35
4	Quartzite, grey, very fine grained; fine lamellae	.1	6.45
3	Interbedded phyllite, grey, and quartzite, grey, fine grained	5	6.35
2	Interbedded quartzite, grey green, fine grained, and phyllite, dark grey; in 5 mm beds	1	1.35
1	Quartzite, dark rusty purple and pink and green weathering light pinkish grey, fine grained	.35	.35
	Yanks Peak Formation		
	Quartzite, pinkish-light green grey, medium grained; 30 m lens exposed	1.2	1.2
	Underlying beds: Yankee Belle Formation		

Sutherland Brown (1957, p. 29) suggested two environments of deposition for the Midas Formation. For the Vic Sandstone he proposed a continuation of the shallow marine environment represented by Yanks Peak Formation quartzite. Above the Vic, deposition would have occurred in deeper water with a restricted circulation. However, this restricted deep-water condition is invoked to account for extensive deposits of black slate and phyllite now included in the Harveys Ridge succession of the Barkerville Terrane.

Young (*in* Campbell et al., 1973) postulated a calm, restricted marine deposition based on dark fine clastics and evidence of organic material. He suggested deposition on a deep shelf or basin because the Midas Formation graded westward into thick black shales. The Midas Formation does not consist of thick black shale in the map area and exposures farther west are not Midas Formation, as mentioned above. Mansy (1970) used similar arguments for restricted marine deposition but also stated that the fine grained quartzites have grain size distributions which are characteristic of fluviatile origins.

Deposition associated with restrictive marine conditions may not have been continuous in the west, but rather involved isolated areas on the margins of the restricted basin, outlined by Young, to the northeast. Midas Formation deposition may have continued in the environment established during Yanks Peak Formation time until after Vic Sandstone time, as suggested by Sutherland Brown (1957). Then restrictive conditions may have begun in the east with some areas of the west also affected. Another possibility is that the lower Midas Formation, including the Vic is synchronous with the upper Yanks Peak Formation farther east. The onset of Midas Formation deposited under restrictive conditions would be synchronous in both the east and west. Western deposition during this time would be into quiet water but both restricted and nonrestricted environments.

Age and correlation. Young (1969, *in* Campbell et al., 1973) assigned the Midas Formation to early Lower Cambrian from evidence of abundant trace fossils, from the gradation into early Lower Cambrian Mural Formation, and from the correlation of Midas Formation with uppermost McNaughton Formation which contains cruziana-like trace fossils. No new evidence has been obtained and this age assignment is accepted.

Mural Formation

The Mural Formation consists mainly of limestone and lesser amounts of dolostone. To the east in Cariboo Mountains it also contains shale. Isolated, unfossiliferous, exposures of Mural Formation limestone can be confused with limestone of the Yankee Belle, Cunningham and Isaac formations of the Cariboo Terrane and with most limestone units of the Snowshoe Group of the Barkerville Terrane. It is distinguished by its stratigraphic position, occurrence of archaeocyathid fossils and from its contact relations with bounding units.

The Mural Formation is widespread, occurring throughout the Cariboo Terrane within the area and is correlated eastward into the Rocky Mountains where it was first named. Campbell et al. (1973) described the formation from the Rocky and Cariboo mountains and presented a reference section at the head of Dome Creek. The thickness increases westward from 150 m near the Rocky Mountain Trench in Cariboo Mountains to 730 m at Turks Nose Mountain which borders the eastern edge of the map area (Campbell et al., 1973, p. 51). Thickness variations within the map area are not well known, because the Mural Formation is in most places wholly or partly bevelled below an unconformity at the base of the Black Stuart Group, and because of structural complications. It is 150 m thick on the south slope of Kimball Ridge where it is stratigraphically bounded by the underlying Midas and overlying Dome Creek formations; much thinner than at Turks Nose Mountain 15 km to the northwest. It is 250 m thick on Anderson Ridge where it is unconformably overlain by the Black Stuart Group.

The lower contact with the Midas Formation is conformable and mainly sharp. The uppermost 2 to 6 m of Midas Formation commonly contain 1 to 2 cm thick lenses of limestone which thicken upsection to 2 to 8 cm beds directly below the massive limestone of the Mural Formation. The Midas Formation pelite becomes more intensely green near the contact and at the headwaters of Roundtop Creek it is interbedded with purple pelites. The character of transition from Midas to Mural Formation can be used to differentiate Mural from Cunningham Formation.

The Isaac-Cunningham transition differs from the Midas-Mural transition in two ways: 1) the Isaac is distinct from the Midas Formation because it is predominantly black pelite and calcareous pelite whereas the Midas Formation is mostly green and grey pelite interbedded with grey or light grey siltite and quartzite; and 2) the Isaac-Cunningham transition is gradational over approximately 100 m, with an upward increase in brown weathering dark grey limestone grading to the grey weathering grey limestone of the Cunningham Formation, in contrast to the rapid transition of 2 to 6 m between Midas Formation, which may have thin carbonate lenses developed adjacent to the contact, and the massive to 10 cm bedded grey limestone of the Mural Formation.

Limestone of the Mural Formation is grey to dark grey where it is relatively unaltered and dark grey to white where altered. It weathers mainly grey but can be buff, white or orange. It generally consists of 0.05 to 0.5 mm metamorphic crystals of calcite with minor amounts of quartz and heavy mineral silt. Bedding is preserved locally (Fig. 10) with features such as oolites and archaeocyathid and trilobite fragments in limestone as on Kimball Ridge. Sedimentological detail is usually lost due to recrystallization, being replaced by 0.5 to 2 cm colour-defined laminations which appear in most cases to parallel

Figure 10. Bedding style of Mural Formation on Kimball Ridge. (GSC 191006)

original bedding. In places these bands show layer-parallel, small-scale isoclinal folds. They are also found in limestones of the Cunningham and Yankee Belle formations. Marble has developed along some faults through Mural Formation limestone. It is mostly cream to white with 0.5 to 2 mm crystals of calcite.

Dolostone weathers grey, is mottled grey and white on fresh surfaces and is coarsely crystalline. It is composed of euhedral to subhedral dolomite crystals and has minor amounts of porosity. It is most abundant in one outcrop area at Cunningham pass. The dolomitization is a late feature associated with deformational controlled replacement as argued for the dolostone of the Cunningham Formation.

Minor amounts of shale are included in the Mural Formation on Kimball Ridge and Turks Nose Mountain as described by Sutherland Brown (1963).

Age and correlation. The Mural Formation is Lower Cambrian as determined from included trilobites and archaeocyathids. Trilobites ranging from early to late Lower Cambrian were identified previously from the reference section in the Cariboo Mountains (W.H. Fritz *in* Campbell et al., 1973, p. 51-53) and from Turks Nose Mountain and the hill east of Bowron Lake (V.J. Okulitch *in* Sutherland Brown, 1963, p. 28). Archaeocyathids of the Lower Cambrian were identified from Turks Nose Mountain, Iltzul Ridge and the hill just east of Bowron Lake (V.J. Okulitch *in* Sutherland Brown, 1963, p. 27).

Three previously unrecorded fossil localities have been found within the Mural Formation south of Cariboo River. On Kimball Ridge a thin bed of orange-weathering limestone contains poorly preserved archaeocyathids and trilobite fragments (52°54'35"N, 121°02'42"W). Exact stratigraphic position is unknown due to structural contacts but the occurrence is probably in the lower middle part of the Mural Formation. Abundant well preserved archaeocyathids were found 5 km southeast of Black Stuart Mountain at an elevation of 1860 m (6100 feet; 52°51'53"N, 121°04'57"W). These archaeocyathids stand out in relief displaying features of their wall structures. Although several species appear to be represented, detailed identification has not been done. The archaeocyathids of this locality occur in a horizon beginning approximately 20 m above the base of the Mural Formation and continue upsection for approximately 10 m. Deformed archaeocyathids occur 4.2 km south-southwest of Black Stuart Mountain (52°51'22"N, 121°08'54"W) in a tan and light grey weathering light grey, finely crystalline limestone. The fossils have been flattened parallel to a well developed cleavage. Relationships to stratigraphic contacts are unknown.

On lithological similarities alone, the carbonate horizon overlying the Midas Formation is the Mural Formation of northern Cariboo Mountains. From paleontological evidence the formation is Lower Cambrian in the map area and therefore is in part biostratigraphically correlative with the Mural Formation of northern Cariboo Mountains. This agrees with the correlations of the Mural Formation between Cariboo Mountains and the map area as presented by Campbell (1978), Campbell et al. (1973), Mansy (1970), Young (1969), and initially by Campbell (1967).

Campbell et al. (1973) correlated Mural Formation of the Cariboo and Rocky mountains with Sekwi Formation, District of Mackenzie; Badshot Formation, Selkirk Mountains; and Donald Formation of the Dogtooth Mountains. No information collected disputes these correlations and they are accepted here.

Dome Creek Formation

The Dome Creek Formation was introduced by Campbell et al. (1973) to include a sequence of dark shale, siltstone and minor limestone overlying the Mural Formation in northern Cariboo Mountains. At its type locality at the headwaters of Dome Creek the formation is up to 1740 m thick, though it may be tectonically thickened (Campbell et al., 1973, p. 54).

The Dome Creek Formation is nearly everywhere absent in the area due to truncation by the unconformity at the base of the Black Stuart Group. The known occurrences are at Kimball Creek and on the logging road along Cunningham Creek. Mansy (1970, p. 64-66) described it as the post-Mural succession on the north bank of Kimball Creek where it consists of 100 to 120 m of black slate. Sutherland Brown (1963) mapped the trilobite-bearing shales of Kimball Ridge as Cunningham Formation, and Lang (1938) mapped them as an unnamed Lower Cambrian formation. South of Kimball Creek on the north limb of the Black Stuart Syncline, approximately 80 m of thinly interbedded grey slate and brown-weathering grey limestone overlying Mural Formation limestone and underlying Black Stuart Group dolostone represents the base of the Dome Creek Formation, eroded away in other parts of the area in pre-Black Stuart Group time. Mansy (1970) also mapped this horizon as Dome Creek Formation.

Age and correlation. Lang (1938, 1947), Sutherland Brown (1963) and Mansy (1970) described Lower Cambrian trilobite collections taken from the shales overlying the Mural Formation on Kimball Ridge. Mansy reinterpreted the collections of Lang (1938, 1947) and Sutherland Brown (1963) as being from the Dome Creek Formation.

Lang (1938, 1947) found the following fossils which were identified by C.E. Ressor who assigned them an Early Cambrian age, except for the new genus related to Kootenia which may be Middle Cambrian;

Paedeumias sp.
Kootenia sp.
Salterella sp.
Bonnia sp.
New genus related to *Kootenia*
New genus of Ollenellid Trilobite

Sutherland Brown (1963) reported:
Bonnia sp.
from the northwestern locality and:
Ogygopsis klotzi Rominger
from the southeastern locality. Determinations were made by V.J. Okulitch who assigned the *Bonnia* sp. to the upper Lower Cambrian and the *O. Klotzi* Rominger to the Middle Cambrian.

Mansy (1970) collected:
Ogygopsis sp.
Bonnia sp.
from the northwestern locality and:
Ogygopsis sp.
from the southeastern locality. Determinations were made by W.H. Fritz who assigned them a late Early Cambrian age.

Northwest of Kimball Ridge, in the canyon of the Cariboo River, an area not mapped for this report, Sutherland Brown (1963, p. 28-29) recovered fossils from a "black fetid limestone". They were identified by V.J. Okulitch as:
Obolus sp.
Glossopleura cf. *Stenorhacis* Rasetti
Paedeumias (Olenellus) gilberti
The brachiopod *Obolus* is Middle to Late Cambrian. *Glossopleura* cf. *Stenorhacis* Rasetti is Middle Cambrian and *Paedeumias (Olenellus) gilberti* is Early Cambrian. This faunal assemblage is probably from Dome Creek Formation. Campbell et al. (1973) reported upper Lower Cambrian and Upper Cambrian fossils from the type section in Cariboo Mountains and only one possible occurrence of a Middle Cambrian *Ogygopsis* sp. They concluded that the Middle Cambrian may be represented by a lacuna (see Campbell et al., 1973, p. 12). Considering the presence of two Middle Cambrian trilobite indentifications from the local area such a lacuna may not extend westward beyond the Cariboo Mountains.

Campbell et al. (1973, p. 56) suggested the Dome Creek Formation is a basinward restricted marine facies synchronous with the eastern Cambrian shelf facies of the Rocky Mountains. Similar basinal facies transitions occur in the Upper Cambrian of northern British Columbia where the shale facies of Kechika Formation correlate with those of the Dome Creek Formation (*see* Cecile and Norford, 1979, and Gabrielse, 1975, for discussions of Kechika Formation and correlative rocks in northern British Columbia).

Black Stuart Group

The Black Stuart Formation as originally mapped by Mansy (1970) is here elevated to group status because it contains distinct units, three or more of which are separated by unconformities and because it spans the Ordovician to Mississippian. The Devono-Mississippian Waverly, Guyet and Greenberry formations are placed in the Black Stuart Group because they intertongue with it and/or are the same age.

The Black Stuart Group is divided into three informal units (following Mansy, 1970, and Campbell et al., 1973: chert-carbonate; black pelite, which both underlies and overlies the chert-carbonate unit; and sandstone, which overlies the other two units) and into three formations.

The dolostone and chert breccia of the chert-carbonate unit are unique to the Black Stuart Group. The black phyllite and argillite is generally more siliceous than other pelites of the Cariboo Terrane but similar varieties occur in the Barkerville Terrane. The association of the Black Stuart Group with volcanic rocks of the Waverly Formation and conglomerate of the Guyet Formation distinguishes it from older units.

The Black Stuart Group primarily underlies the area southeast of Antler Creek, being best exposed and most diverse near Black Stuart Mountain. Black pelite of the group continues northwest in a thin belt to Big Valley Creek. The Black Stuart Group was first named and described by Mansy (1970) as a formation in the Black Stuart Mountain area. Campbell et al. (1973) recognized equivalent strata to the north of Cariboo River and mapped them as far north as Antler Creek. The only other locality where Black Stuart Group is known is at Haggen Ridge northeast of the map area. Mansy and Campbell (1970) reported a thickness of approximately 850 m in the type area. Though Campbell et al. (1973) did not report any total thickness they considered the group to be less than 850 m.

South of Cariboo River, the Black Stuart Group is probably less than 500 m thick. In the Cunningham Creek area, southeast of Mount Tinsdale, the group may be thicker, but this may be due to structural complications.

Chert-carbonate unit

The chert-carbonate unit consists of various types of dolostone, dolostone breccia and conglomerate and chert

Figure 11. Screenwork replacement of light coloured chert in the chert-carbonate unit of the Black Stuart Group on Kimball Ridge. (GSC 191007)

breccia. It is discontinuous, occurs at the base of the formation and has several facies variations throughout the area. These variations will be described from east to west for: Kimball Ridge; the north limb of the Black Stuart Syncline; and Anderson Ridge and the south limb of the Black Stuart Syncline.

On Kimball Ridge the chert-carbonate unit is primarily dolostone; grey to chalky weathering moderately light grey to white and fine to medium crystalline. Locally chert has replaced the dolostone. The chert is very light grey to white and penetrates the dolostone as irregular layers or screens (Fig. 11).

On the north limb of the Black Stuart Syncline, the chert-carbonate unit consists of mainly dolostone breccia and lesser amounts of the same dolostone as on Kimball Ridge. Chert replacement of the dolostone is different, occurring more as regular layers and irregular patchworks. The dolostone breccia consists of light grey dolostone breccia fragments, from 2 to 20 mm across, fine to medium crystalline, that compose up to 40% of the rock. The matrix is sulphurous limy-dolomite to dolomitic limestone; grey to moderately dark grey and finely crystalline. The dolostone breccia undergoes replacement by chert in the same manner as the fine grained dolostone. North of Cariboo River at Limestone Creek the dolostone breccia contains clasts of angular dolostone, chert, cherty pelite and minor white sandstone and quartz sand (Fig. 12). The quartz sand varies in concentration from near 0 to 40%; locally forming calcareous sandstone. Sand-size quartz grains are rounded and average 1 mm across. There is a marked bimodality of quartz grain size with medium-coarse sand being distinct from the silt-size. Campbell et al. (1973, p. 56-57) described coral-bearing limestone from the same outcrop horizon but this was not found. Thickness of the dolostone horizon varies from 1 to 60 m, possibly being thicker on Kimball Ridge where the top is unseen.

Figure 12. Dolostone breccia of the chert-carbonate unit of the Black Stuart Group near the mouth of Limestone Creek. (GSC 191008)

On Anderson Ridge, through the keel of the Black Stuart Syncline, the chert-carbonate unit is mostly chert breccia (Fig. 13) and minor conglomerate. The chert breccia varies from near white to being streaked in shades of grey. It consists of 30% angular chert fragments, 5 to 30 mm across, commonly elongate, defining a prominent lineation parallel to the regional northwest-trending fold axis. The matrix is chert which may be darker or lighter than the clasts. The chert breccia appears to be a replacement of the dolostone breccia and can be up to 100 m thick, but is commonly less than 40 m thick. Conglomerate overlies the chert breccia on Anderson Ridge, 5 km east-southeast of Black Stuart Mountain. It is 8 m thick and consists of 1 to 10 cm clasts of dolostone, white and moderately finely crystalline and grey chert. Clasts are subrounded to angular and vary greatly in concentration. The matrix is composed of quartz sand; greenish brown weathering dark grey.

Bioclastic limestone directly overlies the Lower Cambrian Mural Formation on Anderson Ridge and is included in the chert-carbonate unit. It is grey and light grey weathering grey with very poorly preserved bioclastic debris, some of which resembles *Tentaculitids* and shell fragments. Locally preserved are 5 to 15 cm beds differentiated by various concentrations of debris. In other places augen shapes lie in the metamorphic foliation and may be sedimentary or tectonic breccia fragments (Fig. 14). The limestone may be the one from which Lenz (1977) collected Silurian brachiopods.

North of Antler Creek and southwest of Waverly Mountain, the chert-carbonate unit may be represented

Figure 13. Chert breccia of the chert-carbonate unit of the Black Stuart Group in the keel of the Black Stuart Syncline. (GSC 191009)

Figure 14. Limestone breccia, probably tectonic, possibly of the Lower Cambrian, at the base of the chert-carbonate unit of the Black Stuart Group in the keel of the Black Stuart Syncline. (GSC 191010)

by poorly exposed tan weathering cream moderately crystalline dolostone overlying grey limestone of the Mural Formation. Chert breccia similar to that found at the base of Black Stuart Group south of Cariboo River is poorly exposed (possibly float) in the creek which drains from the south of Waverly Mountain to Pleasant Valley Creek.

Black pelite unit

The black pelite unit consists primarily of dark grey to black shale, slate, cherty argillite, siltstone and phyllite and lesser amounts of dark grey limestone, dolostone and sandy dolostone and black chert. The black pelite can be confused with those of the Isaac and Dome Creek formations of Cariboo Terrane, those of the Harveys Ridge and Hardscrabble successions of Barkerville Terrane and those of the Triassic of Quesnel Terrane.

The black pelite unit can be traced from the type area at Black Stuart Mountain north to Big Valley Creek. It is involved in several thrust sheets and unconformably rests on most older formations excluding the Cunningham Formation. The thickness of the black pelite unit is unknown but is estimated to be less than 500 m.

The pelite is slate where it contains Ordovician graptolites and is mainly siliceous phyllite and argillite where it overlies the Lower Devonian chert-carbonate unit. Siltstone is dark grey and thinly interbedded with the pelite. Limestone is dark grey and black, finely crystalline (secondary); occurs in 1 to 5 cm beds in pelite and in one locality has yielded conodonts. A single locality of light grey oolitic and intraformational conglomeratic limestone occurs on the logging road along Loskey Creek. Dolostone has a restricted occurrence, is dark grey, finely crystalline and in places sandy. Its bedding character is like that of the limestone. Chert is black to grey, pure to muddy and thinly laminated. Locally the chert is secondary where it has replaced amphiporal limestone (Fig. 15).

Sandstone unit

The sandstone unit of the Black Stuart Group consists of grey siltite and fine grained quartzite interbedded with grey and dark grey pelites, light pink to white moderately fine grained quartzite and what appear to be silicified carbonates. The quartzite, siltite and pelite are unique in the Black Stuart Group but resemble those from the Yankee Belle and particularly the Vic Sandstone of the Midas Formation.

Figure 15. *Amphipora* in the black pelite unit of the Black Stuart Group near Black Stuart Mountain. Canadian nickel for scale. (GSC 191011)

The sandstone unit is found south of Cariboo River on Black Stuart Mountain and 3.5 km east-southeast and 2.7 km southeast of the mountain. The thickness of the unit is less than 30 m but nowhere is the top seen. The lower contact is sharp and conformable onto black pelite and siltite of the black pelite unit.

The predominant quartzite and siltite is brownish-olive-grey weathering grey with thin dark grey lamellae, fine grained and micaceous (Fig. 16). Thin section examination of an impure quartzite from Black Stuart Mountain revealed the rock as composed of mostly subrounded quartz and a matrix of the secondary minerals calcite, dolomite, sericite and clay with minor quantities of plagioclase, potassium feldspar, detrital and secondary opaques, zircon and tourmaline. Bedding with grey pelite is from 1 to 20 cm thick and commonly graded. Local sedimentary features include load and flame structures, fine scale lensing and ripples (1 cm amplitude, 3 cm wavelength). Quartzite beds are very evenly bedded with constant thickness maintained for at least 30 m. At Black Stuart Mountain a bed of quartzite is spotted with pink lenticles of medium grained quartz grains. The sandstone unit exposed 2.7 km southeast of Black Stuart Mountain includes white moderately fine grained clean quartzite and black to white chert (Fig. 17). White quartzite and black impure chert form 2 cm interbeds. The chert has irregular patches of white. Flames of chert protrude upward into the white quartzite, suggesting loading of the quartzite onto a soft less dense horizon. The black impure chert is probably a silicified micrite, which would account for the density differential required for production of the load and flame structures associated with overlying quartzite.

Age and correlation. Mansy and Campbell (1970) and Campbell et al. (1973, p. 57) reported the age of the chert-carbonate unit of the Black Stuart Group as Lower Devonian from identification of corals and conodonts. Mansy (1970, p. 70) reported some organic looking small rods from beneath the sandy unit 2.7 km southeast of Black Stuart Mountain. T.E. Bolton (*in* Mansy, 1970) speculated that they may be a type of coral found generally in the Upper Ordovician. Lenz (1977) described brachiopods from limestone breccia blocks taken from the basal unit at its easternmost exposure on Anderson Ridge (52°53'N, 121°03'W). Lenz proposed these brachiopods as Llandoverian or Wendlockian.

Figure 16. The bedding style of fine grained sandstone of the sandstone unit of the Black Stuart Group. (GSC 191012)

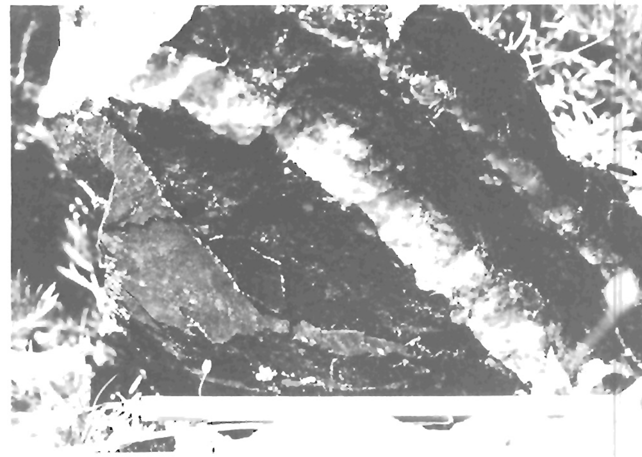


Figure 17. Chert of the sandstone unit of the Black Stuart Group near Black Stuart Mountain. (GSC 191013)

The black pelite unit has yielded Ordovician and Devonian fossils; ages that are older and younger than the chert-carbonate unit. The black pelite unit must therefore contain at least two separate sequences interrupted by deposition of the chert-carbonate unit. The two sequences were not divisible during mapping.

The oldest black pelite unit has yielded deformed graptolites from dark grey slate on the ridge 2.5 km south-southeast of Black Stuart Mountain (52°52'27.5"N, 121°06'59"W; GSC Locality C-53427). B.S. Norford identified them as *Climacograptus?* sp., *Dicellograptus* sp. and *Orthograptus* sp. indicating late Middle or Late Ordovician (Caradoc or Ashgill). The exact position of this collection in the stratigraphic sequence is unknown. The base of the dark grey slate is thrust onto the chert-carbonate unit and onto Cambrian Mural Formation. The fossil locality can be linked to components of the black pelite unit underlying the sandstone unit of Black Stuart Mountain.

The chert carbonate unit has yielded a conodont fragment from the same locality at Limestone Creek (52°58'59.4"N, 121°11'10.7"W; GSC Locality C-82852) as reported by Campbell et al. (1973). It was identified by T.T. Uyeno as *Icriodus woschmidti* Ziegler, and assigned to the latest Silurian to Early Devonian (Pridolian to Lochkovian). This is in complete agreement with ages determined for the fauna collected by Campbell (Campbell et al., 1973).

The younger black pelite unit has yielded two Devonian fossil localities. One of them occurs beneath the sandstone unit 2.8 km southeast of Black Stuart Mountain in black silicified carbonate. It contains 10 to 20% white elongate objects suggested by J.E. Klovan and F. Stoakes (personal communication, 1978) to be possibly *Amphipora* which range from Late Silurian through the Devonian. The second locality is 850 m north of the mouth of Limestone Creek (52°59'6.3''N, 121°11'3.7''W; GSC Locality C-82851) where conodonts were extracted from a thin black micritic limestone bed in black slate. The rocks are exposed near the base of the cliff along the Cariboo River bank. The conodonts were identified by T.T. Uyeno as fragments of the *Polygnathus varcus* Stauffer group of Klapper et al. (1970). He assigned them to the range Lower *Polygnathus varcus* subzone to Lowermost *Polygnathus asymmetricus* zone, that is Late Middle Devonian.

In summary the informal units of the Black Stuart Group range from Late Ordovician to Late Devonian and possibly younger. Upper Ordovician graptolitic pelite may be separated from the Lower Devonian chert-carbonate unit by a Silurian unconformity. The Middle and Late Silurian brachiopods described by Lenz (1977) may be from a limestone unit locally preserved beneath the Lower Devonian chert-carbonate unit. The black pelite overlying the chert-carbonate unit could span the interval from Early to Late Devonian. The sandstone unit is Upper Devonian or younger, because it overlies the black pelite unit.

The oldest rocks of the Black Stuart Group, the graptolitic slates of the black pelite unit, correlate with Ordovician graptolitic pelites of the Road River Formation from northern British Columbia. Gabrielse (1975, p. 11) reported southernmost Road River facies equivalents in the Fort Grahame map area. Northward from Fort Grahame map area Road River Formation equivalents can be traced into the Selwyn Basin of Yukon Territory (Gabrielse, 1975). Detailed stratigraphy of Road River Formation has recently been described by Cecile and Norford (1979) for a part of northern British Columbia and by Gordey (1978, 1979) for the Summit Lake area in Northwest Territories and Yukon. In southern British Columbia correlations are less certain.

The Lower Devonian chert-carbonate unit could correlate with Lower Devonian carbonates of northern British Columbia. The Muncho-McConnel Formation (Taylor and Stott, 1973; Taylor et al., 1979) and related calcareous sandstone and limestone debris flows (Taylor et al., 1979) in the Ware map area and the "tapioca" sandstone sequence in McDame and surrounding area (Gabrielse, 1963) are examples of possible correlative quartz-sand-rich carbonate of the Lower Devonian. In contrast to them the chert-carbonate unit has only local examples of calcareous quartz sand and sandy limestone from Limestone Creek and one locality on Anderson Ridge. Besides quartz the Limestone Creek outcrop also has chert breccia, dolostone breccia, cherty pelite and rounded sandstone clasts which is not characteristic of the Lower Devonian occurrences noted in northern British Columbia.

The younger black pelite unit may range into the Mississippian and is in part correlative with the Earn Group (equivalents of which are informally called the "black clastic") of basinal facies throughout northern British Columbia and Yukon. The *Amphipora* — bearing Middle Devonian black fetid carbonates of the McDame Group (*see* Gabrielse, 1963) may be in part correlative with a silicified amphiporal carbonate found 2.8 km southeast of Black Stuart Mountain.

The age of the sandstone unit is not directly known, but is probably Upper Devonian or younger because it overlies one of the conglomerate facies of the Guyet Formation. Its correlation outside the area is unknown.

Waverly Formation

The name Waverly was first used geologically by Johnston and Uglow (1926) and was subsequently abandoned by Sutherland Brown (1957). The name is reproposed for a formation of volcanic rocks found underlying the conglomerate of the Guyet Formation and which are well developed on the southwest slopes of Waverly Mountain. Sutherland Brown (1957,1963) included these volcanic rocks in the Guyet Formation but did not give them any status.

The Waverly Formation consists of schistose calcareous basaltic agglomerate and flows, pyroclastics, pillow basalt and minor breccia, and chlorite-rich quartz siltite. It is the only volcanic unit recognized from the Cariboo Terrane of the map area. The calcareous agglomerate, flows and pyroclastics distinguish the Waverly Formation from volcanic rocks of the Antler Formation of the Slide Mountain Terrane.

The Waverly Formation underlies a narrow belt extending from Summit Creek southeastward to near Black Stuart Mountain. The reference area of the Waverly Formation is in the area of Waverly Mountain where the unit is the most diverse and has a large areal extent. It was mapped by Johnston and Uglow (1926) as part of the Waverly Formation, which included rocks now mapped as Antler Formation. Sutherland Brown (1957) mapped it as part of the Guyet Formation, and later in the Black Stuart Mountain area as part of the Midas Formation (Sutherland Brown, 1963). Campbell et al. (1973) included the Waverly Formation as part of the Black Stuart or Guyet Formation. The thickness of the Waverly Formation, where present, varies from approximately 2 to 40 m. The lower contact is sharp and suspected to be conformable above the black pelite unit of the Black Stuart Group.

The basalt of the Waverly Formation is characteristically pyroclastic and carbonate-rich. In most areas the basalt is a schistose calcareous greenstone with little indication of original texture or mineralogy. It consists of a foliated matrix of chlorite, semi-opaques, sericite and epidote supporting 0.5 to 4 mm augen-shaped knots of dark grey to black calcite or dolomite (Table 8). Agglomerate with basalt clasts (pebble to boulder size) occurs

Table 8. Mineralogy of some rocks from the Waverly Formation

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26	27	28
Plagioclase			17+					8$	5*		25*	15#	2+									1&						15+
Augite							8&							1:+				1: +					2&					
Hornblende									8*				3&															
Quartz			6+	2&	8&			6@		70@			5&			2+	15@											
Pyrite		3&					1&						4#								3$		2$					
Opaques	4&		4+		3@	1$	3$	10&	10@	6@	12+	15+				8@	5$	4$		5@	5$	1&		2$	1@	5@		
Matrix																												
Plagioclase				30$		40&	30$		40&			30&	30@	25+	35&			50@	20@				45@				60$	
Quartz	5											2					5		8									
Chalcedony					25						10				5	10		6						5				
Chlorite	35	28	22	12	25	20	10	17	18	21	10	28	32	48	20	40	40	20	30	40	42		30	10	50	53	13	30
Sericite	10											10		8			5	5					6					10
Epidote	5	5		4	3		3							8			3				5				2			8
Calcite			6	14	9	14			7			10	10	1						20	10			40		2	5	12
Semi-opaque		7	20	30	15	5	10	6					8		10				20		25	18		5	15		6	
Secondary																												
Biotite					12&	8&									8&				12&									
Calcite								53$		3@						25&	10&											
Dolomite			25&								35&		6&			15&	10&						10$		30&	10&		
Amygdules	$-*	&-*	#-*			+-*	&-+		&-+						&-*				@-+	&-*		@-*	@-&	@-&		$-+	@-+	&-#
Calcite	26	37	6			8	5		7						13				2	18		30	3	8		8	3	8
Chlorite		15	2				10		3						7				6	7		20	2	5		22	7	8
Quartz	5	5					5		2										2	10		10		5				9
Plagioclase							4		5																			
Sericite	10																					20						
Biotite																											2	

Key to grain sizes in millimetres: 4 - 2 = *; 2 - 1 = #; 1 - .5 = +; .5 - .25 = &; .25 - .125 = $; .125 - .0625 = @;

1 Schistose calcareous basalt; hill NW of Antler Ck
2 Amygdaloidal basalt; Antler Ck
3 Tuff; Waverly Mtn
4 Basalt flow; Waverly Mtn
5 Volcanic sediment; Waverly Mtn
6 Agglomerate; Waverly Mtn
7 Pillow basalt; Waverly Mtn
8 Calcareous basalt flow; hill NW of Waverly Mtn
9 Basalt flow; hill NW of Waverly Mtn
10 Chloritic siltstone; hill NW of Waverly Mtn
11 Calcareous diabase; hill NW of Waverly Mtn
12 Diabase; hill NW of Waverly Mtn
13 Basalt flow; hill NW of Waverly Mtn
14 Volcanic sediment; hill NW of Waverly Mtn
15 Basalt flow; ridge S of Waverly Mtn
16 Calcareous basalt; ridge S of Waverly Mtn
17 Volcanic sediment; ridge S of Waverly Mtn
18 Agglomerate; ridge S of Waverly Mtn
19 Basalt flow; Cunningham North Mtn
20 Amygdaloidal basalt; ridge due N of Mt Guyet
21 Calcareous basalt; Alex Allan Ck
22 Amygdaloidal basalt; Cunningham North Mtn
23 Basalt flow; Cunningham North Mtn
24 Volcanic sediment; road along Loskey Ck
25 Greenstone; road along Loskey Ck
26 Amygdaloidal basalt; road along Loskey Ck
27 Basalt flow; road along Loskey Ck
28 Basalt flow; road crossing Tinsdale Ck

Explanation. The mineral percentages were determined by visual estimation with the aid of a petrographic microscope and are listed in the column under the sample number. The symbols represent an average size in millimetres for the particular mineral. Size estimates are not given for the fine constituents of the groundmass. Size ranges are given for the amygdules.

The classification "Secondary" is for coarser grained, distinctively secondary minerals, not necessarily confined to the fine grained minerals in its classification. The category "Amygdule" includes bona fide amygdules but also polycrystalline irregular interstitial filling where these are the proportions of amygdules.

General description. The plagioclase of these rocks has been nearly completely altered. Groundmass plagioclase is variolitic to trachytic microlites to laths and is the most completely altered. Plagioclase phenocrysts and microphenocrysts and euhedral to subhedral and never sufficiently concentrated to be in contact. Only sample 12 has zoned plagioclase. There are no ophitic textures with the sparse clinopyroxene. The augite is partly altered in all samples. Hornblende is not seen but chlorite pseudomorphs appear in several samples. The groundmass of the basalt flows consists of partly altered plagioclase laths and secondary derivatives of probably pre-existing glass. The term basalt is used with uncertainty because there are no analyses of these rocks and the mineralogy does not differentiate between andesite or basalt.

The secondary stilpnomelane is light straw brown to greenish straw brown and forms fine rosettes throughout groundmass and amygdules. Concentration of the stilpnomelane along fractures and amygdule-groundmass contacts occurs in one specimen (19). Secondary calcite and dolomite are disseminated throughout amygdules and groundmass.

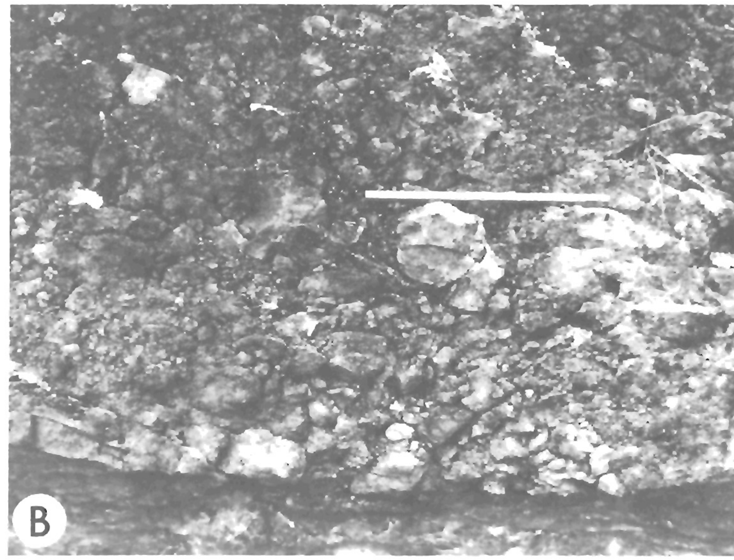
The mineral sequence within the amygdules, from wall to core, is: quartz, chlorite, calcite; where the quartz can be either coarsely crystalline or cryptocrystalline. Sericite, plagioclase and biotite which are also found in the amygdules, but more rarely, crystallized either prior to or synchronous with quartz.

All of these rocks have developed a deformational foliation ranging from schistose to weakly foliated. The more schistose varieties contain abundant carbonate, but not enough to 'float' the volcanic component. Where volcanic fragments float in a calcite matrix such as in sample 14, the original flow foliation of the feldspar microlites is maintained. Those volcanics with very little carbonate show minor pressure solution effects typical of competent rocks within the field area.

in the belt of Waverly from the hill south of Greenberry Mountain southeast to Cunningham North Mountain (Fig. 18). The matrix of the agglomerate is mostly calcite, locally forming 65% of the rock but in most places ranging from 5 to 40%. Basalt containing breccia fragments of laminated grey to dark grey moderately finely crystalline limestone is found on the logging road along Loskey Creek. The layer is 8 to 10 m thick and the fragments are 2 to 7 cm long parallel to their lamination. Flow and pillow basalt are found only near Waverly Mountain where the pillows range from 30 to 100 cm across and have little interstitial material. In the same area are pyroclastic basalts in 0.3 to 1.5 m beds of primarily varying shades of green and minor purple and cross-bedded choritic quartz siltstone; rock types not seen elsewhere.

The Waverly Formation basalt is bounded by Black Stuart Group black pelite at Cunningham Creek and southeast of Black Stuart Mountain. The carbonate-rich

Figure 18. Calcareous basalt agglomerate of the Waverly Formation.
A) Basalt blocks in calcite matrix near Greenberry Mountain. (GSC 191014)
B) Basalt fragments in calcareous matrix on North Cunningham Mountain. (GSC 191015)



agglomerate may represent direct volcanic activity transporting material as flows (Sutherland Brown, 1957, p. 36) and/or airfall debris or sedimentary transport of volcanic material into a carbonate-rich basin. The carbonate matrix is thought to be sedimentary and existed prior to the accumulation of the basalt because of the breccia fragments of laminated calcite found in the basalt of the Loskey Creek exposure. The occurrence of calcite throughout much of the Waverly Formation basalt may indicate that it was derived from the volcanic material. Ultimately both sources for the carbonate may be from the basaltic magma. The sedimentary carbonate could have formed by precipitation from circulating water fed by a carbonate-rich magma.

Age and correlation. The Waverly Formation directly overlies micritic limestone of the Black Stuart Group and therefore is equal to or younger than the age of the limestone; Middle to lower Late Devonian. It corresponds in lithology and age to Devonian basalt of the Earn Group from the Selwyn Basin (Gordey et al., 1982).

Guyet Formation

The Guyet Formation was introduced by Johnston and Uglow (1926) as consisting of conglomerate, sandstone and slate. Sutherland Brown (1957) redefined the formation, adding the Greenberry Limestone Member and basalt. Johnston and Uglow designated the limestone as an overlying formation and included the basalts in the Waverly Formation. Sutherland Brown (1957) placed the Guyet Formation with the Antler Formation in the Slide Mountain Group. It is proposed that the Guyet Formation be further redefined. Firstly, it should be removed from the Slide Mountain Group because it has no affinities with the Antler Formation. The Guyet Formation is interdigitated with the Black Stuart Group and is in tectonic contact with the Antler Formation. Secondly, the Greenberry limestone is reinstated as a formation. Lastly, the basalt of the Guyet Formation is included with agglomerates and volcaniclastics and those rocks are named the Waverly Formation after Waverly Mountain where they are well developed.

The Guyet Formation consists of quartzose sandy conglomerate to breccia, muddy conglomerate to breccia, sandstone and greywacke (Table 9 and 10). The conglomerate with its chert, pelite and local volcanic clasts is distinct from other conglomerate of the Cariboo Terrane. Its clasts and their inclusion in quartz sand or black pelite matrix are distinguishing features. A similar conglomerate occurs in the Hardscrabble Mountain succession of the Barkerville Terrane. Sandstone and blue quartz clasts are found in the Hardscrabble Mountain, but not Guyet Formation conglomerate.

The Guyet Formation is discontinuous throughout the map area being exposed mainly in two localities, Summit Creek-Alex Allan Creek area and from Mount Guyet to the south slopes of Mount Tinsdale. It can be traced from the Big Valley Creek area to south of Cariboo River where it feathers out into black slates of the Black Stuart Group. The conglomerate is found on the islands at the north end of Swan Lake.

Table 9. Mineralogical composition of Guyet Formation conglomerate.

Rock no.	QTZ	ORT	PLG	ZRC	BOP	CHR	CHP	PLT	SLT	VLC	RXQ	STY	MTX	OTH	
1	2				1	37	4	17	9	23			7		
2	9	1		1	5	17	7	2	2	37			22	+	C,MC
4	12					34	16	1	20	+			16		
5	46	1	+		+	14		8	20				13		
6	55	+			1	36	1	2	1		2	2	1		
7	4				2	69	11	6	6				4		
8	16												84	1	M
9	8				+	71	6	5	3	3			4		
10	24	2		+		25		13	10				27		
11	57				1	15	3	6			12	5			
10-1	3				+	42	13	18	11			3		1	SS
26-3D	67	+			+	23	4		3				1		
46-16	66				2	16	5						12	1	T
46-28	35				3	4								7	CRB
113-1	52	+		+	1	33	2	+	+		5	2	7	+	T
113-1A	59	2		+	4	15	3	1			6	3	7		
113-2	42		+		2	35	3	3	+		3	+	10	1	QM,M,T
126-25	43	1	+	+	+	19	5	7	6		4	8	7	1	QM,B
126-26	48	3	+		1	2	1	3	1		+	17	23	1	QM,B

Key

QTZ	= Quartz	PLT	= Pelite	C	= Chlorite
ORT	= Orthoclase	SLT	= Siltstone	CRB	= Carbonate
PLG	= Plagioclase	VLC	= Volcanic	M	= Muscovite
ZRC	= Zircon	RXQ	= Recrystalline quartz	MC	= Microcline
BOP	= Black opaques	STY	= Stylolites	QM	= Quartz mylonite
CHR	= Chert	MTX	= Matrix	B	= Biotite
CHP	= Cherty argillite	OTH	= Others		

\+ = less than 0.5 per cent

All mineral percentages are rounded to the nearest integer

Note: Samples 1 to 11 have been analyzed for strain and are discussed further in Appendix B.

Reference areas for the Guyet Formation are at Mount Guyet and Summit and Alex Allen creeks. Thickness of the formation is variable because it is discontinuous. The formation is thickest where the conglomerate is predominant, as at Alex Allan Creek, Mount Guyet and Mount Tinsdale. In general the formation is from several metres to 400 m thick, commonly being between 100 to 300 m thick. It pinches out near Black Stuart Mountain.

The lower contact of the Guyet Formation is conformable onto parts of the black pelite unit of the Black Stuart Group and may be unconformable on the Waverly Formation north of Mount Guyet. Basalt clasts of the Guyet Formation conglomerate resemble basalt of the Waverly Formation and may have been derived from it.

The sandy conglomerate (Fig. 19) underlies the area of Alex Allan Creek, Mount Guyet, and the island at the north end of Spectacle Lakes. Granule- to cobble-size clasts of green, tan, and grey chert to cherty argillite, grey to black chert, black pelite, light grey micaceous siltite and quartzite, and minor limestone and basalt are supported in a sand-size matrix of quartz, chert, siltite and minor feldspar. The matrix clasts are subrounded to rounded. Granule to pebble chert fragments are angular to subrounded and siltite and quartzite are subrounded to rounded. Pebbles and cobbles of limestone are angular. Pelite and cherty argillite are in most places too deformed to estimate original shapes.

The bedding within the sandy conglomerate is defined by layers of sand in which are supported coarse clasts and layers of sand with no coarse clasts. The sand matrix is continuous across the boundaries. Scoured contacts between overlying conglomerate and underlying sandstone were seen locally but their frequency is unknown due to the type of exposure. The sandstone is planar bedded where bedding could be detected from trains of opaques. Channels of Guyet conglomerate are cut into the underlying black siltstone and pelite of the Black Stuart Group at the junction of Tinsdale and Cunningham creeks. The erosional relief is up to 75 cm and the width of the channels is on the order of 1 to 2 m.

The sandy conglomerates of Alex Allan Creek and Mount Guyet are very similar in clast content, size, shape, percentage and distribution. The Spectacle Lakes occur-

Table 10. Mineralogical textures of Guyet Formation conglomerate.

Rock no.	QTZ	FFLD	CHR	CHP	PLT	SLT	VLC	OTH	
1	4CZ		0	0	0	0	0		
2	3CY	2DY	1DZ	1	1	2CY	0CY		
5	3CY		1CZ						
6	3CZ		2CZ	2					
7	4CY	1DZ	1	1	2				
8	6CZ							5	(M)
9	2CZ		1CZ		0				
10	3DZ		2DZ		2				
11	3		2	2	1				
10-1	2DZ		1	1	0	0			
26-3D	2CZ	0DZ	2		0DZ				
46-16	3DZ		2CZ	2					
46-28	4CZ		3DZ					3CZ	(CRB)
113-1	3CY		1CZ	2					
113-1A	4CY	3CZ	4CY	3	3				
113-2	3CZ		2CY	3	2			2CZ	(QM)
126-25	3CY	3CY	2CZ	3	2	2DZ			
126-26	4DY	4CY	3CZ	3	4	3CZ		3CY	(QM)

Key

Mineral and rock abbreviations are the same as in Table 9.

MGD = modal grain diameter
MGR = modal grain roundness
MGS = modal grain sphericity

Missing values of MGR and MGS are due to deformational changes of the grain shape parameters.

Grain diameter (MM):	Grain roundness:	Grain sphericity:
0 = 4 – 2	A = wellrounded	X = high
1 = 2 – 1	B = rounded	Y = moderate
2 = 1 – 0.5	C = subrounded	Z = low
3 = 0.5 – 0.25	D = subangular	
4 = 0.25 – 0.125	E = angular	
5 = 0.125 – 0.063	F = very angular	
6 = <0.063		

The results represent the textural characteristics of the first ten grains of each grain type encountered during point counting. Data are presented only if at least five grains of the grain type concerned had been counted upon completion of the point count.

Note: Samples 1 to 11 have been analyzed for strain and are discussed further in Appendix B.

rence has similar conglomerate but also has variations with higher chert content and a different chert type. The chert is pure and more varied in colour than the grey-green argillaceous cherts and cherty argillites common in the typical variety of conglomerate described from Mount Guyet-Alex Allan Creek. Chert clasts identical to those within the conglomerate are found within the directly overlying Visean limestone of the Greenberry Formation. On the same island there is a dark grey variety of conglomerate which has about 25% clasts, only half of which are dark and light grey chert. The remainder of the clasts are light grey and tan siltstones and purple to red ochre siltstones rich in iron oxide. The clasts are granules to fine pebbles and are angular to subrounded.

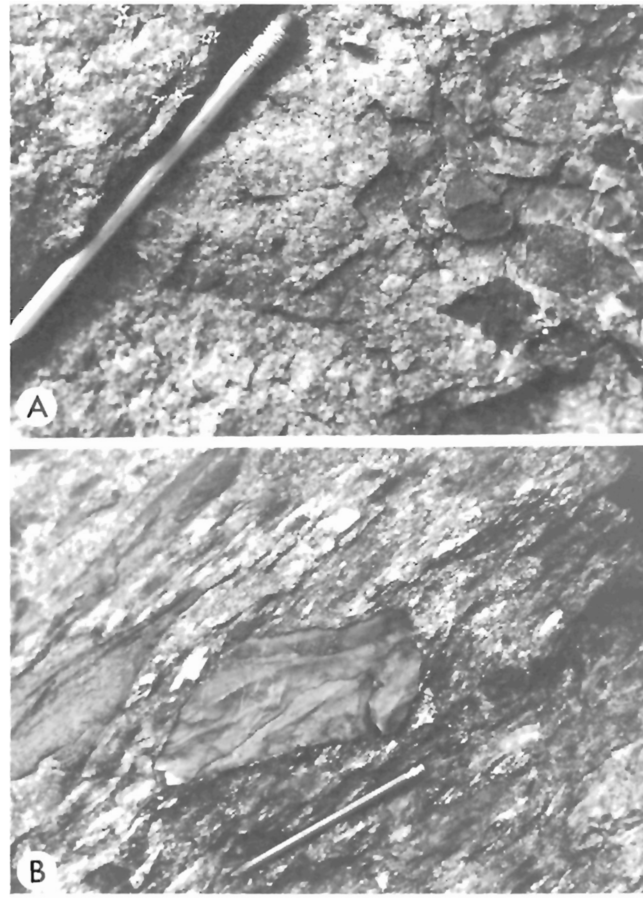


Figure 19. Sandy conglomerate of the Guyet Formation. A) Angular and subangular chert pebbles in quartz and chert sand matrix northeast of Alex Allan Creek. (GSC 191016)
B) Limestone cobble in a foliated matrix like that from a tributary of Alex Allan Creek. (GSC 191017)

The muddy conglomerate has the same clasts as the sandy conglomerate, however, there is a greater diversity in their size and percentage. In the Summit-Alex Allan Creek area there are more basalt, limestone and siltite clasts in the muddy conglomerate; they are also larger (Table 11). Sutherland Brown (1957, p. 33) described this type of conglomerate from Summit Creek where 1 m boulders of grey-green aphanitic basalt occur near the base of the Guyet Formation. Near the mouth of the first west-draining tributary to Alex Allan Creek similar volcanic clast-rich muddy conglomerate is interbedded with the sandy conglomerate. The section along this creek is described below.

The calcareous volcanic clasts contained in the conglomerate of this section are similar to the volcanics of the Waverly Member. The large volcanic boulders at the base of the muddy conglomerate of Summit Creek resemble the Waverly flow at the base of the conglomerate along Alex Allan Creek. Monotonous sandstone and sandy conglomerate 235 m thick occurs 250 m south along strike of the above section. Muddy conglomerate

Table 11. Reference section of Guyet Formation near the mouth of the first west-draining tributary to Alex Allan Creek (location; 53°08'28"N, 121°30'20"W).

Unit	Lithology	Thickness in metres Unit	Above base
12	Conglomerate, quartz, chert sand matrix, pebble clasts of grey to olive chert and cherty pelites, and minor light grey siltite, black pelite and sandstone, pebbles are angular to subrounded	30	69.5
11	Quartzite, grey, coarse grained	2	39.5
10	Conglomerate, black shale matrix, pebble and cobble clasts of grey green chert, quartzite and greenstone and are subrounded	5	37.5
9	Greenstone, flow?, poorly exposed and may be a large boulder	1	32.5
8	Conglomerate, green shale matrix, 40% pebble clasts of green cherty argillite, angular	3	29.5
7	Conglomerate, quartz sand matrix, 20% pebble clasts of light grey pelite	2.5	26.5
6	Conglomerate, matrix grading upward from quartz chert sand to black shale, 40% pebble and cobble clasts of chert, calcareous greenstone, dark grey pelite, light grey siltite and light green diabase, clasts are subrounded to rounded	4	24
5	Quartzite, grey, coarse grained, has an erosional upper contact		
4	Conglomerate, quartz, chert sand matrix, pebbles of grey pelite, grey and green cherts, light grey siltite and minor cobbles of calcareous amygdaloidal greenstone, grades into the overlying unit	1.5	19
3	Conglomerate, black shale matrix, 35% pebbles and cobbles of calcareous amygdaloidal greenstone, grey and green chert and cherty argillite, black pelite, light tan quartzite and green and brown siltstone, clasts are subrounded	4	17.5
	Covered	3	13.5
2	Quartzite, brownish grey moderately coarse grained	.5	10.5
1	Conglomerate, quartz, chert sand matrix, pebbles and cobbles of calcareous greenstone, dark grey and grey cherts, grey pelites, light grey moderately crystalline limestone and light brown siltstone	10	10
	Base unseen		

underlying the area south of Mount Tinsdale to Black Stuart Mountain does not have the volcanic cobbles and boulders and only rarely has granules or pebbles of greenstone.

Muddy conglomerate of the Guyet Formation is found interbedded with black slate of the Black Stuart Formation on Black Stuart Mountain and on the ridge east of there. It is the same as that found in the Tinsdale-Cunningham Creek area north of Cariboo River.

A dark grey greywacke and shale sequence with 10 to 75 cm beds overlies the muddy conglomerate on Mount Guyet. The contact with the underlying conglomerate is gradational over 3 m with interbedded conglomerate and greywacke. The greywacke is normally graded and becomes progressively finer upsection until it is interbedded with dark grey to black slate. The top of the sequence is not seen in this locality but is capped by Greenberry Formation limestone eastward along the ridge from Mount Guyet.

Sutherland Brown (1957, p. 34) described greywackes from the Guyet Formation conglomerate which have an average feldspar content of 12%. The greywacke he described is in part believed to include muddy conglomerate and finer equivalents but the high feldspar content could not be confirmed. Greywacke from the southern slopes of Mount Tinsdale does contain comparable amounts of feldspar to those reported by Sutherland Brown and may have been included in his report. These greywackes, however, overlie Greenberry Formation limestone and are therefore younger than the Guyet Formation conglomerate. On the west bank of Alex Allan Creek opposite the first west-flowing tributary there are outcrops of a feldspar, quartz, and quartzite granule and pebble conglomerate. These grey to dark grey rocks are well graded indicating inverted stratigraphy, if the grading is normal. From dark slate the rocks coarsen upsection. The mineralogy of these rocks is unlike that of the Guyet Formation conglomerate previously described, and is more akin to rocks ascribed to the Snowshoe Group.

Uglow (Johnston and Uglow, 1926) and Sutherland Brown (1957,1963) proposed that the clasts of the Guyet Formation conglomerate were derived from underlying Cariboo Group rocks. Campbell et al. (1973) disagreed. The clasts of the Guyet Formation conglomerate cannot be principally from the Cariboo or Snowshoe groups because those groups do not include primary chert, cherty argillite, tan siltstone and volcanic rocks. The chert and cherty argillite and some of the black pelite clasts of the Guyet Formation conglomerate contain recrystallized radiolarians (Fig. 20). These clasts must therefore be Paleozoic and cannot have been derived from non-radiolarian-bearing Hadrynian rocks. Sutherland Brown (1957) advocated chertification of Cariboo Group rock fragments to account for the chert within the Guyet Formation conglomerate, but this is unlikely considering the presence of radiolarians in some of the chert clasts. The major erosional unconformity is at the base of the Black Stuart Group and is of Early Ordovician age, therefore

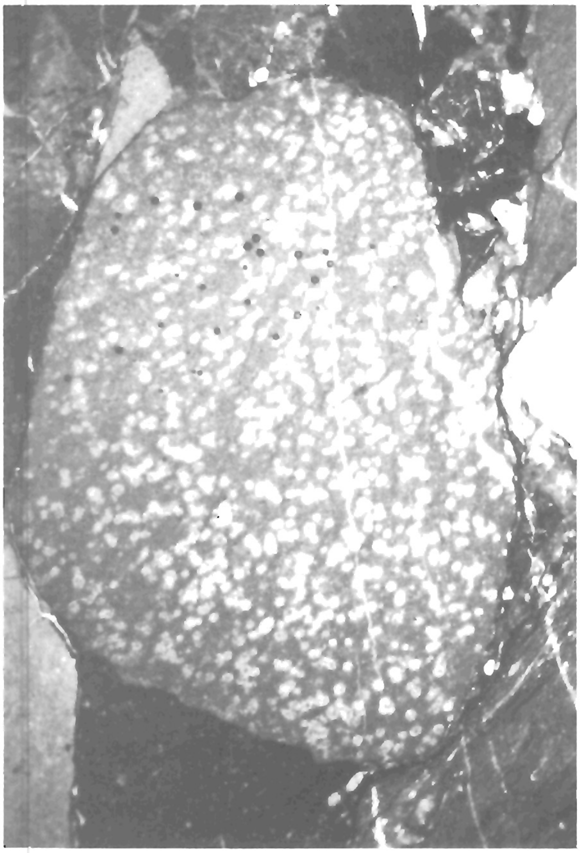


Figure 20. Radiolaria in a chert clast of the Guyet Formation conglomerate. Clast is 6 mm wide along the horizontal. (GSC 191018)

the Guyet Formation conglomerate is not associated with this erosional event. The Guyet Formation conglomerate has been deposited into Black Stuart Group black shales and the provenance of the clasts is mainly beyond the bounds of the present outcrops of the Guyet Formation.

The cherts and cherty sediment clasts of the Guyet Formation resemble Antler Formation sediments and were probably derived from a similar terrane. Because no such terrane is known to have existed east of the map area the source rocks were west of the depositional site. The volcanic clasts resemble the Waverly Formation and may have been derived locally or from Waverly equivalents to the west. Sources for the quartz, siltstone, quartzite, and meager feldspar are not known.

Age and correlation. The Guyet Formation cannot be dated directly. Its position beneath the Greenberry Formation, and most likely below upper Middle Devonian Black Stuart Group, brackets the conglomerate to the Late Devonian and/or Early Mississippian. Part of the conglomerate may be younger, as seen from the relationships on the island at the north end of Swan Lake. There, clasts of a chert in pebble conglomerate are also found within the directly overlying Tournaisian Greenberry Formation limestone. The Guyet Formation conglomerate is a time and lithological correlative of the Earn Group of Yukon (equivalents of which are informally called "black clastic") and its correlatives in northern British Columbia (Gabrielse et al., 1977; Taylor et al., 1979).

Paleoenvironmental history of the Black Stuart Group

The Black Stuart Group was deposited onto a gently undulating Ordovician unconformable surface (*see* section on Pre-Columbian palinspastic reconstruction for details). Where exposed the unconformity is nearly parallel to bedding, cutting through underlying stratigraphy at shallow angles.

The oldest unit of the Black Stuart Group is graptolitic shale. The unit is discontinuous and is known with certainty from only one area. The configuration of the unconformity may have had some influence on its distribution and facies, but because the shape of the unconformity is not known in detail, this possibility can not be pursued. Lithostratigraphic details from the graptolitic shale unit are: 1) the fine grained nature, 2) the even finely laminated bedding, 3) the absence of ripples, cross-laminations, graded bedding or dewatering and density inversion structures (flames, loads, convolution), 4) the absence of active burrowing organisms, 5) the moderate organic content (from dark grey colour) and 6) the presence of graptolite remains. Those features are sufficient to indicate that the shales were deposited in a quiet, stagnant basin receiving a limited input of clastics. Benedict and Walker (1978, p. 583), who summarized paleobathymetric indicators into table form, thought that flat thin laminae, abundant fine grains and graptolitic remains suggest water as shallow as 40 m but possibly much deeper. The source area and the direction of transport of these shales is unknown.

The chert-carbonate unit overlies the graptolitic shale of the black pelite unit and is restricted to the southern part of the map area. It changes character from the massive dolostone of Kimball Ridge southwestward to dolostone breccia. Sedimentary features of the dolostone breccia include: 1) clasts of chert, quartz, siltstone and dolostone near Limestone Creek, 2) pebbles and cobbles of dolostone in quartz sand found interbedded with dolostone breccia on Anderson Ridge, and 3) a limestone to limy dolomite matrix to the dolostone breccia fragments. The confinement of the dolostone to the unit suggests dolomitization at the time of deposition. The clastic rock of the southwestern facies indicates mechanical deposition (not recorded to the northeast) but could be either transported or in situ debris. The sand and chert fragments suggest that at least some of the material was from elsewhere. The dolostone of Kimball Ridge may have been partly broken up and transported to the southwest. The chert-carbonate unit is reasoned to have been deposited in shallow water because 1) penecontemporaneous dolomitization is a very shallow-water phenomenon and 2) conodonts from the unit are all shallow-water forms (*see* Seddon and Sweet, 1971). It may be in part redeposited in the southwestern facies with clasts either derived in situ or from nearby sources.

The black pelite unit overlying the chert-carbonate unit of the Black Stuart Group is Middle Devonian and may range from Lower to Upper Devonian. Its contact with the chert-carbonate unit is sharp and suspected to be conformable. Environmental stratigraphic detail includes: 1) fine grain, 2) even, finely laminated bedding, 3) occasionally graded silt lenses and laminae, 4) absence of ripples, crosslaminations, dewatering and density inversion structures, evidence of burrowing organisms, and fossils other than *Amphipora* and conodonts, 5) moderate to high carbon content, 6) low to moderate silicification, 7) variable quantities of thin black carbonate bands, and 8) presence of an oolitic grey limestone. Intercalated within the black unit are the Waverly and Guyet formations.

The black pelite of this unit is similar to the graptolitic shale, except it is more carbonaceous and siliceous. There are no physical paleobathymetric indicators which suggest deep water, though the basin would have been quiet and restricted with little clastic input. The conodonts from the upper Middle Devonian carbonates are all species of *Polygnathus*, which is suggestive of deep water environments (Chatteron, 1976) but the sample is far too limited to support this conclusively. Chatterton attributes a diverse fauna of *Polygnathus* with poor representation of other forms to a low energy basin or slope environment.

The intercalation of the Waverly Formation volcanics into this succession may indicate a shallowing of the basin because the Waverly is directly overlain by the grey oolitic limestone of the black pelite unit. Laminated grey limestone breccia fragments are found within the basalts of the Waverly. The laminated limestone is suspected to have been derived from a similar environment as the oolitic limestone. The oolites indicate wave-action; usually less than 10 to 15 m below sea level. Above the oolitic limestone are more black shales indicating a calming of current action which may have resulted from deepening of the water synchronous with a waning of the effects of Waverly volcanism. The Waverly Formation is overlain by both the black pelite unit of the Black Stuart Group and the Guyet Formation conglomerate.

Sedimentological detail of the Guyet Formation includes: 1) angular to rounded polymictic clasts, 2) grading displayed by decrease from pebble-clast size to the pervasive quartzite-chert-sand matrix, 3) erosion contacts at the base of coarse beds, 4) interbedding of muddy and sandy conglomerate, 5) changes in clast content, 6) even bedding, where seen, 7) massive sandstones, 8) suspension of solitary large clasts in finer massive matrix, and 9) general fining upsection in thick sequences. The environment of deposition is not obvious from this scanty information.

Several other observations are useful: 1) the conglomerate is intercalated in the black pelite unit of the Black Stuart Group, 2) the contact between these two units is only partly erosional, and 3) clast types possibly from the Black Stuart Group are scarce. These points imply the Guyet Formation conglomerate was deposited into a moderately deep-water environment. Sutherland Brown (1963) suggested (from the size of the clasts) that the Guyet was probably deposited at the mouth of a torrential stream. Such a stream must be erosional and the Guyet conglomerate does not record that erosion. Large clasts supported in a mud matrix (muddy conglomerate) is a textural anomaly for a conglomerate deposited in a fluviatile environment. The conglomerate is considered instead to have been brought to the depositional site as debris flows travelling into moderately deep water. The texture and grading characteristics of the conglomerate are features of proximal debris flows (*see* Krause and Oldershaw, 1979, and Walker, 1978).

Overlying the Guyet conglomerate are the black pelite and sandstone units in the Black Stuart Mountain area and the Greenberry limestone to the north. The upper black pelite unit is considered have been deposited penecontemporaneously with the northern Guyet Formation conglomerate and to be deposited in an environment similar to that of the older black pelite unit. The sandstone unit suggests a change of conditions with current induced sedimentary structures and a coarsening of grain size. The water was no longer stagnant but its depth need not have been different than during the time of deposition of the black pelite unit.

Greenberry Formation

Elevation of the Greenberry limestone member of the Guyet Formation to a formation, as originally suggested by Johnston and Uglow (1926), is proposed. Sutherland Brown (1957) made the Greenberry a member of the Guyet Formation because he believed it was thin, discontinuous and volumetrically unimportant, but the Greenberry is more continuous than he thought, and is distinct from, and everywhere overlies, the conglomerate of the Guyet Formation.

The Greenberry Formation consists of crinoidal limestone, and is locally altered to dolostone and chert. It is distinguished in the field from older limestone by its abundant crinoid fragments, but not so easily from the Sugar limestone of the Barkerville Terrane which is also crinoidal.

The Greenberry Formation is poorly exposed, underlying a narrow zone from Big Valley Creek in the northwest to Mount Tinsdale in the southeast. Any of its localities display the general characteristics of the unit and no type section is suggested. Its bedded character is best displayed north and east of Tinsdale Creek. The Greenberry Formation is reserved for the Mississippian crinoidal limestone of the Cariboo Terrane. The Greenberry Formation may be up to 30 m thick, but averages about 10 m thick.

The lower contact is rarely exposed but where seen is sharp against a thin shale bed on Guyet Formation conglomerate. It contains minor clasts of chert similar to those of the directly underlying conglomerate in the island at the north end of Spectacle Lakes implying a continuity of stratigraphy not seen to the west.

The crinoidal limestone is light grey weathering light grey to grey and contains debris of corals, brachiopods, echinoderms, bryozoans and gastropods. Microscopically it has minor amounts of finely crystalline quartz, heavy minerals, secondary pyrite, conodonts and fish teeth. The size and percentage of crinoid and other bioclastic debris is highly variable, with the rock ranging from crinoid wackestone to crinoid rudstone. The rudstone has the greater diversity of secondary fauna. Transitions from rudstone to wackestone are gradational in some localities (Fig. 21), but elsewhere are represented by abrupt irregular boundaries. All the biological material is fragmentary and shows no evidence of being in or near growth position. In most places the limestone is massive and devoid of sedimentary structures, however, in the outcrop south of Mount Tinsdale 15 to 40 cm beds are internally graded with respect to crinoid fragment size. The bedding and generally massive accumulation of broken fossils suggests that the unit was deposited as a debris flow rather than in situ.

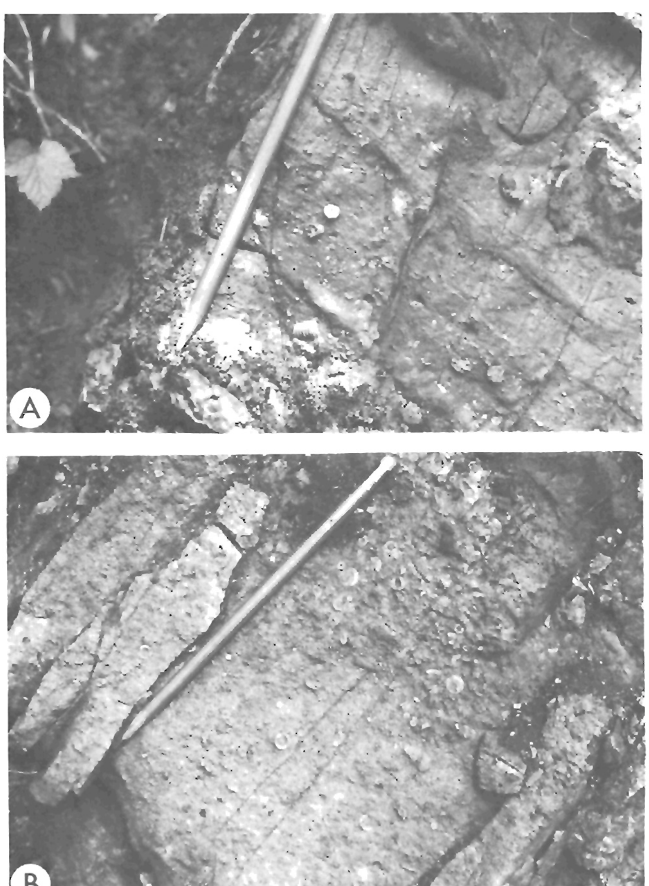


Figure 21. Greenberry Formation limestone near Mount Tinsdale. A) normally graded. (GSC 191019) B) inversely graded. (GSC 191020)

The Greenberry Formation has undergone various amounts of tectonism and neomorphism. Tectonic effects include pressure solution features, crushing-shearing phenomena and later profuse fracturing and calcite and quartz veining. Many outcrops display well developed shear foliation with rotated, granulated, and disrupted bioclasts. The light grey bioclastic limestone on lower Stewart Creek has two tectonic planes at 30° to each other. The first phase is characterized by surfaces of remobilization and pressure solution with carbonate-filled pressure shadows adjacent to crinoid fragments. The tails of the pressure shadows lie in the plane of flow. This surface of flow is cut by 1 to 2 mm spaced surfaces along which clasts have been granulated and disrupted.

Neomorphism was never sufficient to obliterate all bioclastic features, but coarsely crystalline carbonate was developed locally. Silicification, dolomitization and replacement by pyrite are local effects. Where the limestone has been partly silicified, such as, southwest of Mount Tinsdale, the silicification is in discrete layers 4 to 20 cm thick. The layers are continuous across outcrop widths of 10 to 30 m and appear to mimic original bedding. Chert and dolomite replace limestone both completely and preferentially. The chert is white or more commonly light grey and dark grey as in outcrops north of Nine Mile Lake and south of Mount Tinsdale. Dolomite is predominantly cream and is invariably associated with chert. In places pyrite selectively replaced crinoid stem fragments on a fine scale and is a common but minor secondary mineral.

Age and correlation. The Greenberry Formation is Lower Mississippian as determined from conodonts collected from many exposures. More precisely, the conodonts recovered are confined to the Upper Tournasian and Lower Visean (Orchard and Struik, 1985). The conodonts, identified by M.J. Orchard, are listed in Appendix A along with his comments. The age suggested by the conodonts agrees with earlier assessments from poorly preserved corals and brachiopods (Johnston and Uglow, 1926). Although the Greenberry Formation is the same age as many other crinoidal limestones in the Canadian Cordillera and elsewhere it does not appear to be a lateral continuation of any of them. Rather the Greenberry may be a local carbonate buildup on a hardground provided by the Guyet Formation. The conodont *Mestognathus* has not been found in the Rocky Mountains (North American Terrane) to the east and may represent a provinciality which distinguishes the Cariboo from the North American Terrane.

Alex Allan Formation

The Alex Allan Formation is here introduced to include Pennsylvanian black micritic limestone, and shale. It is similar to the interbedded shale and limestone of the underlying Black Stuart Group except for its thicker beds and more abundant limestone.

The formation underlies the area from Summit Creek northwest to the edge of the map area (Fig. 1). It is

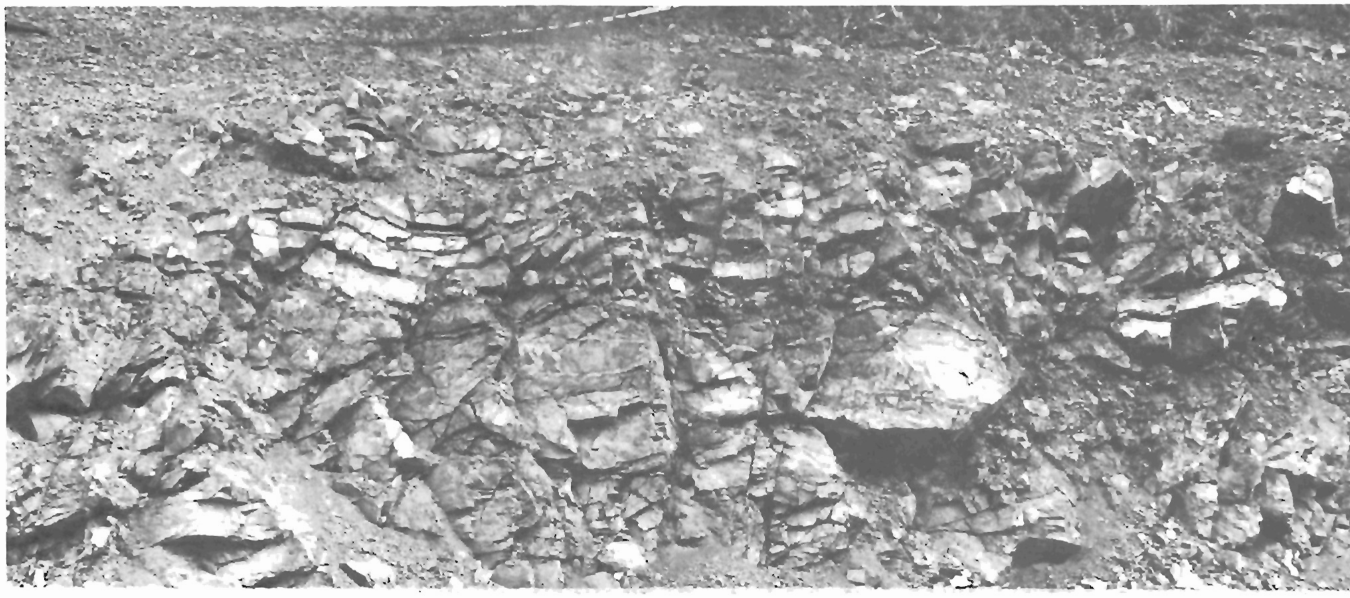


Figure 22. Bedding style of the Alex Allan Formation at its type locality on the road to Bowron Lake Provincial campsite, approximately 200 m north of where the road crosses Alex Allan Creek. (GSC 191021)

exposed in isolated outcrops locally with Greenberry and Guyet formations. A good type locality is a roadcut (Fig. 22) near the intersection of Allex Allan and Summit creeks; hence the name, Alex Allan Formation. The unit is given formation status, although it would not appear on most regional scales, because its conodont fauna distinguish it as unconformable on the Greenberry Formation. At its type locality it is 5 m thick, the maximum seen.

Contacts with underlying units are not exposed except at Summit Creek where the limestone is interbedded with black slate indistinguishable from slate apparently continuous with muddy conglomerate of the Guyet Formation. From relationships and fossil data from other areas there must be an unconformity concealed in the slate succession.

Grey to black micritic limestone, shaly and silty limestone and graphitic shale dominate the unit. Several beds at the Summit Creek locality contain wisps of shale (Fig. 23) and others have accumulations of fine grained bioclastic debris. Beds of limestone vary from 3 to 70 cm in thickness with thinner interbeds of shale and are commonly disrupted into augen-like lenses. Secondary pyrite appears throughout most of the unit.

Age and correlation. The age of the Alex Allan Formation is Middle Pennsylvanian as determined from conodonts examined by M.J. Orchard (*see* Appendix A, and Orchard and Struik, 1985). There is a mixture of age specific conodonts within the unit, representing Latest Mississippian to Middle Pennsylvanian. The mixing is thought to be due to either reworking or extremely slow deposition. Reworking is favoured (Orchard and Struik, 1985) because the same mixed fauna can be extracted from several beds within the unit (Fig. 22). No attempt is made to correlate beyond the field area.

Paleoenvironments of the Greenberry and Alex Allan formations

The Greenberry Formation contains abundant macrofossil debris and yields a conodont fauna indicative of a shallow-water environment. The conodont *Mestogna-*

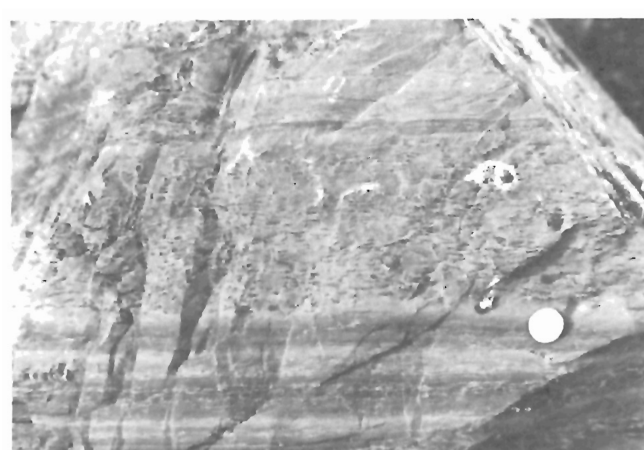


Figure 23. Wisps of pelite, which may be fossil ripples, in the black micrite of the Alex Allan Formation; Canadian quarter for scale. (GSC 191022)

thus is the most diagnositic for shallow-water, being deposited in oolitic or thick-bedded bioclastic limestone (Austin, 1976, p. 223). The macrofossil debris is fragmentary and shows no sign of having been deposited in place. Bedding is planar and the bioclasts are locally graded. The meagre information suggests redeposition of the bioclasts possibly as debris flows. However, the absence of any deeper water fauna mixed with the shallow may argue against far travelled redeposition.

The Alex Allan Formation consists of muddy limestone and shale which contain a mixed conodont fauna, all of which are compatible with deep water deposition. The absence of *Cavusgnathus*, which represents very shallow-water conditions (*see* von Bitter, 1976, and Merril and Martin, 1976), hints at deeper water deposition for the Alex Allan Formation. The lack of macrofossils and the presence of black fine grained clastics suggest a quiet water situation deeper than the wave base. The mixed fauna may result from very slow deposition or from reworking of older material, either case being compatible with a quiet moderately deep-water environment.

Unnamed Pennsylvanian limestone

Pennsylvanian crinoidal, fusulinid limestone found on the island at the north end of Spectacle Lakes is distinguished from the dark grey micritic limestone of the Alex Allan Formation by being more lightly coloured and bioclastic. It is similar to Greenberry Formation limestone but has a much lower percentage of crinoid fragments and contains fusulinids.

The unit is confined to the island on Spectacle Lakes and is exposed with Greenberry Formation limestone and Guyet Formation conglomerate. It was previously included in the Greenberry member of the Guyet Formation by Sutherland Brown (1963) and Campbell et al. (1973). Contacts with nearby units are not exposed.

Light grey weathering grey finely crystalline limestone supports approximately 6% poorly preserved fusulinids, 5% crinoid stem fragments (1-3 mm long) and minor forams and conodonts. The limestone is pervasively veined with white calcite (fracture filling).

Age and correlation. Conodonts from this unit are like the Middle Pennsylvanian ones of the Alex Allan Formation (M.J. Orchard, personal communication, 1984). The fusulinids may be Lower Permian (C. Ross, written communication, 1980) or Middle Pennsylvanian (W.R. Danner, written communication, 1980).

Unnamed Permo-Triassic clastic rocks

Olive-grey greywacke and slate overlies the Greenberry Formation on Two-Bit Creek in northern Wells map area and on the south slopes of Mount Tinsdale in Cariboo Lake map area. The rocks are similar to feldspathic micaceous quartzites in parts of the Snowshoe Group of Barkerville Terrane. The distribution of the unit is a function of structural extrapolation and interpretation. The lower contact with Greenberry Formation on Two-Bit Creek is sharp and apparently sedimentary.

The greywacke clasts consist mainly of subrounded quartz with small amounts of potassium feldspar, plagioclase, muscovite and chlorite. The matrix is of sericite, chlorite and calcite. The grain size ranges from silt to very coarse sand and beds locally display grading. The slates are mainly olive, platy and thinly bedded. Beds of coarse grained greywacke are thick and massive whereas in fine grained sequences they are thin.

Age and correlation. The sequence overlies the Greenberry limestone and does not occur between the Greenberry and Alex Allan formations. Parts of the limestone on the island at the north end of Spectacle Lakes are age equivalent to the Alex Allan Formation but some of it may be Lower Permian. It is presumed that the olive-grey clastics overlie these limestone units and therefore are younger and may be Permian and/or Triassic.

Intrusive rocks

Intrusive rocks of the Cariboo Terrane are divided into three categories on the basis of relationships with host rocks and their lithology. They are intrusions of diabase, quartz porphyry and lamprophyre. The Mount Murray intrusives were defined to include intrusives of both the Antler Formation and Cariboo Group (Sutherland Brown, 1957, 1963). Because there is doubt as to correlation of intrusive rock between these units, the term "Mount Murray intrusives" is not used here.

Diabasic intrusives

Dykes and sills invading the Black Stuart Group south of Cariboo River and a dyke cutting Cariboo Group in the Roundtop Mountain area are volumetrically the most important of a series of diabasic intrusives (Table 12). Less conspicuous and smaller intrusions of diabase occur in scattered localities.

Two of the dykes from the Black Stuart Mountain area contain myrmekite. Edges of oligoclase grains show various stages from unaltered to the development of worm-like quartz intergrown with oligoclase or albite. Though the quartz growths are usually around edges they locally invade cores. Complete replacement of the original oligoclase appears to result in the symplectic intergrowth of quartz and albite.

A black aphanitic basalt sill intrudes the black pelite unit of the Black Stuart Group on Cunningham Creek just downstream from its junction with Roundtop Creek. This sill is included with the diabases because it could be their finer equivalent. It is the only one of its kind found in the area. Crosscutting relationships with host rocks demonstrate a Devonian or younger age because the youngest rock intruded is the black pelite unit of the Black Stuart Group.

Table 12. Compositions of diabase from the Cariboo Terrane

	1	2	3	4
Plagioclase (altered)				
laths 0.3–1.5 mm long		40		
blocky crystals 0.3–2.5 mm				
An_{25} – An_{35}	50		65	50
Augite (partially chloritized)				
0.1–0.4 mm		5		
0.4–1 mm		13		
Chlorite (interstitial)	27	30		
Biotite (secondary)				
olive brown			10	15
rusty brown		2	8	10
Stilpnomelane (secondary)			5	
Quartz, 0.05–0.3 mm	5		5	
Sericite (secondary)		5		5
Fe-dolomite (secondary)	10			5
Opaque				
Fe-oxide	8		4	3
pyrite		5		
Myrmekite (secondary)			8	12

Key

1. Dyke on ridge south of Middle Mountain
2. Black Stuart Mountain area
3. Black Stuart Mountain area
4. Black Stuart Mountain area, olive-brown biotite is an alteration of hornblende

Quartz porphyry intrusives

A quartz porphyry to quartz-feldspar porphyry sill can be mapped intermittently from Jubilee Creek to the area west of Two Sisters Mountain. It ranges up to approximately 8 m in thickness. A similar intrusive is found in the Antler Formation near its base in the Mount Tinsdale area. The Proserpine dykes may be ankeritized equivalents of this quartz porphyry series.

The mineralogical content of the sill was estimated from thin section:

Quartz, phenocrysts, 1-5 mm	5-15%
Potassium feldspar, phenocrysts, 1-6 mm	0- 8%
Oligoclase, phenocrysts, 2-4 mm	0- 3%
Matrix of fine quartz, sericite and fine blades of oligoclase	80-85%

The sill intrudes Guyet Formation and strata overlying Middle Pennsylvanian Alex Allan Formation limestone and therefore must be Pennsylvanian or younger. It is folded and therefore was emplaced during, or prior to, the major folding event.

The felsic intrusives were intruded prior to some folding and after the emplacement of the Antler Formation in the early Mesozoic. The major folding event is Jurassic and/or lower Cretaceous. The age of the quartz porphyry intrusives is therefore bracketed between Permian and Late Cretaceous.

Lamprophyre intrusives

The only occurrence of lamprophyre found is the one reported by Holland (1954, p. 23) and Sutherland Brown (1957, p. 45). It intrudes the northeast-trending fault at Roundtop Mountain. The dyke is later than the fault and is therefore younger than the pervasive deformational event (Columbian). It is considered to be Tertiary, but may be Late Cretaceous.

Economic geology

The Cunningham Formation is host to some replacement deposits of galena and sphalerite. For example, on the Cunningham Forest Road at Roundtop Creek, limestone is altered to dolostone with minor galena.

The Black Stuart Group on Anderson Ridge hosts quartz-barite veins with galena and lesser amounts of sphalerite (Reed and Lovang, 1981). Although the mineralization is reported to be from the Mural Formation it is more likely to be hosted in the chert-carbonate unit of the Black Stuart Group.

Structural and metamorphic geology

The Cariboo Terrane consists of essentially one structural package; defined as a deformed sequence of rock separated from others by an angular unconformity. The package consists of all units from Kaza Group to Permo-Triassic sedimentary rocks.

The structures of the package form a spectrum but for purposes of discussion are divided into 3 categories based on conditions of formation and superposition. From oldest to youngest they are 1) flow, 2) ductile shortening, and 3) brittle shortening and extension. The strain environment is emphasized over the superposition of structural elements. As a consequence little importance is attached to designating structures as S_1 or S_2 etc.; although for comparison purposes this is done in its simplest form in Table 13. The metamorphism is included in the discussion of the structures because it reflects to a large degree the strain parameters. The structural spectrum began with flow folding possibly in latest Triassic and ended with brittle extension due to uplift in the Eocene.

Flow

This category of structures includes isoclinal, small scale rooted and rootless folds in limestones of the Cunningham, Yankee Belle and Mural formations. The limbs of these folds are parallel to a colour lamination and parting that parallels bedding as displayed by lithological transitions. They are found only locally in recrystallized limestones, and no larger scale isoclinal structures have been found associated with them. Too few measurements of

Table 13. The structures of the Cariboo Terrane organized into a series of deformations (D) of surfaces (S), and folds (F).

Event	Structure	Metamorphic mineral
D_4	Extension-cross faults striking NE and steeply dipping to S.	
	Extension faults striking NNE and N with steep dip and having oblique-slip and some strike-slip.	
D_3 (F_3)	Open folds, warps, kinks and crenulations (F_3). Normal and reverse strike faults steeply dipping.	Chlorite?
D_2 (S_2, F_2)	Pervasive axial surface cleavage (S_2); asymmetric (W-directed) flow to concentric folds (F_2) that trend and plunge to NW; thrust faults E-dipping	Quartz, albite, biotite, chlorite muscovite chloritoid,
D_1 (S_1)	Displayed in carbonate. Bedding-cleavage (S_1); isoclinal recumbent rootless fold (F_1) thrust?	Quartz, albite?, muscovite

fold axes were recorded from these features to conclude anything on their orientation with respect to younger folding. Those measurements taken indicate that hinges both parallel and diverge to varying degrees from the average orientation of the major folds. Layer parallel veinlets of calcite crosscut the isoclines and are themselves folded by later phases. Figure 8 shows this vein-isocline relationship and the development of a crosscutting, and therefore later, parting.

Ductile shortening

Cleavage, folds and faults of this category display characteristics of ductile flow.

Cleavage

The term cleavage as used here follows the definition from Bayly et al. (1977, p. 9):

> "Set of closely spaced secondary planar parallel fabric elements which impart mechanical anisotropy to the rock, without apparent loss of cohesion."

Included under this definition are such features as foliation and schistosity.

The cleavage is pervasive throughout Cariboo Terrane rocks of the map area. It strikes dominantly west-northwest and dips moderately to steeply to the north-northeast. Variations in orientation are displayed by the Lambert net plots of Figure 24 whose domain divisions are shown in Figure 25. The cleavage consists of parallel surfaces of mica and is axial planar to outcrop and regional scale folds.

The cleavage is defined by metamorphic muscovite, quartz, albite, chlorite and locally biotite, and its character is governed by rock type. The metamorphic micas are generally parallel suggesting the peak temperature of metamorphism coincided with the formation of the cleavage. Exceptions are on Anderson Ridge where biotite randomly overgrows muscovite foliation surfaces, on Middle Ridge where chloritoid overgrows muscovite foliation (Fig. 26) and near Waverly Mountain where felted stilpnomelane overgrows the chlorite foliation of the Waverly Formation basalt tuff. The temperature of metamorphism is mostly 300-400°C as defined by the colour alteration index of conodonts (from CAI chart of Epstein et al., 1977) extracted from the Black Stuart Group and Greenberry Formation. Locally it may have been higher as the conodonts are from stratigraphically high units.

The effect of rock type on cleavage will be discussed for impure quartzite and siltite, quartzite, conglomerate, limestone and pelite.

The impure quartzite and siltite of the Cariboo Group has cleavage defined by parting along white mica and minor chlorite foliation surfaces. The micas grow in a preferred orientation and form a meshwork around resistant quartz and feldspar grains. The result is not discretely spaced cleavage surfaces, but rather, a continuous mass of oriented mica interrupted by the resistant coarser grains. The parting is along the orientation direction and the surfaces have a topography governed by several resistant grains. A similar type of cleavage occurs in conglomerates of the Snowshoe Group, however, the clasts are invariably flattened and stretched in the plane of mica foliation. Thin pressure solution selvages associated with the large clasts induce a spaced parting not evident in the finer grained rocks. The relief of the parting surfaces is half of the short axis of the largest grain.

Microscopically, pressure-solution features form an integral part of the cleavage. Solution selvages composed of opaque matter are erratically spaced, parallel to the mica elongation, discontinuous and localized. They are volumetrically more important in poorly sorted, coarser grained clastics. Quartz grain shape is in part controlled by solution.

In Yankee Belle Formation impure quartzite of the Roundtop-Middle Ridge system preferentially weathered cleavage is spaced at 0.5 to 2.5 cm intervals, imparting a lenticular appearance to the impure quartzite (Fig. 27).

Quartzite of the Yanks Peak Formation displays foliation, dissolution surfaces, and subgrain trains, but cleavage on outcrop scale is not well developed. Low sericite contents define the foliation. Dissolution surfaces are best developed when cleavage is parallel to bedding and, like the subgrain trains, are not usually detectable with the unaided eye.

Sandy conglomerate of the Guyet Formation, like the Yanks Peak Formation quartzite, has a poor cleavage expression in outcrops. As for the Yanks Peak Formation, this poor expression is due to lack of abundant matrix micas, and because the conglomerate is essentially an L-tectonite; that is, a rock which has been deformed such that the strain ellipse is more cigar-like than pancake-like. Strain measured in eleven samples of Guyet Formation conglomerate is mainly constrictional and records some 40% extension parallel to the regional fold axes. This extension formed during the period of ductile deformation and is unrelated to later brittle extension. Detailed descriptions of the strain measuring technique and the deformation characteristics of the conglomerate are given in Appendix 3. Cleavage is poorly developed in the sandy conglomerate, not because it is much less deformed than the Cariboo Group, but because of its lithology and style of deformation. The polymictic nature of the conglomerate has caused differential cleavage development between clasts.

Within the muddy conglomerate of the Guyet Formation, cleavage relationships between clast and matrix are more obvious than in the sandy conglomerate. The shale matrix supports a well developed cleavage that penetrates through both matrix and clasts of similar lithology. This contradicts the implication of Sutherland Brown (1957) that schistose clasts occur within an unfoliated matrix and agrees with the observation of Campbell et al. (1973, p. 58 and 79). The muddy conglomerate has a cleavage best described as intermediate between rough and smooth spaced, from the classification of Powell (1979, Fig. 4, p. 27). Where there are pre-existing layers at angles to the cleavage it has the features of zonal and discrete crenulation cleavage as described by Gray (1977). The spacing of the cleavage is governed by the clast size.

Limestone displays various styles of cleavage, generally poorly expressed. Best developed is bedding-parallel parting associated with dissolution surfaces. The parting surfaces have a 0.05 to 4 mm spacing and are quite smooth. Sericite flakes may occur along them when the limestone is moderately fine or more coarsely crystalline. Where there is a quartz-silt content, recrystallization localizes along the silt lenses and very thin planar dissolution selvages border the zones of different crystal size.

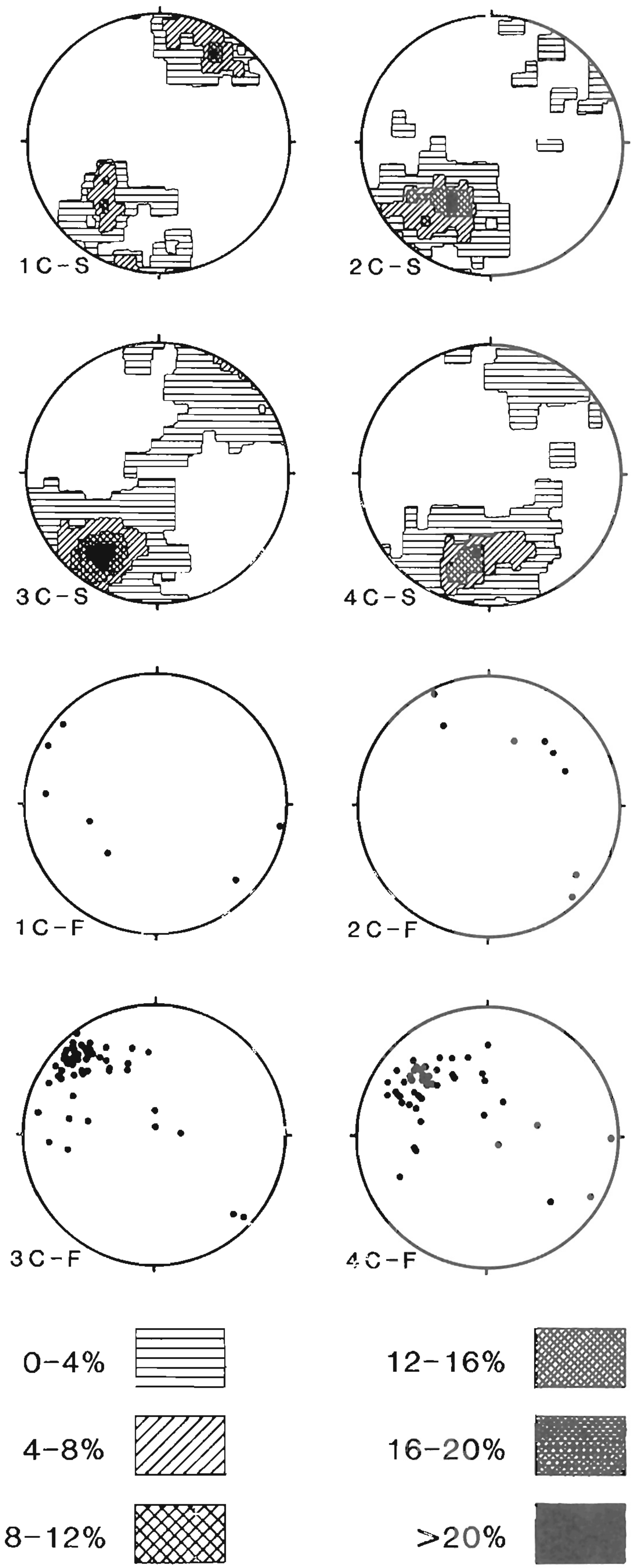


Figure 24. Equal area net plots of the orientation of ductile style fold axes and the poles to cleavage for four structural domains of the Cariboo Terrane. Of the three characters in the labels, the first refers to the number of the domain (see Fig. 25), the second is a C for Cariboo Terrane and the third is either an S for cleavage surface or an F for fold axis.

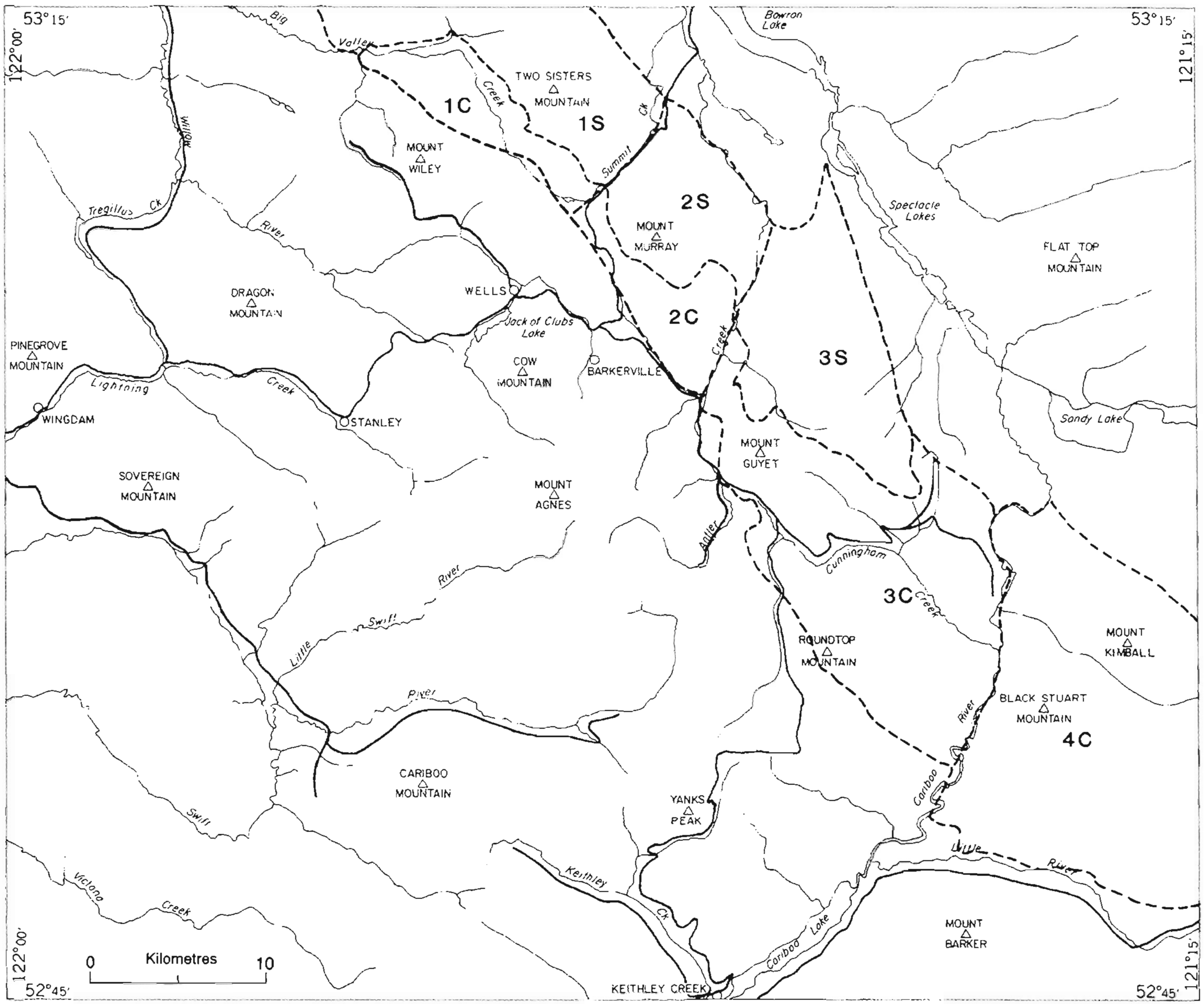


Figure 25. Map of the arbitrary structural domain divisions of the Cariboo Terrane (C) and the Slide Mountain Terrane (S).

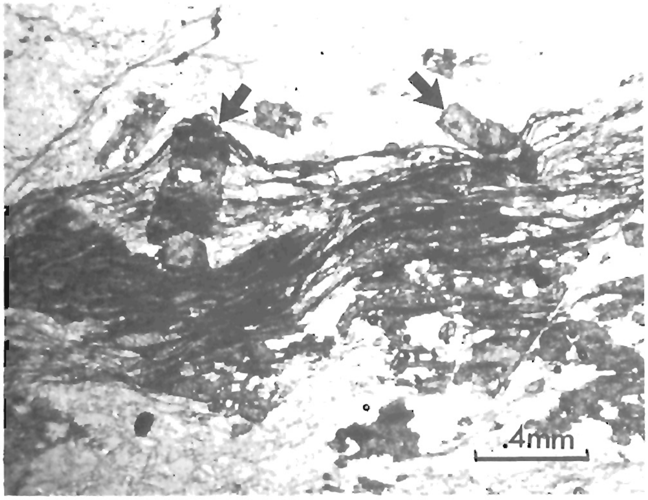


Figure 26. Chloritoid crystals (indicated by arrows) grown at angles to the stylolite and white mica foliation in fine grained quartzite of the Yanks Peak Formation on Roundtop Mountain.

Figure 27. Spaced cleavage in micaceous quartzite of the Yankee Belle Formation near Middle Mountain. (GSC 191024)

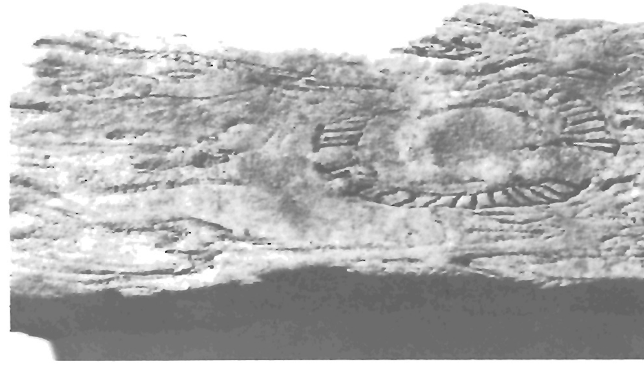


Figure 28. Stretched archaeocyathid cross-section in Mural Formation limestone at the headwaters of Sunshine Creek.

Figure 29. Open and tight folds in Yankee Belle Formation micaceous quartzite near Middle Mountain. The tight fold is adjacent to a reverse fault, fold side up. The white rectangular patch at the bottom of the photograph is a notebook for scale. (GSC 191023)

Stylolites rarely attain amplitudes greater than 3 mm and are most often perpendicular to bedding. Locally there is a spaced parting at angles to bedding. An example of the mobility of the Mural Formation carbonate is shown in Figure 28 where archaeocyathids are flattened along a finely spaced cleavage.

The cleavage is pervasive throughout all pelite of the section. It is predominantly of the smooth closely spaced type of Powell (1979). Rarely is the secondary mica sufficiently pervasive to produce a continuous cleavage. In relation to rock classification the pelite cleavage varies from slaty to phyllitic. Nowhere within the mapped area are the rocks actually schists.

Folds

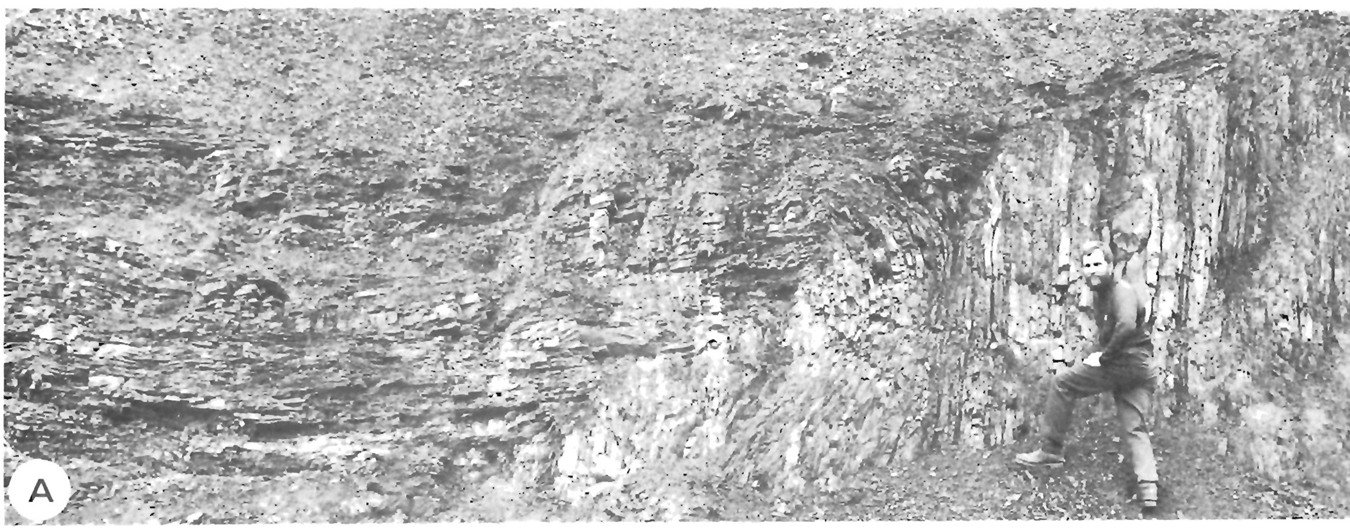
Fold axes generally plunge between 10 and 30° to the northwest. Folds are relatively open on a large scale forming broad asymmetrical anticlines and synclines such as those in the Black Stuart and Kimball Mountain areas. Outcrop scale folds are open to isoclinal, generally having interlimb angles of between 35 and 70°. Similar to the macroscopic folds, they are asymmetric and westward-verging. The steep limbs may be overturned to the west and are frequently offset by east-dipping reverse and thrust faults. Their geometric classification ranges from concentric to similar, commonly being of class 1c (see Ramsay, 1967, p. 365-366). Pictures of some of the fold types can be found in the publications of Sutherland Brown (1957, Plates VIB and VIIB) and Holland (1954, Plates V, VI, VIIB and VIIIB) and in Figure 29. The fold style varies directly with lithology and proximity to faults. Lithologically, the different response of quartzite versus pelite is the most obvious.

Quartzite forms open to tight folds on both outcrop and regional scale. Commonly the folds are not continuous because of forelimb fault disruption. Backlimbs and hinge areas approximately maintain bedding thickness. The forelimbs, where faulted, show dramatic thinning, accompanied by recrystallization of the quartzite. The recrystallization can leave the quartzite with the appearance of a quartz vein. An example of which is the "quartz comb" on Middle Ridge as mapped by Holland (1954).

Pelite is a common rock constituent of the area and shows the most diversity of fold styles, ranging from asymmetric open and tight to asymmetric-to-recumbent tight and isoclinal. The black pelite unit of the Black Stuart Group is locally isoclinally folded on the outcrop- and hand-specimen-scale with the best examples of this in the Tinsdale-Cunningham Creek and Black Stuart Mountain areas (Fig. 30). Pelite of the Isaac Formation folds in the same manner as the black pelite unit of the Black Stuart Group. Impure quartzite, limestone and silty pelite have fold styles intermediate between those of pure quartzite and pelite.

Near faults the intensity of folding increases. Tight and isoclinal folds are localized in the hanging and footwalls of thrust faults. Fold amplitudes increase and wavelengths decrease adjacent to high angle reverse faults of the Roundtop-Middle Mountain area (Fig. 27).

Conical folds of Cariboo Group of the Roundtop Mountain area were first described by Sutherland Brown (1957, p. 48-49). It is suspected that they are common in the area but are obscured by lack of outcrop. Because of this obscurity, folds are projected into structural cross-sections as though they were nearly cylindrical. The cross-



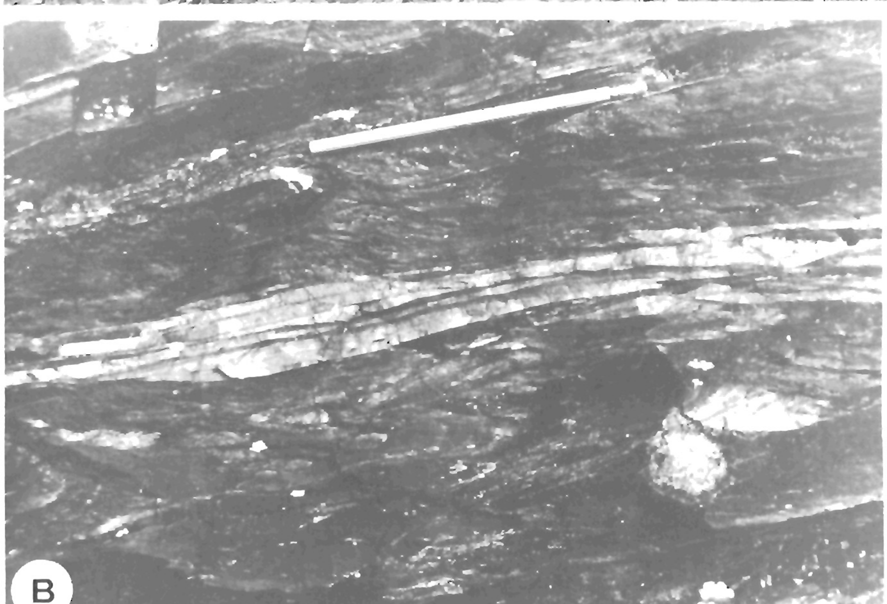


Figure 30. Isoclinal folds in the black pelite unit of the Black Stuart Group.
A) Refolded isocline on Tinsdale Creek. Fold axes of isoclines parallel the outcrop. (GSC 191025)
B) Isoclines near Black Stuart Mountain. (GSC 191026)

sections, therefore are schematic in relation to amplitude, wavelength and symmetry of included folds.

Faults

Westward-directed thrust faults formed during the period of ductile shortening. They are folded and cut folds in the sympathetic manner displayed in the Rocky Mountain Fold and Thrust Belt (*see* Dahlstrom, 1970). The thrusts cut bedding at shallow angles and locally follow bedding planes for several kilometres. They cut downsection to the east as locally observed and can be inferred to do so from outcrop and known dips.

As shown in cross-section E-E' (*see* maps, in pocket), thrusts formed prior to folding maintain similar cutoff angles to depth. They follow 'flats' near the base of the Black Stuart Group and Yankee Belle, Yanks Peak and Cunningham formations. To satisfy displacement and geometry requirements for cross-section A-A', the thrusts must root as a single fault in the Isaac Formation under Anderson and Kimball ridges. East of the map area this thrust may root in the Matthew Fault. The thrust is not drawn in the thick Isaac Formation of northeastern Kimball Ridge in cross-section D-D', because its relative position with respect to the Isaac-Cunningham Formation contact is unknown. Folded thrusts in cross-section D-D', northeast of the major vertical fault are drawn to cut shallowly downsection to the east to be consistent with similar faults to the south.

Surface exposures of folded thrust faults allow minimum displacement estimates of 1 to 3 km. From the low angle between thrust and bedding and the stratigraphic levels which have been juxtaposed, true displacements must be somewhat larger. A fuller consideration to thrust displacement is given in the section on the palinspastic reconstruction of cross-section E-E'.

Moderately dipping thrust faults in the Roundtop-Middle Mountain area cut the steep limbs of west-verging asymmetrical folds outlined by the Yanks Peak Formation quartzite. The limbs of the folds displaced by the faults are thinned and in places form lenses along the fault surface.

The thrust faults are younger than part of the black pelite unit of the Black Stuart Group which they override. They are related to the regional folding and are therefore as young as the folds. Even though some features indicate an age sequence to these structures, they are considered penecontemporaneous.

Brittle shortening and extension

Cleavage, folds and faults of this category display characteristics of brittle deformation.

The cleavage is axial planar to crenulations and minor folds. It is only of local importance, occurring primarily in the Cariboo Group. Rarely, phyllite of the Black Stuart Group has a poorly developed crenulation cleavage. The third phase cleavage from the area south of Cariboo River strikes west-northwest and dips steeply northward. It coincides approximately with the modal attitude of the ductile cleavage but is more northwesterly in strike.

The crenulations and minor folds are restricted to impure quartzite and phyllite and are found primarily in Cariboo Group and locally in the Black Stuart Group and Guyet and Greenberry formations. They deform cleavage associated with ductile shortening. The attitude of their fold axes approximates that of the ductile folds in being northwest-southeast.

Kinkfolds are locally developed in pelite of the Cariboo Group and Black Stuart Formation. They have amplitudes from 0.5 to 3 cm and wavelengths of 3 to 15 cm. The hinge zones are disrupted, forming a spaced axial planar parting. There is insufficient data to relate their attitudes to other structures.

Faults

Brittle faults are steep dipping and both parallel and dissect the orogen. They are divided into three sets by common orientation and crosscutting relationships. The first set of brittle faults consists of steep reverse and normal faults which strike parallel to the orogen. They offset the ductile thrusts. Dip-slip displacements, as measured from the cross-sections, vary from less than 250 m to greater than 1 km.

The second set consists of northeast-trending transverse faults which offset all earlier faults. They are usually southeast side down with a component of strike slip. They dip steeply to vertical and are most conspicuous in the area between Cariboo River and North Cunningham Mountain. A fault of this category is plugged with a Tertiary lamprophyre dyke on Roundtop Mountain. The fault is older than the dyke and younger than the ductile folds.

The third set consists of north- and east-northeast-trending right-lateral strike-slip faults that crosscut all previous structures with the possible exception of the second set of transverse faults. These faults transect the map area along Antler Creek between Mounts Waverly and Howley and accumulate a displacement of some 4 km. They are considered to be Late Cretaceous, coinciding with right-lateral strike-slip movement in northern British Columbia and Yukon (*see* Tempelman-Kluit, 1979). Sutherland Brown (1957) described strike-slip faults which are reactivated north-trending transverse faults.

Joints

Jointing is pervasive throughout the Cariboo Terrane. The density and continuity are variable. Parallel joints of a single outcrop have no preferred spacing and can vary from 0.5 to several metres. They may be of only local importance or continue for up to 10 to 15 m or farther beyond the confines of the outcrop. Within a single outcrop there may be up to 4 or 5 joint orientations. Only 1 or 2 of these will be prominent. There is a difference in modal joint orientations between the Cariboo and Slide Mountain terranes.

As with cleavage orientations, joints have been plotted by stratigraphic group and domain (Fig. 31 and 32). The modes are well defined for the Cariboo Group in all domains. For the Black Stuart Group and Guyet and Greenberry formations the data have more scatter. The modal attitudes of the poles to joints and the fold axes are similar enough to suggest that the joints are primarily of the 'ac' type.

Pre-Columbian palinspastic reconstruction

Palinspastic reconstructions removing effects of the Columbian Orogeny show possible configurations of the unconformity at the base of the Black Stuart Group prior to Columbian deformation. Cross-section E-E' and the Kimball Mountain area from cross-section D-D' were used to make the palinspastic sections of Figure 33.

Two reconstructions are presented, each with a different horizontal datum. Figure 33B has the horizontal datum at the base of the Mural Formation to satisfy the hypothesis of no pre-Black Stuart deformation. Pre-Black

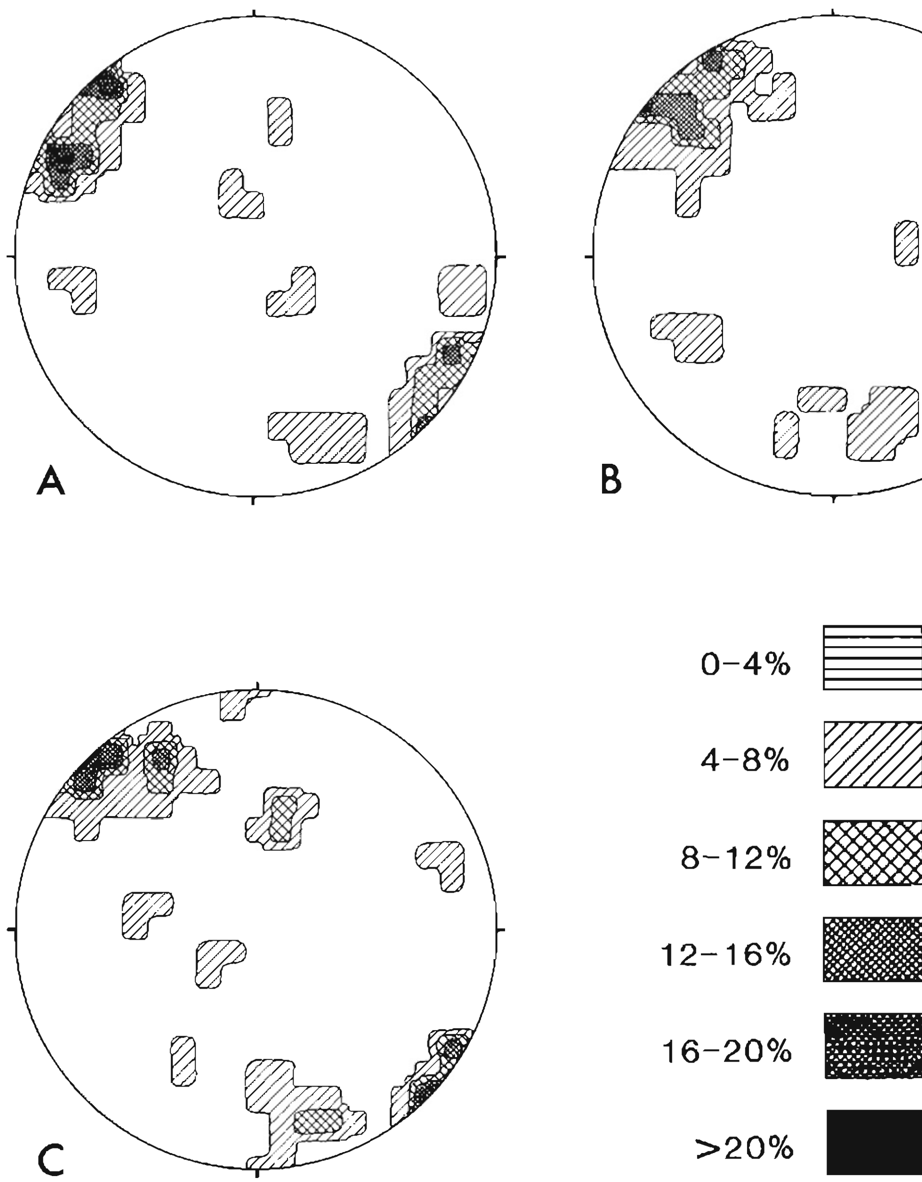


Figure 31. Equal area net plot of the poles to joints within rocks of the Kaza and Cariboo groups of the Cariboo Terrane. A, B and C are averaged plots from domains 2, 3 and 4 (Fig. 25) with 22, 31 and 26 points respectively.

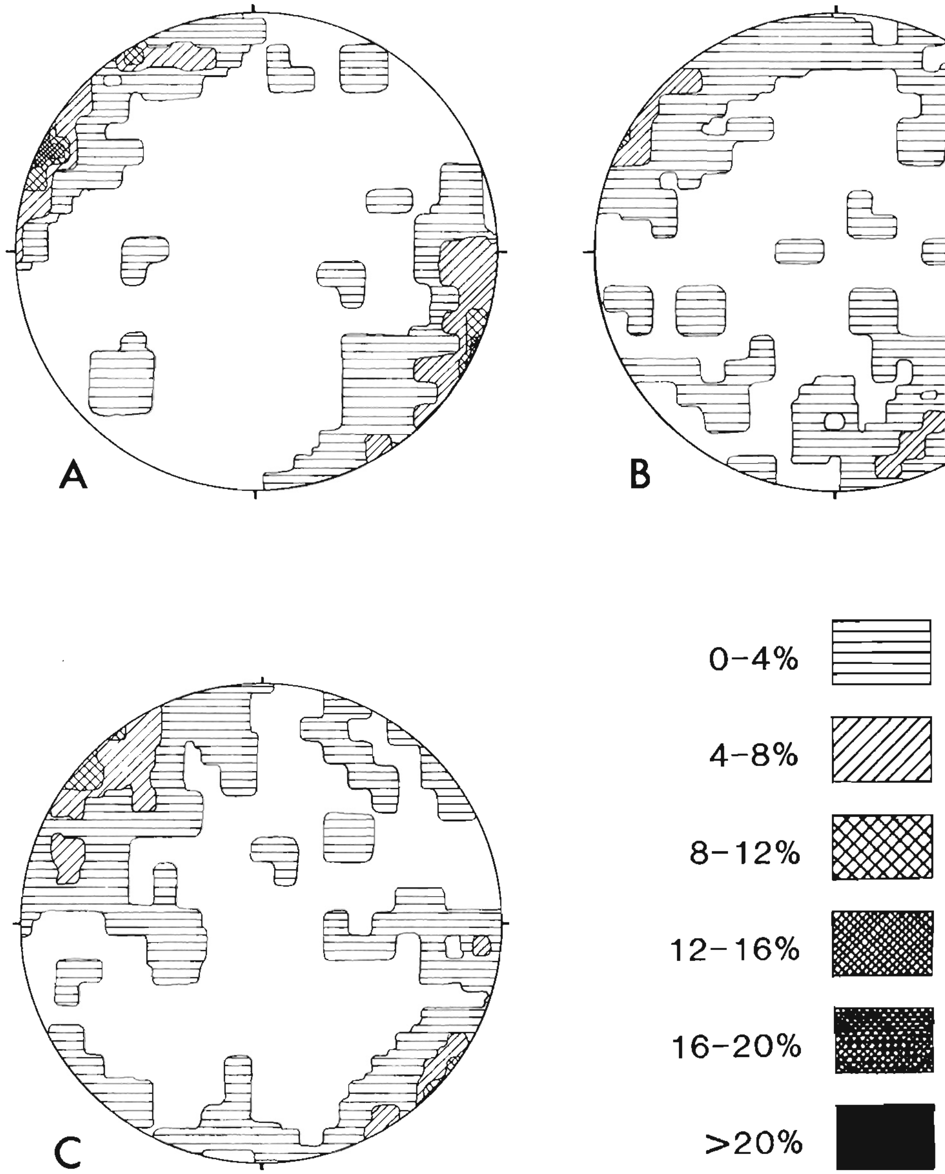


Figure 32. Equal area net plot of the poles to joints within rocks of the Black Stuart Group, and Guyet, Waverly and Greenberry formations of the Cariboo Terrane. A, B and C are averaged plots from domains 2, 3 and 4 (Fig. 25) with 67, 93 and 85 points respectively.

Stuart deformation and peneplanation is satisfied in Figure 33A which has the basal Black Stuart unconformity as the horizontal datum.

Restoration required that several areas be approximated as to their bed length because it was not fully displayed within the cross-section. These areas are on the northeast bank of Sunshine Creek[1] and on the ridge system between Little River and Sunshine Creek. Estimates of bed length were made as short as possible to minimize shortening and maximize the topography of the pre-Black Stuart unconformity.

The palinspastic reconstructions (Fig. 33) show considerable variation in stratigraphic thicknesses along the line of section. These thickness changes are partly a result of Columbian tectonics, but are also in part depositional variations. Because shortening has been minimized the thickness variation is accentuated. Stratigraphy generally thins to the west with a prominent arch in the area now northeast of Sunshine Creek.

[1]Informal name used in this report for the first tributary draining southeastward into Separation Creek.

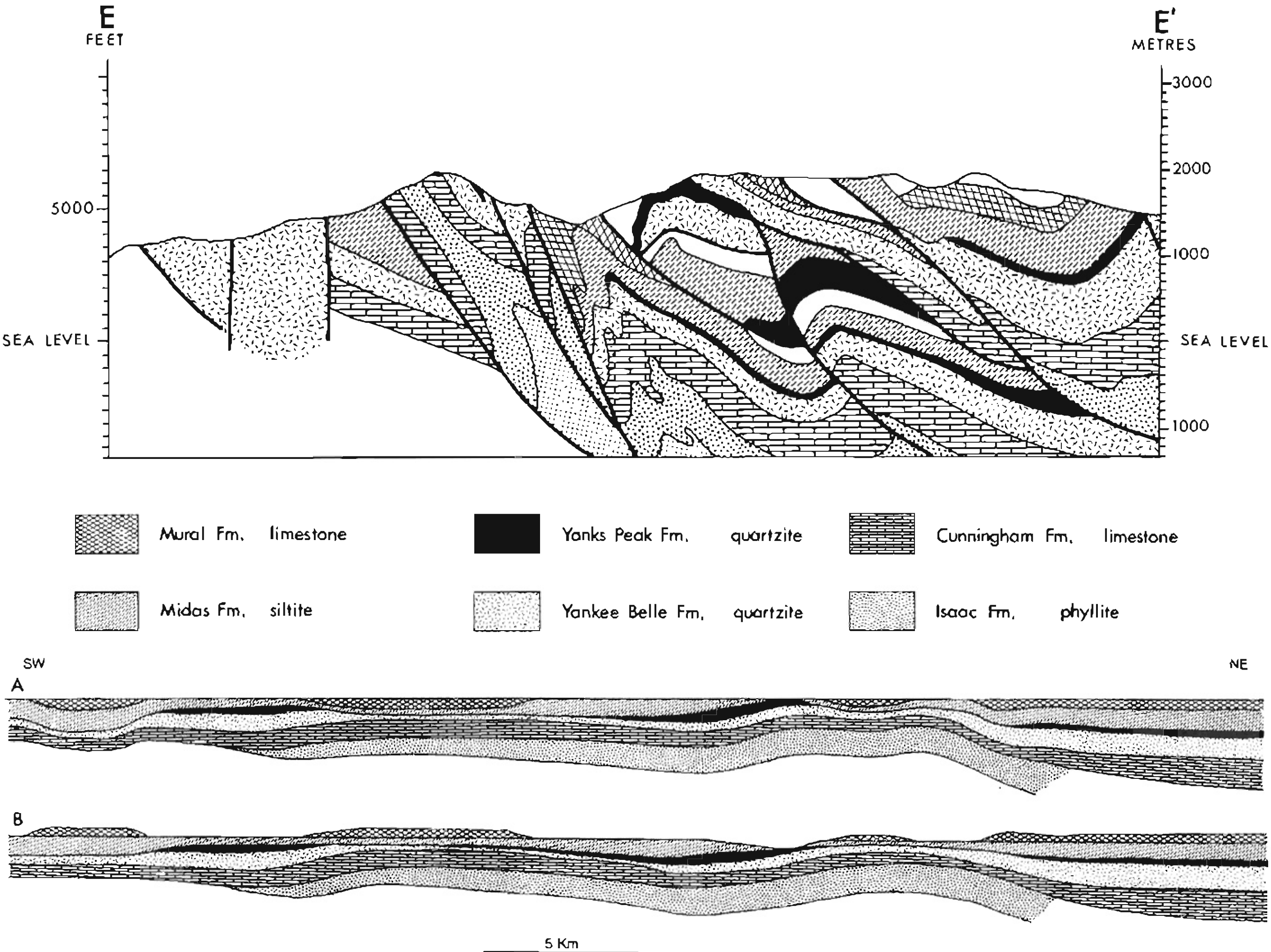


Figure 33. Structural cross-section E-E' from near Black Stuart Mountain palinspastically restored to before the Columbian Orogeny. Palinspastic section 'A' assumes the unconformity at the base of the Black Stuart Group is horizontal. Palinspastic section 'B' assumes the base of the Mural Formation is horizontal.

In terms of Columbian tectonics the palinspastic section demonstrates approximately 70% shortening in the area between Kimball Mountain and the hill northeast of Little River. This is the minimum amount of shortening possible with the structures of cross-sections E-E' and D-D'. The majority of this shortening is accommodated by thrusting.

The section in Figure 33A assumes pre-Black Stuart tectonism and peneplanation. It shows that pre-Black Stuart deformation must have been mild in the area of cross-sections E-E' and D-D'. The possible deformation is minor enough to be confused with the disconformable situation depicted in Figure 33B.

It is possible that the stratigraphic relationships at the unconformity are due to block faulting. Areas where the highest level of stratigraphy is preserved would represent downfaulted blocks. Such a block faulted regime would indicate a pre-Black Stuart tensional event. Then, three possible explanations for the mode of formation of the pre-Black Stuart unconformity configuration are: 1) tension, causing a block faulted regime, uplift and subsequent peneplanation, 2) uplift and differential erosion causing disconformable conditions, and 3) compression, causing broad-scale warping, uplift and subsequent peneplanation.

Northwestward from the palinspastic section the Black Stuart is known to cut downsection from Mural Formation to Yankee Belle Formation. This is not a gradual northwestward bevelling, but one of low topography.

The unconformable relationships thus far described are at odds with the conclusions of Johnston and Uglow

(1926) and Sutherland Brown (1957,1963). They postulated that a major deformational event occurred prior to the deposition of the Guyet Formation. The deformation was considered Devono-Mississippian because of the suspected Carboniferous age of the Guyet Formation. It was named the Cariboo Orogeny by White (1959). From the revised stratigraphic relationships of this study, the major unconformity is not at the base of the Guyet Formation, but at the base of the Black Stuart Group. The unconformity is Ordovician and does not represent a major deformation within the Cariboo Terrane of the area and it therefore does not satisfy the definition of the Cariboo Orogeny.

Stratigraphic control of structure

Pre-existing structures appear to influence the formation of superimposed structures. Pre-existing structures include stratigraphic features such as lithology, thickness change and bedding orientation and conventional tectonic structures. Within the area structural control is dominated by the competency of Yanks Peak Formation quartzite, the discontinuity of Yanks Peak Formation quartzite, the "arches" due to thickness changes and the broad-scale warping that occurred during the lower Ordovician. First, the structural features seen in the cross-sections of the map area will be described in relation to the above mentioned controls. Then, structural controls will be discussed with the aid of a clay model.

The most prominent feature of the palinspastic reconstruction of cross-section E-E' is the "arch" which is now located on the ridge northeast of Sunshine Creek. On the west flank of this arch is a thick wedge of Yanks Peak Formation quartzite. The area of the arch is a locus for thrusts which carry the thick stratigraphic package underlying Kimball and Anderson ridges. Mural Formation is preserved beneath the unconformity both west and east of the lens of Yanks Peak Formation quartzite in the western part of the palinspastic sections (Fig. 33). The thrust carrying Cunningham Formation over Mural Formation cuts through the stratigraphy west of the Yanks Peak Formation quartzite lens and on the eastern side of the remnant of Mural Formation. The thrust appears to have been localized between these two competent units.

Cross-section C-C' shows the faulted fold pair of the Roundtop Mountain area. There are two prominent features in this area; the rapid westward downcutting of the unconformity and the local abundance of Yanks Peak Formation quartzite. The ductile phase thrusts superimpose the quartzite and the described unconformity. The thrust relationship is like that described for the western thrust of cross-sections E-E' and D-D'. The thrust cuts upsection at the western edge of the Yanks Peak Formation quartzite lens. With a horizontal datum for the base of the Black Stuart Group, the western limb of the present syncline would have existed as an east-dipping monocline prior to the Columbian event.

Clay models were made to have two principal features. They had angular unconformities and layers of different thickness and viscosity. These features are suspected to influence the formation of deformational structures. The clay models were placed in a vise and slowly squeezed horizontally. One experiment is shown through various stages of deformation in Figure 34. The series of pictures in Figure 34 is only meant to illustrate the idea that tectonic structures are controlled by pre-existing features. The experiment is not meant to prove the point, only to illustrate it. Figure 34A shows the configuration of layers prior to squeezing. An anticline-syncline pair with a larger scale anticline is cut off by an unconformity. Of the three viscosities represented the intermediate grey is the most viscous, the darkest grey the least viscous and the lightest grey the moderately viscous. Following the series of pictures in Figure 34 it is obvious that: 1) the thrusts initiate in the vicinity of the hinge zone of the anticline or "arch", 2) the initial fold pair becomes accentuated, and 3) the final faulting and folding is concentrated on the limbs of the initial central syncline. Figure 35, a cross-section of the model, shows the control the most viscous layer had on the position of the thrust. The thrusts border exactly the area where the most viscous layer is in contact with the unconformity. A peculiarity of this feature is that beds separated by the lower thrust appear not to match.

It is suggested that stratigraphic arches react similarly to the syncline of the experiment and that the combination of arches and pre-existing "warps" of the Ordovician event governed the position and scale of the features produced during the Columbian Orogeny.

ROCKS OF THE BARKERVILLE TERRANE

The Barkerville Terrane is bounded on the east by the Pleasant Valley Thrust across which it adjoins the Cariboo Terrane, and on the west by the Slide Mountain and Quesnel terranes (Fig. 2). It is underlain by an unknown basement and overlain by the tectonically emplaced Slide Mountain Terrane. The Barkerville Terrane continues beyond the map area northwest to near Prince George and southeast to near Clearwater. It is treated separately from surrounding rocks because it has a unique stratigraphic succession and is bounded by faults.

It is recommended that most existing names for units within this terrane be abandoned. The strata are here divided into one formal and several informal units (Table 14). The divisions are mainly informal because of uncertainties concerning stratigraphic order. The Snowshoe Group is the formal unit; includes most rocks of the Barkerville Terrane and has fourteen informal subdivisions: Ramos, Tregillus, Kee Khan, Keithley, Harveys Ridge, Goose Peak, Agnes, Downey, Eaglesnest, Bralco, Hardscrabble, unnamed carbonate, Island Mountain and Tom. The Permian Sugar limestone (Orchard and Struik, 1985) rests with unknown relationship on the upper part of the Snowshoe Group.

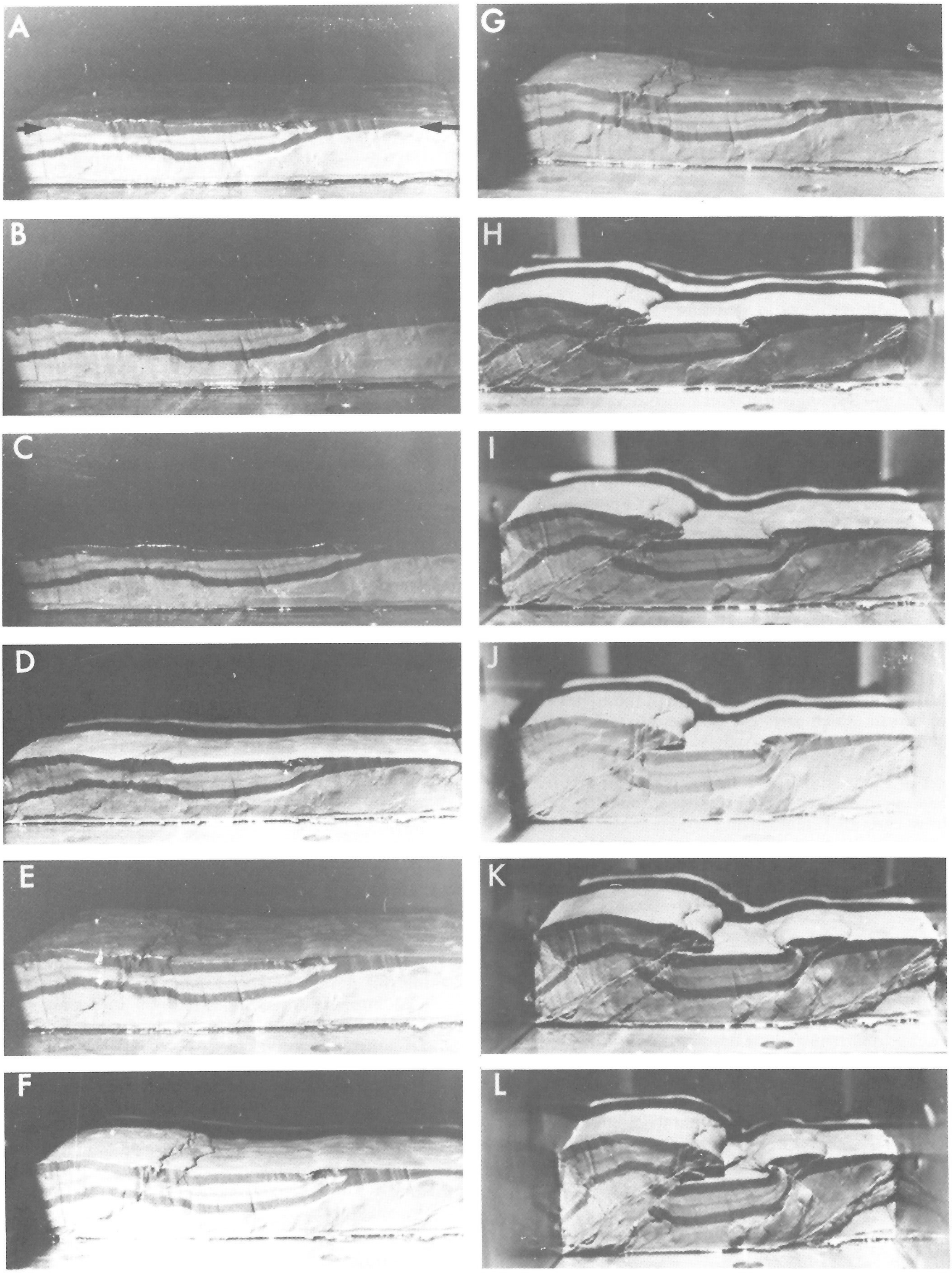

A
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Figure 35. Cross-section sawn through clay model of frame L of Figure 34, shown looking from the opposite direction. Note that the thin most viscous layer (Yanks Peak Formation; arrow) is cut by thrust faults exactly at either end of where it directly underlies the unconformity beneath the least viscous layer (Black Stuart Group). (GSC 191028)

Rocks of the Snowshoe Group resemble those of the Horsethief Creek Group of the Cariboo Mountains and much of the Eagle Bay Formation near Barriere.

Bowman (1889) included all rocks of the Barkerville Terrane in the Cariboo Series. Johnston and Uglow (1926), Hanson (1935) and Lang (1938) subdivided the Cariboo Series into various formations. Holland (1954) changed Cariboo Series to Cariboo Group and completely revised the stratigraphy, introducing new formation names. Sutherland Brown (1957, 1963) extended the usage of Cariboo Group eastward to the Cariboo Mountains; added a new formation from strata found in the Cariboo Mountains, and suggested that the group was mostly lower Paleozoic. Campbell (1961, 1963) applied the terminology unrevised in the Quesnel Lake map area. Later he (Campbell et al., 1973) revised the sequence of strata in the Cariboo Group by: 1) adding two new formations, 2) renaming some of the rock units as newly established Devonian Black Stuart Formation, 3) mapping the former uppermost formation of the Cariboo Group (Snowshoe Formation) as the Hadrynian Kaza Group, and 4) suggesting most of the Cariboo Group was Hadrynian and Cambrian. Campbell et al. (1973) suspected that rocks of the Cariboo Group whose type sections were on Yanks Peak might not belong to the Cariboo Group as used by

Figure 34. Opposite. Sections A to L, cut through a deformed clay model to illustrate the effect of pre-existing structure, in this case an anticline-syncline pair beneath an unconformity (under the upper dark layer), on the formation of new structures. The layers marked by shades of grey have different viscosities; the lightest is of intermediate viscosity (micaceous quartzites and limestone representing the Kaza Group, Yankee Belle, Midas and Mural formations), the middle tone is most viscous (quartzite representing the Yanks Peak Formation) and the darkest tone is least viscous (pelite representing the Isaac Formation and Black Stuart Group). Note that thrust faults root in the cores of the anticlines and that they generally enhance the amplitude of the pre-existing fold. (GSC 191027)

Table 14. Table of formations for the Barkerville Terrane

PALEOZOIC	LOWER PERMIAN		*SUGAR limestone (0-10 m)*	*Grey crinoidal limestone*
				Unconformable on Hardscrabble Mountain; not in contact with Island Mountain
	UPPER PALEOZOIC	SNOWSHOE GROUP	*ISLAND MOUNTAIN AMPHIBOLITE (< 150 m)*	*Amphibolite, tuff and siliceous mylonite*
				Fault contact on Eaglesnest; not in contact with Hardscrabble Mountain
			HARDSCRABBLE MOUNTAIN (≤ 150 m?)	*Black siltite, argillite and muddy granule conglomerate*
				Unconformity?
	PALEOZOIC		*BRALCO (< 100 m)*	*Grey limestone, locally pelletal, commonly marble, includes undifferentiated phyllite*
				Conformable with Downey; not in contact with Eaglesnest
			EAGLESNEST (≥ 150 m)	*Grey and olive micaceous feldspathic poorly sorted quartzite and phyllite*
				Lateral equivalents?
			DOWNEY (≥ 150 m)	*Olive-grey micaceous feldspathic poorly sorted quartzite and phyllite, marble, metabasaltic volcaniclastics*
				Conformity
			AGNES (< 60 m)	*Light grey conglomerate in part with calcareous matrix*
				Lateral equivalents
			GOOSE PEAK (< 250 m)	*Light grey poorly sorted quartzite, phyllite, minor black siltite*
				Conformity?
			HARVEYS RIDGE (≤ 300 m)	*Black micaceous poorly sorted quartzite, siltite and phyllite; minor muddy conglomerate, limestone and basaltic metavolcaniclastics*
				Unconformity?
PRECAMBRIAN	HADRYNIAN		*KEITHLEY (≤ 300 m)*	*Light grey quartzite; olive micaceous poorly sorted quartzite, siltite and phyllite*
				Conformity?
			KEE KHAN (≤ 75 m)	*Marble, olive phyllite, sandy marble*
				Conformity
			TREGILLUS (> 400 m)	*Olive-grey micaceous poorly sorted feldspathic quartzite and phyllite, conglomerate*
				Lateral equivalents?
			RAMOS (> 300 m)	*Olive micaceous poorly sorted feldspathic quartzite and phyllite, black siltite and phyllite, amphibolite, marble, minor basaltic and felsic volcaniclastics*
				Not in contact
	HADRYNIAN OR PALEOZOIC		*TOM (≥ 175 m)*	*Olive-grey micaceous poorly sorted feldspathic quartzite, phyllite and schist; quartzose mylonite*

The thickness of units is approximate and may vary substantially from the given estimates.

them in the Cariboo Mountains. They essentially redefined the Cariboo Group providing reference sections in the Cariboo Mountains. The Cariboo Group of Yanks Peak and of Cariboo Mountains are two different sequences, and it is proposed that the Cariboo Group be defined by the reference sections in the Cariboo Mountains as presented by Campbell et al. (1973) and that the sequence at Yanks Peak be called the Snowshoe Group.

Igneous intrusions of the terrane consist mainly of diorite and gabbro sills with less quartz porphyry rhyolite.

All rocks have been regionally metamorphosed to low and middle greenschist facies.

Stratified rocks

Snowshoe Group

Ramos succession

The Ramos succession is mainly interbedded micaceous quartzite and phyllite with subordinate siltite, amphibolite, marble and tuff. Very coarse grained, very micaceous olive quartzite is characteristic of this unit of the Snowshoe Group. Tuff and siltite appear high in the section whereas marble and amphibolite are mainly low. The Ramos is diverse because the subordinate rock types are discontinuous and poorly exposed. The succession is intensely sheared at its contact with the western Crooked Amphibolite of the Slide Mountain Terrane. The Ramos may be in whole, or in part, related to the Tregillus.

The Ramos succession underlies the western border of the map area from Ramos Creek southeast to Keithley Creek. The most varied assemblage occurs near Cariboo Mountain. Ramos rocks were previously included in the Snowshoe Formation by Campbell (1963), Campbell et al. (1973) and Campbell (1978). There is no designated type section for this succession, however, exposures in road cuts along Ramos Creek, on Sovereign and Cariboo mountains and in Keithley Creek are representive of much of the upper part of the unit. The lower sections are only poorly exposed in the low lying areas near Fontaine Creek, and Little Swift and Swift rivers. The total thickness of the unit is unknown, but it exceeds 300 m.

The base of the Ramos succession is not exposed. Contacts with other units are suspected to be faults except along Keithley Creek where it may stratigraphically underlie or interfinger with the Keithley succession.

Interbedded quartzite and phyllite are the most abundant rocks within the Ramos succession (Fig. 36). The quartzite is olive to grey on fresh surfaces, weathering grey to olive grey, is poorly sorted, and on average is medium to coarse grained. Almost everywhere it is micaceous with the notable exception of a sequence on Sovereign Mountain. Quartz clasts are glassy clear, grey and minor blue. In thin section the quartz grains are polycrystalline, commonly with 3 or more crystals per grain, but the coarser the grain, the fewer the crystals. Many grains have microcrystalline tails or a concentration of small crystals on their edges. The tails parallel the foliation of the muscovite matrix. All but the finest recrystallized quartz has undulatory extinction. Potassium and plagioclase feldspar content varies greatly from place to place but is everywhere less than 10%; averaging approximately 3%. The potassium feldspar, the predominant feldspar, is microcline and orthoclase. Beaded perthite comprises about half of the orthoclase. The plagioclase is oligoclase or andesine; varying from An_{25} to An_{35}. The feldspars are partly altered (1-90%) to sericite, epidote and calcite. Zircon is the primary accessory mineral. Metamorphic muscovite, chlorite and biotite occur up to 45% defining several overlapping metamorphic foliations. Metamorphic garnet, ankerite, siderite, calcite, pyrite and tourmaline occur locally.

Figure 36. Isoclinally folded fine grained micaceous quartzite of the Ramos succession of the Snowshoe Group on the logging road northwest of Keithley Creek (Canadian quarter for scale). (GSC 191029)

The phyllite is olive with lesser amounts of grey. Near Cariboo Mountain equivalent rocks are olive-grey schist. The phyllite consists of fine grained quartz, white mica, chlorite and accessory pyrite, ankerite, siderite and epidote. The quartz is in trains of less than 0.2 mm sutured crystals that follow the foliation of the rock. Albite is in some places intergrown with the quartz. The white mica is commonly crenulated and overgrown by 0.5 mm muscovite and chlorite. The schist found along the Swift River and on Cariboo Mountain is characterized by 0.2 to 0.5 mm muscovite, chlorite and biotite overgrowing the sericite foliation and which are in turn deformed. Garnet is a common accessory mineral in the schist.

Contacts between the quartzite and pelite are sharp and bedding is generally rhythmical. As with the Tregillus, areas of non-micaceous quartzite form thick massive sequences devoid of pelite. Graded bedding occurs locally, most commonly as the decrease of the grain size of all fractions but in some places as a decrease of grain size in only the coarse outsize fraction; rare occurrences of channel scour were noted.

Black siltite and very fine grained quartzite interbedded with black phyllite are scattered throughout parts of the belt of Ramos rocks. They are similar to rocks of the Harveys Ridge succession.

Amphibolite and associated marble and calcareous amphibolite are common in the area of Cariboo Mountain and Swift River. The amphibolite is olive, being spotted with white feldspar or marble to varying degrees. It is strongly foliated and almost invariably the foliation is refolded (Fig. 37). It is composed mainly of actinolite where the amphibolite is devoid of calcite and of tremolite where calcite is present. Epidote, albite, chlorite, sphene and opaques are the most common minor constituents. The actinolite and tremolite crystals are less than 2 mm long in most places. Larger ones occur where the amphibolite is associated with garnet-bearing schist as

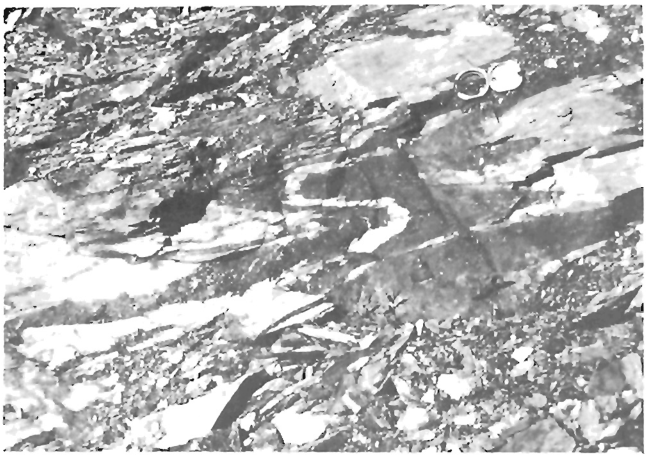


Figure 37. Superimposed folds in amphibolite within the Ramos succession of the Snowshoe Group in a logging clear cut south of Little Swift River near its junction with Swift River. (GSC 191030)

near Swift River. The oriented amphibole and the lamellae of mineral segregations define the rock foliation. Chlorite occurs as sheaths and mats rather than as oriented crystals.

Contacts of the amphibolite with bounding quartzite and schist are sharp. Locally the amphibolite contains thin interbeds of light grey medium grained quartzite and schist. Minimum thicknesses of the amphibolite are in the order of 1 to 10 m; variations and total extent of the amphibolite horizons are unknown.

Several impure marble units occur south of the Fontaine Creek area. The marbles are dark grey, weather buff and light grey, and are composed of calcite with impurities of micas and quartz. They are interlayered with grey and silvery schist or phyllite and are in sharp contact with olive micaceous quartzite and phyllite. Individually they are discontinuous and less than 50 m thick.

Chlorite phyllite along Ramos Creek, and in scattered exposures south to Swift River is interpreted as metamorphosed tuff. In the section along Ramos Creek the tuff occurs near the top and consists of 1 to 2 m thick beds in black and olive phyllite and fine grained quartzite. Along Keithley Creek tuff is interlayered with dark grey and olive phyllite near the upper contact of the Ramos with the Harveys Ridge succession.

Other minor rock types included in the Ramos succession are rhyolitic tuff near Sovereign Mountain, and micaceous quartzite with coarse grained dètrital muscovite found near Eskeridge and Weaver creeks.

Because the quartzite-pelite sequences of the Ramos succession have similar sedimentary characteristics to those of the Tregillus rocks they are thought to have been deposited in a similar environment under similar conditions and possibly from the same source terrain. Different from the Tregillus, however, is the occurrence of amphibolite, marble, black siltite and tuff. The tuff of the Ramos indicates volcanic activity, which is absent from the Tregillus except possibly where it is in contact with the Kee Khan marble. The association of marble with some of the amphibolite suggests that the amphibolite may be derived from impure carbonate. Quartzite-pebble conglomerate in the Ramos near the settlement of Keithley Creek is similar to that of the Tregillus.

The age of the Ramos succession is assumed to be Hadrynian, coeval with the similar Tregillus succession (whose age also is unknown) which may be its lateral equivalent. As such, it may also correlate with the Horsethief Creek Group of the Selkirk Mountains.

Tregillus clastics

The Tregillus clastics are a sequence of micaceous quartzite, phyllite, schist and orthoquartzite with a quartzite clast conglomerate near its top. The conglomerate and contact relations with the overlying Kee Khan marble are the only distinguishing features. The interbedded quartzite and pelite are much like those of the Ramos, Eaglesnest and Tom successions.

Tregillus clastics underlie an area in the northwest corner of the Wells map area (Map 1635A, in pocket); mainly north of Highway 26 and west of Willow River. They were previously mapped as Kaza Group (Snowshoe Formation; Campbell et al., 1973) and are continuous with some of the rocks mapped as undifferentiated Cariboo Group in Prince George map area (Tipper, 1961). They may be wholly or partly equivalent to the Ramos succession. Metamorphism in the unit varies laterally near the garnet isograd. Retrogressive metamorphism is common throughout the sequence. The minimum thickness is approximately 400 m.

The base of the Tregillus rocks is not exposed. The sharp, conformable upper contact with the Kee Khan marble is placed at the first calcareous rocks above the quartzite conglomerate.

The Tregillus consists of interbedded quartzite, pelitic rock and minor conglomerate. The quartzite weathers grey and brown, and is grey and light grey on fresh surfaces. It is mostly medium- to coarse-grained and poorly sorted with variable amounts of matrix mica. Quartzite with little to no matrix mica is lighter grey than other varieties. Quartz clasts are mostly glassy clear and a few are blue. Potassium and plagioclase feldspar are present locally up to 10%. Accessory minerals include zircon and rare rutile. Metamorphic muscovite, chlorite and biotite occur up to 30%, with the muscovite defining the rock foliation which is generally overgrown by biotite and chlorite. Metamorphic garnet and tourmaline occur locally. The pelite interbedded with the quartzite varies from dark grey to grey and olive grey. It consists mainly of fine white micas and quartz. Metamorphic biotite and garnet are common near Tregillus and Aura Fina creeks.

Bedding is even and rhythmical with quartzite beds averaging 40 cm in thickness and pelite beds generally thinner. Contacts between quartzite and pelite are sharp.

In some places the non-micaceous quartzite forms thick massive sequences devoid of pelite; the best example of this is on the hill northwest of Tregillus Creek.

The conglomerate is light grey and light brown. It consists of pebbles and granules of white moderately coarse grained well sorted quartzite in a poorly sorted matrix of sand-sized clear and blue glassy quartz (Fig. 38). Granule conglomerate is more abundant than coarser grained varieties. The percentage of quartzite clasts decreases with decreasing grain size. The conglomerate is less than 15 m thick and is usually 1 to 5 m thick. It is overlain by a thin sequence of quartzite and pelite in turn overlain by the calcareous rocks of the Kee Khan marble.

The poor sorting and local graded bedding of the quartzite and its even and rhythmic interbedding with the pelite suggest the rocks were deposited as proximal turbidites (examples of which were shown by Walker, 1978, Fig. 2, and by Poulton and Simony, 1980, Fig. 7 and 8). The lack of crossbedding and sole markings also indicates a proximal position on a submarine fan complex. The massive sequences of quartzite may be channels within the complex.

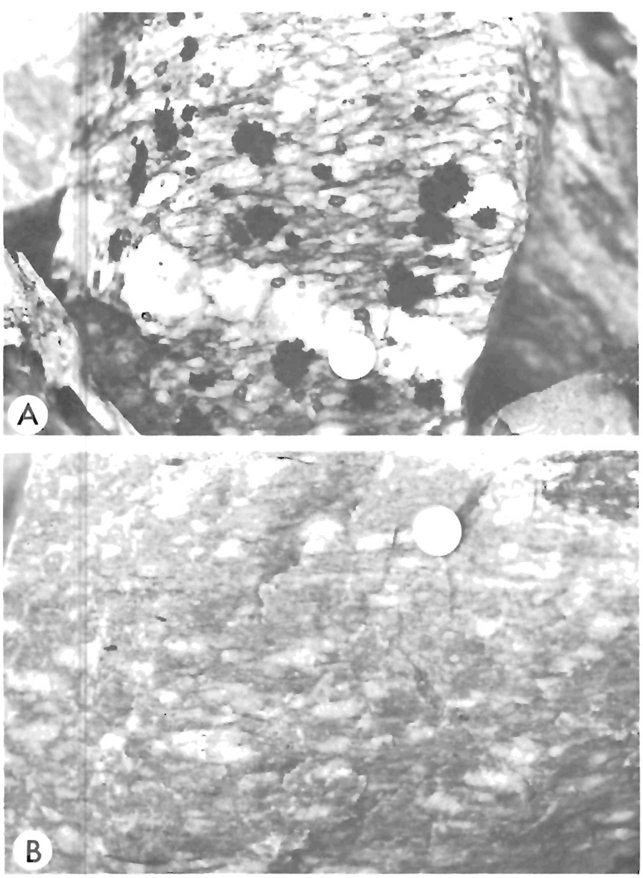


Figure 38. Conglomerate from the upper part of the Tregillus succession of the Snowshoe Group. The clasts are white quartzite. The coin scale in each frame is a Canadian dime. A (GSC 191031) and B (GSC 101032) come from the southwest slope of the ridge north of Lightning Creek at Pinegrove Creek (presently a ski resort).

The age of the Tregillus is unknown. The rocks are similar to those of the Hadrynian Windermere Supergroup, particularly the Horsethief Creek Group (*see* Poulton and Simony, 1980) and as contradictory evidence is lacking, they are correlated with that group.

Kee Khan marble

The Kee Khan marble is a thin discontinuous sequence of marble, calcareous quartzite and phyllite. It is distinguished from other marble horizons by its magnetite-bearing, chrome-green phyllite and position above the conglomerate of the Tregillus, but otherwise the grey calcareous rocks of the Kee Khan are similar to others of the Snowshoe Group.

The Kee Khan marble is limited to the northwest corner of the Wells map area, following the distribution of the Tregillus clastics. The Kee Khan was previously mapped as Kaza Group (Snowshoe Formation) by Campbell et al. (1973). Marble and phyllite of the unit form the headwall of Kee Khan Creek and a base for the top of the ski lift of the Troll Resort at Pinegrove Creek. This exposure and those in bulldozer trails on Pinegrove Mountain are representative of the Kee Khan marble at its thickest, approximately 75 m.

The Kee Khan marble conformably overlies micaceous quartzite of the Tregillus clastics. The contact is generally sharp and is drawn at the first calcareous quartzite or chrome-green phyllite above the quartzite conglomerate of the Tregillus.

Grey foliated marble is the dominant lithology of the Kee Khan rocks. It weathers orange, brown or grey. It is composed of 0.5 to 1.5 mm calcite crystals with subordinate quartz, micas and minor feldspar. The calcite and quartz are equant; the muscovite, chlorite and biotite define foliation surfaces. The marble occurs in less than 10 m thick layers separated by calcareous and noncalcareous phyllite and quartzite.

Phyllites of the Kee Khan are characteristically chrome-green and olive-grey and grey varieties. The chrome-green phyllite consists of chlorite, sericite, quartz, epidote and minor magnetite. In places they are calcareous and gradational with marble and, in others, sandy and gradational with phyllitic quartzite. Locally thin layers of orange weathering marble interbed with the phyllite (Fig. 39). The olive-grey and grey phyllite is generally not calcareous and occurs as discrete interbeds with the marble and quartzite.

The quartzite is poorly sorted, medium- to coarse-grained, micaceous and commonly calcareous. Quartz and feldspar clasts are the same types as those in the Tregillus. The micaceous matrix is chlorite-rich where the quartzite grades into the green phyllite.

The chrome-green phyllite is interpreted as derived from both a terrigenous clastic and volcanic source. The

Figure 39. Isoclinally folded impure orange weathering marble and green pelite of the Kee Khan succession of the Snowshoe Group on the logging road on the east side of the foot of Tregillus Lake. (GSC 191033)

quartz and sericite suggest a clastic source and the abundant chlorite and epidote a volcanic one. The quartzite represents the continuation of the sedimentary conditions of Tregillus time interrupted by the anomalously quiet carbonate deposition.

As with the Tregillus clastics the age of the Kee Khan marble is unknown. Because the marble unit is conformable with the Tregillus it is part of the same tectonic interval and therefore is considered Hadrynian and correlative with the Windermere Supergroup.

Keithley succession

The Keithley clastics consist of phyllite, fine micaceous quartzite and orthoquartzite. The orthoquartzite is unique among units of the Snowshoe Group but may be confused with Yanks Peak Formation orthoquartzite east of the Pleasant Valley Thrust. The olive and olive-grey phyllite is similar to others of the group but is most continuous in, and defines, this unit. The phyllite could be confused with that of the Yankee Belle Formation east of the Pleasant Valley Thrust. The Keithley clastics wedge out laterally in places as the facies changes to coarse grained micaceous quartzite.

The Keithley clastics lie along the western part of the map area from Pinegrove Mountain southeastward to Cariboo Lake. They were previously mapped as the Yankee Belle and Yanks Peak formations by Holland (1954) and Campbell (1963) and as Snowshoe Formation by Campbell et al. (1973). These rocks are renamed to avoid confusion with the Yankee Belle and Yanks Peak formations east of the Pleasant Valley Thrust with which they have been previously correlated. It is argued, partly on lithological difference, but mostly on differences in bounding stratigraphy that the sequences separated by the fault are not the same. The Keithley rocks are named after Keithley Creek that drains the Snowshoe Plateau, an area where the rocks are relatively well exposed. A reference locality is the abandoned type section of the Yankee Belle and Yanks Peak formations on Yanks Peak as described by Holland (1954). Thickness of the unit is less than approximately 300 m throughout the area.

The lower contact of the Keithley succession is poorly understood. The Keithley overlies the Kee Khan marble at Dilman Creek, suggesting stratigraphic continuity from the Tregillus through to the Keithley. However, near Keithley Creek the Keithley overlies the Ramos. The contact is not well exposed, therefore it is not known whether either of the relationships is a stratigraphic one. The Keithley is underlain by a marble unit 3.5 km northeast of Cariboo Mountain. The black rocks that underlie the marble are anomalous for the Ramos succession, resembling more the rocks of the Harveys Ridge succession. Thinly laminated dark grey fine quartzite and siltite included in the Keithley near Wingdam are uncharacteristic of the unit and are only included in the succession by default. Their contact relationships with other rocks of the area are unknown.

Olive to grey phyllite and fine grained quartzite characteristic of the Keithley can be found on Yanks Peak, north-northeast of Cariboo Mountain, on Pinegrove Mountain and near Dilman Creek. The rock weathers olive grey or light brown; sometimes rusty. The phyllite and quartzite interbeds are thin or medium thick and commonly indistinct. Locally coarse grained micaceous quartzite beds (less than 50 cm thick) are interbedded with the finer rock. The coarse grained quartzite resembles that of the Ramos succession. At French Snowshoe Creek the Keithley phyllite contains 1 to 4 mm thick interlaminae of white dolomite, ankerite and/or quartz.[1] They appear to be veinlets parallel with the schistosity. Here also are abundant 2 to 20 mm pyrite cubes scattered throughout the phyllite. Grey glassy quartz veins and pods occur in many exposures of Keithley rock and are most characteristic of the Keithley and underlying Ramos successions. The pods are lenticles of quartz vein intruded prior to translation along the schistosity. The only other unit in which the glassy grey quartz veins occur with any abundance is the Goose Peak.

The orthoquartzite of the Keithley is known confidently from Yanks Peak, north-northeast of Cariboo Mountain and on Pinegrove Mountain. It is white, light grey, pink or light brown. The quartz grains are well sorted, medium- to coarse-grained and glassy or white. The matrix is composed of fine grained quartz with little or no impurities of sericite. The orthoquartzite is massive with no internal sedimentary features such as grading or crossbedding. It is less than 15 m thick and forms very resistant rounded outcrops. Although commonly missing, where present, it forms the top of the unit. Holland (1954) mapped it as the Yanks Peak Formation.

The Keithley succession is underlain by various units suggesting: 1) it unconformably overlies them, 2) it is

[1]This feature is also described by Holland (1954, p. 17)

faulted against them, or 3) the underlying units undergo lateral changes. The lower contact is too poorly exposed to determine whether it is an unconformity or fault, but rocks near the contact do not show any deformation that may be associated with faulting. Lateral changes of the Ramos succession to rocks of the Keithley succession are indicated in several places.

East of Keithley Creek, the Keithley is overlain by the Harveys Ridge succession and underlain by the Ramos. West of Keithley Creek, Harveys Ridge-like rocks directly overlie the Ramos with no intervening Keithley. The omission may be due to a facies change from the Keithley phyllite to the Ramos quartzite; as noted by the presence of Ramos-type quartzite included within the Keithley phyllite along the lower part of French Snowshoe Creek, east of Keithley Creek. Northeast of Cariboo Mountain the Keithley olive phyllite contains beds of Ramos-type quartzite; unlike the occurrence at French Snowshoe Creek this one is near the top of the Keithley sequence. At Ramos Creek, Keithley-type phyllite is interbedded with the quartzite sequence of the Ramos and this sequence underlies Harveys Ridge-type black siltite. From these relationships the Keithley and Ramos are interpreted as partial facies variations of the same horizon. The Tregillus and Ramos successions may also be partly equivalent.

The age of the Keithley succession is unknown but may be Precambrian.

Harveys Ridge succession

The Harveys Ridge succession consists of black and grey siltite, micaceous quartzite, phyllite, limestone and minor dolostone. It is dominated by black and dark grey rocks. The interbedded black siltite, phyllite and micaceous quartzite with black quartz grains are characteristic of the unit, although all occur in other sequences, and in isolated outcrops could be confused with those from other successions within the Barkerville Terrane. Any of the quartzite coarse enough to show the poor sorting and type of quartz grains, however, cannot be confused with rocks of the Cariboo Terrane. The limestone and dolostone are restricted to the upper part.

The Harveys Ridge succession traverses the area from Willow River southeastward to Cariboo Lake. It has also been traced beyond the map area south to Ishkloo Creek. The Harveys Ridge rocks were previously mapped as Midas and Snowshoe formations by Holland (1954) and Campbell (1963) and as Snowshoe Formation by Campbell et al. (1973). The rocks are renamed to avoid confusion with the Midas Formation of the Cariboo Terrane and because the present definition includes rocks from various previously defined formations. As with the Keithley rocks, differences of lithology and bounding stratigraphy distinguish the Harveys Ridge from any unit east of the Pleasant Valley Thrust. The succession is named after Harveys Creek and Harveys Ridge where it is relatively well exposed. No type section is proposed for this unit although Harveys Ridge exposures are typical of the sequence. The type section of the Midas Formaton as defined by Holland (1954) at Yanks Peak is included in the Harveys Ridge succession. The thickness varies considerably, but is estimated to be everywhere less than 300 m.

The Harveys Ridge succession overlies, in the northwest, the Kee Khan marble and Keithley clastics, and in the southeast, the Ramos and Keithley. The contact that has been seen, with the Keithley orthoquartzite, is sharp although there is an insufficient area of it to determine whether it is an unconformity.

The black siltite is dense and featureless. It commonly contains 1 to 3 mm laminae of white quartz parallel to the bedding. It occurs in 2 to 20 cm beds separated by black or grey phyllite. Holland (1954, p. 19, 20) called this rock a silty quartzite. The siltite and phyllite are in places altered to light grey siltite and sericitic phyllite respectively. The alteration follows bedding in some places and in others cuts across the beds as a blotchy mottling. Pyrite, siderite and ankerite are common secondary minerals of the phyllite and less so of the siltite. Local silty phyllite displays poorly developed foliation. Green calcareous phyllite occurs rarely within the siltite-phyllite sequence.

Harveys Ridge quartzite can be black, grey or olive grey; pure or micaceous. The pure quartzite (orthoquartzite) is always black and is generally sorted and of medium grain size. It is featureless, resistant and commonly cut by numerous white quartz veins, some of which are vuggy. The micaceous quartzite characteristic of the Harveys Ridge is black or dark grey, poorly sorted, and of variable grain size. Quartz clasts are mainly clear, but up to 10% black and 1% blue glassy ones are common. In thin section the quartz can be seen to contain trails of fine bubbles. All the quartz grains larger than 0.3 mm have undulatory extinction and are broken into subgrains. Potassium and plagioclase feldspar may form up to about 8% of the rock but average less than 3%. The potassium feldspar is most altered, up to 30%. The olive-grey quartzite is much the same as that of the Ramos, Eaglesnest or Tom successions. A rusty brown weathering grey ankeritic quartzite occurs locally. Holland (1954, p. 20) described it in the vicinity of Yanks Peak where it appears to be confined. A similar rock is found in black siltite and phyllite at the head of Penny Creek.

Dark grey limestone forms a thin horizon near the top of the Harveys Ridge sequence. It is rarely present but can be found throughout the map area. The calcite crystals range from 0.1 to 1 mm across. The laminations that exist are metamorphic and nothing can be said about the original bedding. Holland (1954) suggested that exposure of the limestone may be sporadic because of partial removal below an erosional unconformity.

The sedimentary characteristics of the Harveys Ridge differ little from that of the Ramos or Tregillus quartzite-pelite sequences. It shows the same regularity of bedding, frequency of graded bedding and poor sorting which could denote these rocks as proximal turbidites. The sequence, however, is finer grained than the Ramos or Tregillus and therefore may represent more distal parts of a submarine fan. Contrary to this suggestion, however, is the absence of crossbedding and sole markings common in distal facies. Local sorted, medium grained sand must reflect either a change of environment or, more reasonably, sporadic influxes of clean sands possibly due to storm action on beach or longshore deposits.

The green calcareous phyllite was found only on the Snowshoe Plateau, at the head of Little Showshoe and Aster creeks and on Harveys ridge near the 'Jeep' trail to Keithley Creek. This rock resembles metamorphosed tuff. Either the volcanic activity these rocks represent was limited or far removed because the beds in all places are less than 3 m thick.

The age of the Harveys Ridge is unknown. It is assigned to the Paleozoic by correlation with the Eagle Bay Formation. The Harveys Ridge rocks resemble those of unit 3 of the Eagle Bay Formation at Forest Lake near Barriere (Preto et al., 1980; Preto, 1981). Although the correlation with the Eagle Bay Formation is made with some confidence, the age of the rocks of unit 3 is somewhat less secure.

Goose Peak quartzite

The Goose Peak is composed of coarse grained quartzite with minor amounts of fine grained quartzite and phyllite. The quartzite is characterized by its coarseness, low percentage of matrix mica and persistent feldspar content. Thick massive beds may be distinguished from the Keithley orthoquartzite by their poor sorting. The unit can be distinguished from other clean quartzite by the well defined 50 to 100 cm thick bedding. The unit is a lateral facies change of the Pine conglomerate.

The Goose Peak is located southeast of Mount Agnes to Cariboo Lake. It has also been traced south of the map area into the Goose Range. Within the map area the thickest accumulations occur on the ridge southwest of Pine Creek; however, it is thicker to the south. These rocks were previously mapped as the basal conglomerate of the Snowshoe Formation by Holland (1954), undifferentiated Snowshoe Formation of Campbell (1963) and as unit 4a by Struik (1982a). No type section for this unit has been designated; it is named after a reference area on Goose Peak south of the map area (Struik, 1983). The quartzite is lensoidal on a regional scale, attaining a maximum thickness estimated at 250 m.

The contact with the underlying Harveys Ridge rocks is sharp on the Snowshoe Plateau. At Mount Agnes the Goose Peak quartzite, overlies grey micaceous quartzite and phyllite in turn overlying black siltite assigned to the Harveys Ridge.

The Goose Peak quartzite is light grey or locally dark grey or pink, weathering grey or brown. It is made up predominantly of clear white and minor black quartz with up to 10% feldspar. It is poorly sorted and much is coarse- or very coarse-grained, some being granule conglomerate. The clasts are generally subrounded. The potassium feldspar is coarser grained, more abundant and more altered than the plagioclase. Minor detrital muscovite found in the matrix also is part of some quartz-feldspar grains. The matrix is composed of sericite and fine grained quartz and minor plagioclase. The quartz and some of the plagioclase is secondary.

Phyllite of the Goose Peak is mainly olive grey and minor black. The olive-grey variety is like that of the Keithley unit. It occurs as thicker beds than the black and is associated with the finer quartzite of the Goose Peak. Some of the olive phyllite is silty. The presence of olive-grey phyllite increases to the south where it is associated with the same type of grey quartz veins as those of the Keithley succession. The black phyllite occurs as thin interbeds with the coarse grained quartzite.

The quartzite occurs as 20 to 100 cm thick beds, rarely attaining 200 cm thick. The beds are defined primarily by the change in percentage of the coarse fraction of the quartz clasts. A fining upwards of the coarse fraction exists in some places, but mostly the individual beds are of a different average grain size. Quartzite with the olive-grey phyllite is generally more thinly bedded.

The poor sorting, coarse grain size, grading of the coarse fraction, and low volume of pelitic rock, suggest that these rocks represent a debris flow deposit, possibly distal to the laterally equivalent Agnes conglomerate. The sequences of quartzite with olive-grey phyllite may be the turbiditic deposits at the border of the debris flows, and the black phyllite, the accumulation in the debris flow channel during quiet periods. There is little evidence that the contact with underlying units is an unconformity and the influx of coarse detritus may simply represent a change in topography or climate in the source terrane.

None of the Goose Peak is dated, however, as it stratigraphically overlies the Harveys Ridge succession it too may be Paleozoic. No correlatives of the Goose Peak beyond the Quesnel Highlands are known.

Agnes conglomerate

The Agnes conglomerate consists of granule to boulder clasts of white to grey poorly sorted quartzite. The poorly sorted quartzite clasts appear to be unique to the Agnes conglomerate. Black phyllite forms minor thin beds within the conglomerate. The matrix, and rarely some of the clasts of the conglomerate, may be limestone. The Agnes conglomerate is a lateral facies variation of the Goose Peak quartzite.

The conglomerate is found mainly near Mount Agnes, on the Snowshoe Plateau and on the ridges flanking Pine Creek. Small exposures exist at Cornish Mountain and Mount Amador. The conglomerate was previously mapped as the basal Snowshoe Formation by Holland

(1954) and as unit 4a by Struik (1982a). The conglomerate is named after Mount Agnes near which a large area of it is exposed. Although no type section has been designated, exposures east of Pine Creek, on Harveys Ridge and near Mount Agnes are characteristic of the unit. The conglomerate is everywhere deformed and clasts have been elongated; locally the deformation has nearly obliterated the boundaries of the clasts. Where there are only quartzite clasts in a quartzite matrix the conglomerate may be difficult to identify. The unit is generally less than 30 m thick, possibly attaining 60 m in the vicinity of Pine Creek.

The Agnes conglomerate almost everywhere overlies the Harveys Ridge succession; an exception is at Cornish Mountain where it structurally overlies the Downey succession. In places the lower contact is sharp, with conglomerate directly on black phyllite or siltite or limestone, and in others interbedded with micaceous quartzite and phyllite, or a sequence of Goose Peak quartzite intervenes.

The conglomerate is light grey to grey and weathers similarly. It consists of quartzite, pelite and locally limestone clasts in quartzite, pelite or limestone matrix (Fig. 40). The quartzite clasts are sorted and poorly sorted white and dark grey orthoquartzite with lesser amounts of grey micaceous quartzite. The orthoquartzite is similar to that of the Keithley orthoquartzite except that it is not as well sorted. It is similar to the Goose Peak quartzite but is nearly always devoid of matrix mica and feldspar. Clasts of the orthoquartzite are medium- to coarse-grained and consist of glassy white and clear quartz. The micaceous quartzite consists of glassy clear, black and minor blue quartz grains which are on average medium grained. This quartzite resembles that of the Harveys Ridge and parts of the Eaglesnest successions. The pelite clasts are volumetrically of little importance consisting mainly of black and olive-grey phyllite. The black siltite and grey limestone clasts resemble those of the Harveys Ridge succession.

The matrix of the conglomerate consists mostly of glassy white clear, dark grey and minor blue quartz and minor feldspar and sericite. Where the matrix is calcite or is limy, as on parts of the Snowshoe Plateau (Holland, 1954, p. 22) and at Mount Burdett (Fig. 41), it weathers orange, brown or light grey. Quartzite clasts supported in the limy matrix tend to be more angular than those in either the quartzite or pelite matrix. Conglomerate with a black phyllite matrix is restricted to Luce, Little Snowshoe and French Snowshoe creeks. Quartzite clasts make up from about 10 to 20% of the conglomerate with a phyllite matrix. In some cases the rock could be referred to as a conglomeratic phyllite.

Clasts occupy from 40 to 60% of the conglomerate. They are subangular to subrounded with the mean grain size of pebbles. The quartzite clasts are normally larger than the pelite clasts. The conglomerate is massive; locally it has interbeds of black phyllite.

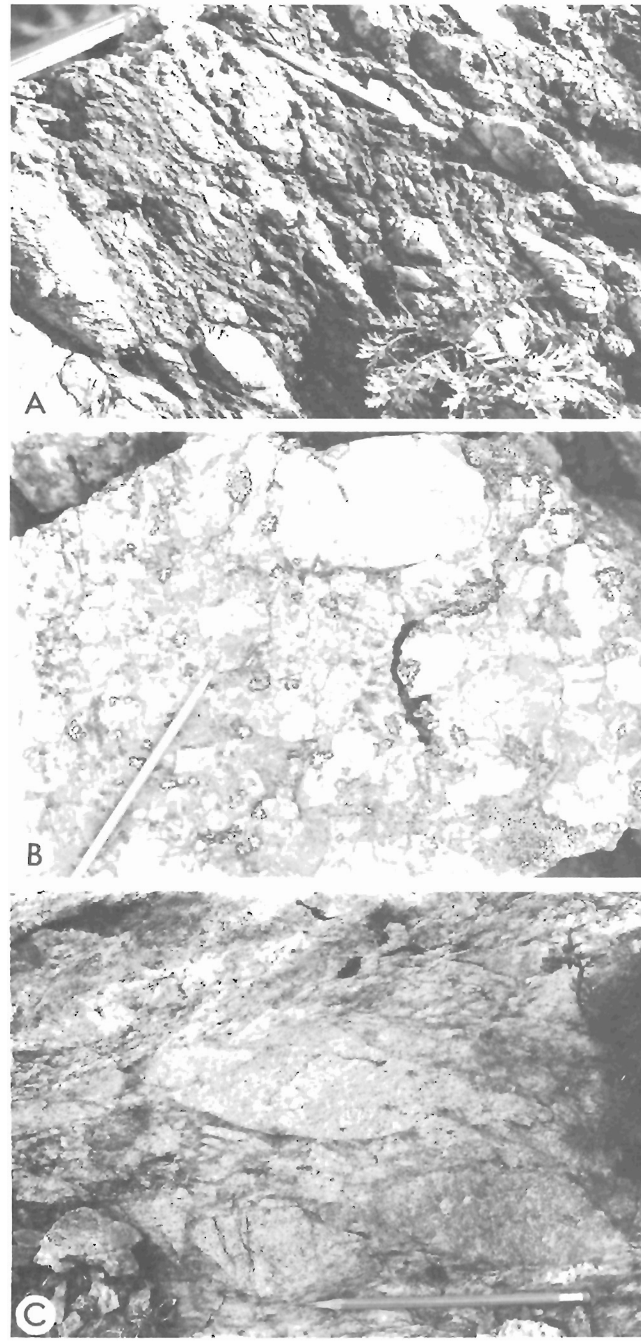


Figure 40. Conglomerate of the Agnes succession of the Snowshoe Group.
A) From Harveys Ridge (GSC 191034),
B) From southwest of Mount Agnes (GSC 191035) and
C) From near Mount Burdett. (GSC 191036)

The poor sorting, conglomerate grain size, massive character and lack of fluvial or glacial sedimentary features suggest that this conglomerate may be a debris flow deposit. However, except for the muddy conglomerate of the Snowshoe Plateau area little of the fine grained

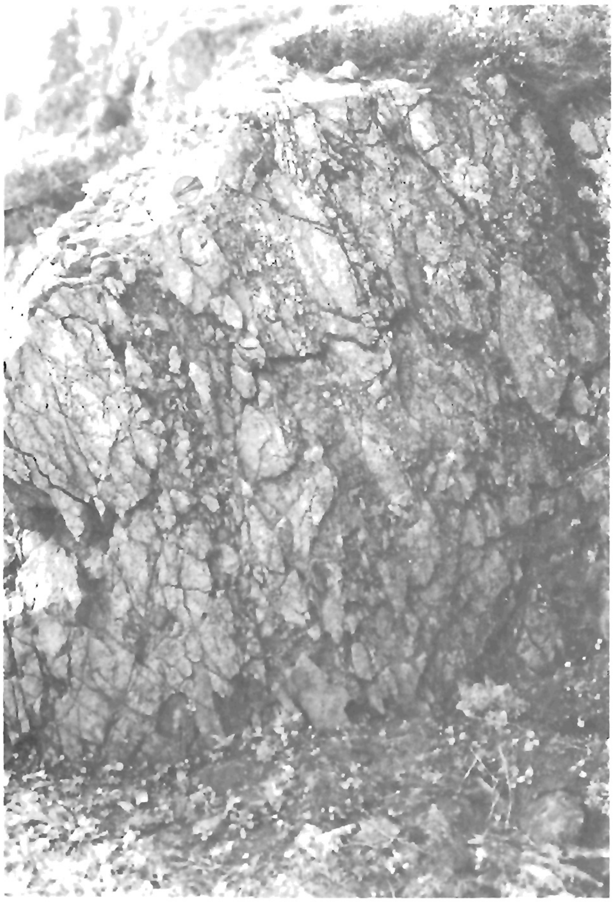


Figure 41. Calcareous conglomerate of the Agnes succession of the Snowshoe Group near Mount Burdett; looking southeast. (GSC 191037)

fraction is associated with this type of deposit. The disorganization of the conglomerate indicates a proximal debris flow deposit.

The Agnes conglomerate like the Goose Peak quartzite is not dated. Because it is a lateral equivalent of the Goose Peak it also is considered to be Paleozoic. Conglomerate of the Milford Group in Kootenay Arc is similar to the Agnes conglomerate, especially the exposure at Cornish Mountain.

Eaglesnest succession

The Eaglesnest succession is composed of various types of micaceous quartzite and phyllite. The rocks of this unit are much the same as those of the Tregillus, Ramos, Tom and Downey successions. Unlike these other units, however, it is usual for the Eaglesnest to have alternating olive and grey quartzite-phyllite sequences. The unit is defined primarily from its position above the Harveys Ridge and Goose Peak successions in the area of Dragon Mountain to Meridian Mountain.

The Eaglesnest succession underlies the area from Tregillus Creek southeast to Mount Burdett. These rocks were previously mapped as undifferentiated Snowshoe Formation by Campbell et al. (1973) and as the Dragon succession of unit 5 by Struik (1982a). No type section is assigned for this unit. Reference exposures exist at the head of Jack O' Clubs Creek and at Eaglenest and Dragon mountains. The sequence attains a minimum thickness of approximately 150 m.

The unit everywhere overlies the Harveys Ridge, Goose Peak or Agnes successions. The lower contact of the Eaglesnest succession near Mount Agnes consists of an interbedded sequence of rocks like those of Harveys Ridge, Agnes and Eaglesnest. A characteristic quartzite (purple-tinged grey, green mica bearing, feldspathic) of the Eaglesnest in the same area and at Aster Creek is gradational with underlying black siltite like that of the Harveys Ridge. Although the contact is not necessarily conformable, these relationships suggest that the sequence is in stratigraphic continuity.

Quartzite and phyllite of the Eaglesnest form olive and grey sequences, but between the head of Jack O' Clubs Creek and Aster Creek there is a purple-tinged variety of the grey quartzite. The quartzite of the Eaglesnest succession is poorly sorted, micaceous and ranges in grain size from fine to very coarse. It consists of clasts of 50 to 75% glassy clear, light grey to grey and minor blue quartz, up to 8% feldspar and minor zircon. Secondary minerals include the matrix of white mica and chlorite and accessory limonite, pyrite, ankerite, siderite and tourmaline. The purple-tinged variety of quartzite contains minor amounts of 1 to 3 mm long grains of green mica. The phyllite is either olive or grey and is much the same in appearance as other phyllites of the Snowshoe Group.

Bedding features of the quartzite-phyllite sequences include a wide range of thicknesses and more grading than in other units of the Snowshoe Group. Average bed thickness is 35 cm, with quartzite making up about 70% of the sequence. Grading occurs within and between beds (Fig. 42); a feature uncommon in other parts of the Snowshoe Group.

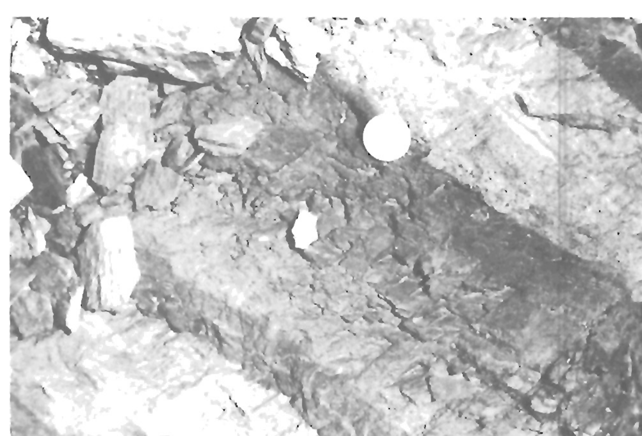


Figure 42. Normally graded quartzite to pelite of the Eaglesnest succession of the Snowshoe Group along Highway 26 in the pass at the head of Devils Canyon. (GSC 191038)

The sedimentary environment of the Eaglesnest is interpreted to be the same as the other quartzite-pelite sequences of the Snowshoe Group; debris flow transitional to turbidite.

The Eaglesnest succession is undated and thought to be Paleozoic only because it overlies the Harveys Ridge succession, assumed to be Paleozoic. Correlation of the Eaglesnest succession with units beyond the map area was not attempted, but in a general way it may relate to the grit-pelite sequences of the Eagle Bay Formation of south-central British Columbia.

Downey succession

The Downey succession is composed of micaceous quartzite, phyllite, marble, limestone, calcareous quartzite and tuff. The unit is characterized from others of the Snowshoe Group by its abundant marble and tuff. The quartzite commonly is brown weathering because of abundant porphyroblasts of ankerite and siderite.

The Downey succession underlies a narrow belt from Big Valley Creek in the north to the head of Cariboo Lake in the south. The rocks were previously mapped as Baker member of the Richfield Formation by Hanson (1935), Upper Snowshoe Formation by Holland (1954) and Sutherland Brown (1957), as undifferentiated Snowshoe Formation by Campbell (1961) and Campbell et al. (1973), and as the Downey Creek succession of unit 5 by Struik (1982a). No type section is assigned for this unit. Reference exposures are the roadcut along Downey Creek and those along Grouse and Cunningham creeks. The thickness of the succession is unknown, but is estimated to be more than 150 m.

The lower contact is poorly understood. At Mount Barker the Downey succession overlies the Harveys Ridge in apparent stratigraphic continuity. A similar relationship is implied on Harveys Ridge although the contact is not seen. The tuff and carbonate horizons on Mount Barker are near the basal contact, similar to the situation along Cunningham Creek. It is an apparent contradiction that both the Downey and Eaglesnest successions stratigraphically overlie the Harveys Ridge; possibly because they are laterally equivalent, or one or the other is unconformable on the Harveys Ridge.

Quartzite and phyllite of the Downey succession are grey, olive and brown, weathering grey and olive. Some areas of phyllite are green and brown where associated with possible tuffaceous rock. The quartzite-phyllite sequences may be altered to light grey and white in areas of hydrothermal activity such as the gold deposits near Wells and Barkerville. The quartzite is poorly sorted, micaceous and ranges in grain size from fine to coarse. It consists of 50 to 80% quartz (glassy clear, light grey to grey and minor blue), up to 10% feldspar and minor zircon. Secondary minerals include the matrix of white mica and chlorite and accessory limonite, pyrite, ankerite, siderite and tourmaline. Minor quantities of sphene, garnet, apatite and epidote appear in heavy mineral separates. The quartzite is locally calcareous. The phyllite consists of quartz silt and fine white mica and chlorite. Accessories commonly include ankerite, siderite, pyrite and limonite and locally chloritoid and biotite. Green chloritic phyllite contains abundant epidote and is considered to be at least in part of volcanic origin. In many places this phyllite is thinly interbedded with caramel-brown and purple phyllite in 10 to 30 m sequences commonly associated with marble.

The limestone and marble are grey and interlaminated grey and white. Crystal sizes range from 0.3 to 2 mm. The carbonate is everywhere strongly foliated apparently due to shear. It is commonly interlayered with green and grey phyllite and overlies, and is interbedded with, volcanic tuff. It is this interbedding of tuff and marble that distinguishes the Downey from other units of the Snowshoe Group. The Downey carbonate is the productive host for the replacement gold-bearing pyrite of the Barkerville gold belt.

The volcanic rock of the Downey is only poorly studied consisting primarily of tuff. Diorite sills were described from the Cariboo Gold Quartz Mine by Skerl (1948). Diorite is also found interlayered with the limestone marble along Cunningham Creek and at Mount Barker. Although labelled as diorite these rocks are more likely dioritic tuff.

Beds of the quartzite-phyllite sequences range from to 150 m thick. Graded bedding exists but is not common. Contacts between the quartzite and phyllite are generally sharp. The quartzite in places grades into marble and phyllite sequences. Contacts between marble and phyllite are mostly gradational and with the tuff are sharp. More detailed descriptions of the Downey succession in the vicinity of the Mosquito Creek Mine were given by Alldrick (1983).

The depositional environment of the Downey is considered to have been a marine shelf periodically inundated with clastic debris in the form of turbidites and debris flows. The carbonate and pelite are representative of low clastic input and the tuff of minor volcanic debris shed from a distant source.

Age and correlation. There are two occurrences of microscopic organic remains within the Downey; one in calcareous quartzite along Sugar Creek (Fig. 43), the other in dark grey sandy limestone on the 2200 logging road where it crosses Big Valley Creek. The fossils were determined by B.L. Mamet (University of Montreal) to be:

GSC Locality C-102829
(lat. 53°10'23", long. 121°41'57")
Silicified kamaenid algae or halysisid algae

GSC Locality C-102827
(lat. 53°10'37", long. 121°34'59")
Algae
Probable ostracod?
Echinoderm
Bryozoan

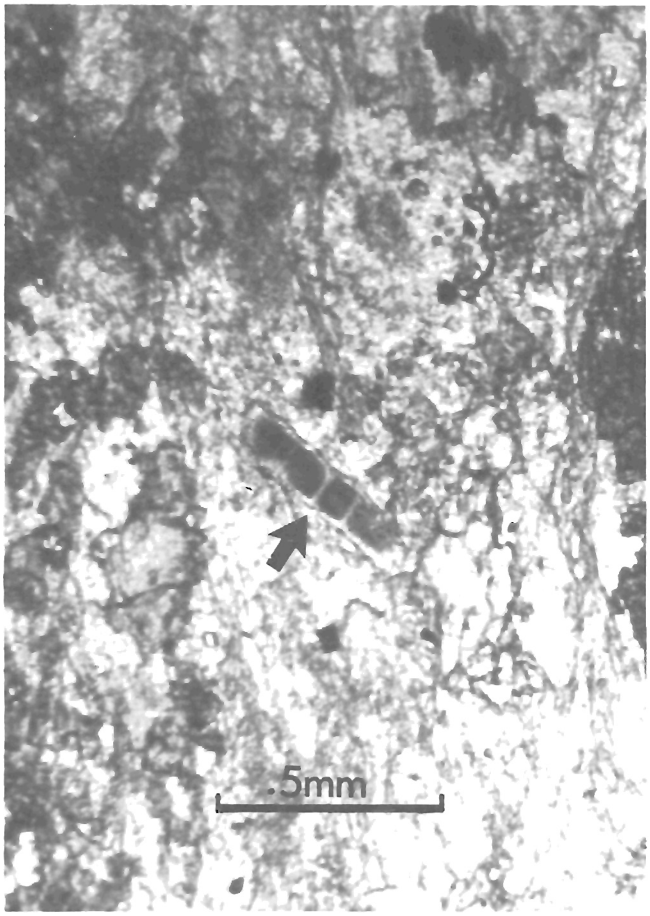


Figure 43. Photomicrograph of silicified kamaenid or halysisid algae from the Downey succession of the Snowshoe Group on Sugar Creek. (GSC 191039)

He suggested that the fossils exclude a Precambrian age and can be from any level of the Paleozoic, Late Cambrian and younger, though less likely Cambrian. The Downey succession is correlated with the part of the Eagle Bay Formation of southern British Columbia that includes the calcareous and volcanic rocks of Dixon Creek (unit 3 of Preto et al., 1980; Preto, 1981) and the tuff and agglomerate (unit 9 & 10 of Preto et al., 1980) that outcrop along the Adams Lake road. In that these rocks of the Eagle Bay Formation underlie the dated Mississippian rocks of the Eagle Bay Formation, the Downey succession is probably lower Paleozoic.

Bralco limestone

The Bralco limestone consists mainly of limestone, marble and minor phyllite. The limestone is similar to that of the Mural and Cunningham formations of the Cariboo Terrane. The marble is similar to that of the Downey succession but is purer carbonate. The only distinguishing characteristic of the Bralco limestone is its stratigraphic position above the Downey succession.

The Bralco limestone is confined to the belt paralleling the Pleasant Valley Thrust from Summit Creek southeastward to Cariboo Lake and beyond the map area to the north arm of Quesnel Lake. The limestone was previously mapped as part of the Cariboo Schists by Bowman (1889), as part of the Barkerville Formation by Johnston and Uglow (1926) and Lang (1938), as the Midas Formation by Holland (1954), as the Isaac Formation by Campbell et al. (1973) and as the Cunningham or Mural Formation by Struik (1981, 1982b). There is no type section for the unit; a reference exposure is at the headwaters of Penny Creek. Thickness of the carbonate is highly variable, and is less than 100 m.

The Bralco limestone overlies the black siltite and phyllite of the Downey succession along the length of its exposure. Similar carbonate appears in outcrop areas of the Downey succession in the northwestern part of the map area.

The light to dark grey limestone and marble consists of 0.2 to 2 mm calcite crystals and near Roundtop Mountain contains up to 10% algal pellets up to 2 mm across. The noncarbonate component is minor consisting mostly of fine grained quartz and lesser heavy minerals. The carbonate is commonly laminated on a 1 to 3 cm scale.

In the vicinity of Penny Creek the Bralco limestone appears to be thrust faulted with brown-grey phyllite, possibly an alteration of the Downey phyllite. Although this phyllite is included in the Downey succession it may be part of the Bralco limestone sequence. Part of the phyllite resembles that of the Midas Formation on Roundtop Mountain.

Age and correlation. The age of the Bralco limestone is thought to be Paleozoic because pelletoidal limestone (Fig. 44) found near Roundtop Mountain (GSC Locality C-102828 Lat. 52°55'14" Long. 121°20'3") contains echinoderm fragments, as identified by B.L. Mamet. It may correlate with the Tsinikin Limestone of the Eagle Bay Formation.

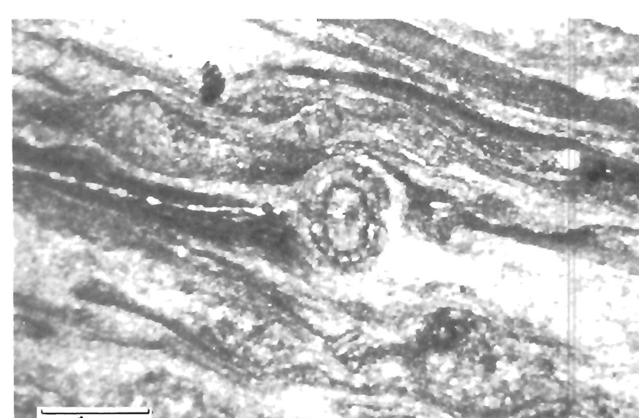


Figure 44. Photomicrograph of echinoid fragments within the Bralco marble of the Snowshoe Group near Roundtop Mountain.

Hardscrabble Mountain succession

The Hardscrabble Mountain succession consists of siltite, phyllite and muddy conglomerate. The rocks, except for the muddy conglomerate, are similar to those in other parts of the Snowshoe Group. The conglomerate resembles part of the Guyet Formation conglomerate of the Cariboo Terrane east of the Pleasant Valley Thrust but is more feldspathic and has blue quartz clasts.

The Hardscrabble Mountain succession underlies the area south of Big Valley Creek in a narrow belt southeastward to Cariboo Lake. It follows the Downey succession and lies along the Pleasant Valley Thrust. It was previously mapped as Rainbow Member of the Richfield Formation by Hanson (1935), as Midas Formation by Holland (1954) and Sutherland Brown (1957), as Isaac Formation by Campbell et al. (1973), and as unit 4 by Struik (1982a). The unit is not formal and is not assigned a type section. Reference exposures exist at the head of Hardscrabble Creek and on Hardscrabble Mountain. Reasonably good exposures can be found on Island Mountain. The thickness is estimated to be greater than 100 m near Hardscrabble Creek but thinner along the eastern margin of the Downey succession.

The Hardscrabble Mountain apparently overlies the Downey and Bralco successions stratigraphically. To the north at Cooper Creek it is in sharp contact with the underlying Downey, but southeastward to Island Mountain it is locally gradational. Fold and cleavage relationships in the area of Island and Barkerville mountains suggest that the Hardscrabble Mountain stratigraphically overlies the Downey. Alldrick (1983) reported graded beds from the Downey rocks in the Mosquito Creek Mine which indicate the same relationship. Quartzite pebble to cobble conglomerate on Cornish Mountain, however, is very similar to that of the Agnes conglomerate and if they are the same, the sequence from Cornish Mountain to Hardscrabble Mountain would be overturned and, 1) the Hardscrabble Mountain would stratigraphically underlie the Downey, and 2) the Hardscrabble Mountain would be equivalent to the Harveys Ridge succession. This interpretation was proposed by Struik (1982a, b). It requires overturned nappe structures greater than 6 km across the structural trend, but no evidence for such a fold exists southeastward along the belt.

The black siltite and phyllite of the Hardscrabble Mountain is identical in hand specimen to that of the Harveys Ridge. The 2 to 20 cm interbeds commonly contain 1 to 3 mm laminae of white quartz parallel to the bedding. The rocks are locally altered to light grey siltite and sericitic phyllite in areas affected by hydrothermal activity, as in the gold mining area of Island Mountain and Lowhee Creek. The alteration is both parallel to and crosscuts bedding, completely or as a mottling. Pyrite, siderite and ankerite are common secondary minerals of the phyllite and less so of the siltite.

The black muddy conglomerate consists of clasts of 40 to 65% quartz, 1 to 6% feldspar, up to 30% rock fragments of black phyllite, grey chert, and less of cataclastic quartz. The quartz is glassy clear, black and minor blue. The feldspar is altered to clays on the average of 20%. Secondary muscovite, chlorite and albite form the matrix. Siderite, rutile and tourmaline are secondary accessories. The conglomerate on the hill north of Mount Wiley contains much more chert clasts than on Hardscrabble Mountain and resembles the Guyet Formation conglomerate east of the Pleasant Valley Thrust.

Subordinate rock types include limestone, tuff and barite. The limestone is dark grey or light grey where it is mainly finely crystalline marble. It is found at Sheperd Creek where it may be equivalent to the Bralco limestone. The tuff resembles that of the Downey succession and is confined to Sheperd and Pinus creeks. The bedded barite is located southwest of Roundtop Mountain. It was pointed out by Geoff Hodgson, then of RioCanex. It is 50 cm thick and its lateral exposure is confined to an exploration trench.

The environment of deposition of the Hardscrabble Mountain is postulated to have been a distal turbidite fan with periodic influxes of muddy conglomerate debris flow. The minor limestone is confined to the upper and lower portions of the unit and reflects a change of source material from that required for the grit-pelite sequences of the Downey succession.

Age and correlation. The age of the Hardscrabble Mountain succession is assumed to be Paleozoic solely because it overlies the Downey and Bralco successions and because it appears to underlie the Lower Permian Sugar limestone. Some of the succession resembles the Mississippian unit EBP of the Eagle Bay Formation of Schiarizza and Preto (1984).

Tom succession

The Tom succession consists of micaceous quartzite, phyllite, sheared quartzite and schist. There is nothing sedimentalogically unique about the Tom succession within the Snowshoe Group. It is much like the Tregillus, Ramos, and Eaglesnest grit-pelite sequences. It is distinguished by its position above rocks correlated with the Downey succession. The position is structural and the Tom probably correlates with another unit of the Snowshoe Group.

The Tom succession is confined to the area between the Willow River and Big Valley Creek. A klippe of the unit rests on Island Mountain. The rocks were previously mapped as undifferentiated Snowshoe Formation by Campbell et al. (1973) and as the Tom succession of unit 5 by Struik (1982a). There is no type area of the unit, which is generally poorly exposed. Its thickness is unknown.

As the Tom is everywhere thrust onto underlying units, stratigraphic relationships with other units are unknown.

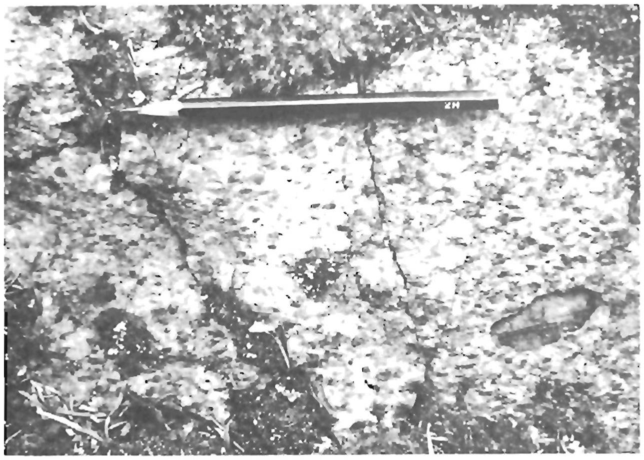


Figure 45. Grit of the Tom succession of the Snowshoe Group on Island Mountain where it overlies the Island Mountain amphibolite. (GSC 191040)

Olive and grey micaceous quartzite consists of poorly sorted clasts of quartz (glassy clear, white and minor blue) and minor feldspar (Fig. 45). Secondary muscovite, albite and minor chlorite form the matrix. Zircon and secondary tourmaline compose most of the accessories.

Sheared quartzite contains a similar mineralogy as the micaceous quartzite, except that it has less matrix mica and contains 1 to 4% garnet. The quartz grains are pencil forms with ratios of long to short axes of 4 or 5 to 1. The garnet forms an interstitial growth with fractured mylonitic quartz and occurs as fragments of distended grains separated by sericite. The sheared quartzite is interlayered with muscovite schist and is mainly near the contact with the underlying Island Mountain amphibolite.

The thrust sheet containing the Tom succession has no known root and must now represent a klippe, possibly of a large overturned fold of Tom succession and Island Mountain amphibolite. The distended and interstitial garnets together in the sheared quartzite sequence suggest a history of garnet growth outlasting shearing. The age and correlation of the Tom succession is unknown.

Sugar limestone

The Sugar limestone consists entirely of crinoidal limestone. It is unique in the Barkerville Terrane as the only recognizable crinoidal limestone.

The limestone is exposed in one locality at the head of Sugar Creek, after which it is informally named. It was first described by Struik (1980) as part of the Greenberry limestone of the Guyet Formation. Later (1982a, b) I distinguished it from the Greenberry limestone which is restricted to the Mississippian. The thickness is less than 15 m.

The Sugar limestone rests sharply on black siltite and phyllite of the Hardscrabble succession on the southwest slope of Hardscrabble Mountain. There is no indication that the contact is a fault.

The grey limestone consists of some 30% of 0.5 to 2 mm size crinoid stem fragments, less than 6% silt size quartz and feldspar grains (some of which may be secondary), and a finely crystalline dusty calcite matrix. The rock is disrupted by approximately 5% stylolites. Locally the limestone is silicified to light grey chert in 2 to 6 cm bands, possibly defined by bedding.

The disorganized bioclasts suggest reworking prior to deposition in the limestone. The reworking may be a result of wave action or sediment flow in a subaqueous environment. There is no mixing of shallow and deep water fauna within the limestone to indicate the transport of shallow water debris into deeper water.

Age and correlation. The Sugar limestone is Lower Permian as determined from extracted conodonts (Orchard and Struik, 1985; *see* Appendix A). The existence of this Lower Permian limestone is critical in the reconstruction of the geological history of the area. It implies that within the Barkerville Terrane there may be an unconformity of Paleozoic age. Because the stratigraphic sequence below the unconformity is completely different from that underlying the Permian of the Cariboo Terrane it implies that the Paleozoic stratigraphy and/or structural history of the two terranes is different. Either of these cases requires that the terranes be sufficiently separated in space to allow for the different geological histories.

Several units of the Barkerville Terrane are correlated with units of the Eagle Bay Formation. The Eagle Bay Formation is part of the Kootenay Terrane, which has an unconformity between lower Paleozoic clastic rock and overlying Mississippian to Permian(?) rock. The unconformity below the Lower Permian Sugar limestone may be the same.

Island Mountain amphibolite

The Island Mountain amphibolite consists mainly of amphibolite with minor amounts of tuff(?) and cataclastic quartzite. This amphibolite is distinguished from that of the Ramos succession by its lack of associated carbonate and coarsely crystalline amphibole. The finely crystalline amphibolite of these units appears much the same in hand specimen.

The amphibolite forms the peak of Island Mountain and underlies the drainage area of Tom Creek. It was previously mapped as a hornblende gneiss of the Snowshoe Formation by Sutherland Brown (1957), as undifferentiated Snowshoe Formation by Campbell et al. (1973) and as unit 5e by Struik (1982a). Good reference areas include Island Mountain and Tom Creek. Thickness of the amphibolite is less than 150 m.

The Island Mountain amphibolite is believed to be thrust onto grit, pelite and limestone (assigned to the Downey succession) because of the stratigraphic omis-

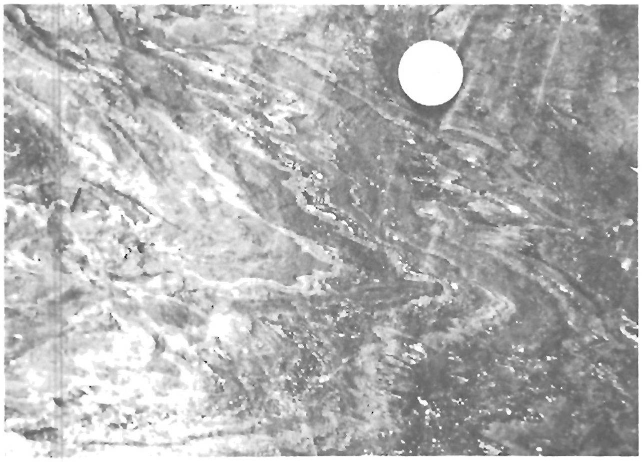


Figure 46. Deformed Island Mountain amphibolite from Tom Creek composed mainly of actinolite, hornblende and plagioclase (white patches). (GSC 191041)

sion of the Hardscrabble Mountain and Bralco successions. It is included in the Showshoe Group because it structurally underlies the Tom succession.

The amphibolite (Fig. 46) consists of 50 to 85% actinolite and/or hornblende, 0 to 20% plagioclase, 3 to 15% epidote, 2 to 8% sphene, 2 to 6% chlorite, 0 to 4% muscovite, 0 to 3% biotite and up to 10% quartz. The amphibole is prismatic and more or less aligned within foliation surfaces. The amphibole has formed in two generations, the first consists of poikiloblastic blocky hornblende and the latter of prismatic actinolite. Plagioclase is wholly restricted to secondary growths of varying composition, mainly sodium-rich. Late feldspar and quartz are intimately intergrown in a myrmekitic style. Needles and blades of epidote occupy the matrix as trains, and irregular crystals as random patches. Sphene is randomly dispersed throughout the matrix as 0.05 to 0.2 mm crystals. Chlorite, muscovite and biotite are aligned in the foliation surface defined by the amphibole. Biotite is locally altered to chlorite. The 0.05 to 0.5 mm quartz crystals form late crosscutting stringers and fine matrix intergrowths.

Thinly laminated cherty tuff(?) occurs in the sequence on Island Mountain. In the same sequence are the sheared quartzites which resemble those from the Tom succession near Tom Creek. The tuff may actually be a cherty mylonite as it consists mainly of microcrystalline quartz. The tuff and mylonite are less than 1 m thick horizons in an otherwise amphibolitic sequence.

The Island Mountain amphibolite resembles the Crooked Amphibolite (Slide Mountain Terrane?) that occurs along the western margin of the Barkerville Terrane. If these amphibolites correlate then the Island Mountain amphibolite outlines the lower limb of a large easterly-verging recumbent nappe. The Island Mountain amphibolite would have been thrust eastward(?) onto the grit-pelite sequence of the Tom succession as recorded by the sheared rock of the Tom. The structural package of amphibolite and Tom succession would then have been involved in the east-verging recumbent fold thrust onto the more easterly Snowshoe Group. In such a scenario the Antler Formation (Slide Mountain Terrane) must be thrust later onto the structural package that includes the Island Mountain nappe. This is very difficult to reconcile with the Crooked Amphibolite unit as the root zone for the Antler Formation thrust sheet.

Intrusive rocks

The intrusive rocks of the Barkerville Terrane consist mainly of diorite with subordinate rhyolite and rhyodacite.

Diorite

The diorite forms sills throughout the terrane and is most abundant along the western margin. Thickness of the sills range from 0.4 to 30 m. All diorites are included in this description even though they may be of different generations.

Primary mineralogy of the diorite is dominated by augite and plagioclase with less quartz, opaques and sphene. The augite takes up 20 to 70%, consists of 0.5 to 4 mm size crystals and is variably altered to hornblende. Plagioclase occupies from 3 to 45% of the rock, the percentage variation due mostly to the intensity of alteration. Secondary minerals include hornblende, epidote, chlorite, albite, muscovite and calcite. Hornblende, as coarse grains, replaces the augite. Other secondary minerals are finely crystalline and form the matrix.

As it is locally isoclinally folded, the diorite is mostly (if not totally) older than Jurassic tectonism. The sills are similar to the diorite of the Slide Mountain Terrane chemically (Table 15), mineralogically and texturally.

Rhyolite and rhyodacite dykes and sills

The rhyolite and rhyodacite occur sporadically throughout the terrane as dykes and sills. They are generally from 1 to 10 m thick. There are at least two generations of felsic intrusion, one prefolding and the other postfolding.

The older intrusions include the Proserpine dykes as described by Holland (1954, p. 23) and Sutherland Brown (1957, p. 41, 42). Proserpine dyke samples were examined from the old mining road crossing Conklin Gulch. They are highly ankeritized brown weathering felsic rock. The samples collected were too badly altered to make reasonable feldspar determinations. The potassium feldspar content reported by Sutherland Brown (1957, p. 41) is sufficient to classify these as granitic rocks.

Sutherland Brown (1957) reported that the Proserpine dykes were folded and schistose, suggesting intrusion prior to the last phase of isoclinal folding. A light brown weathering quartz porphyry dyke intruding Hardscrabble

Table 15. Major element analysis of dioritic rocks from Antler Formation and Snowshoe Group

ANTLER FORMATION

SAMPLE	0-48	0-134	0-135	0-137	0-144A	0-155
SiO_2	45.8	51.3	50.2	48.8	54.6	50.7
TiO_2	.52	1.68	1.70	.62	1.08	1.03
Al_2O_3	9.9	15.3	15.4	15.6	15.3	15.6
Cr_2O_3						
Fe_2O_3	1.6	7.4	5.1	4.1	4.9	2.3
FeO	9.6	4.6	5.0	3.5	4.2	6.9
MnO	.28	.19	.20	.16	.16	.18
MgO	14.3	4.63	5.98	7.85	5.05	7.14
CaO	11.3	7.24	8.98	12.5	7.42	8.59
Na_2O	.6	5.0	2.8	2.4	4.2	3.2
K_2O	.06	.43	.24	.57	.41	1.00
H_2O T	4.7	2.3	3.4	2.9	2.7	2.9
CO_2	.3	.6	.2	.7	.1	.5
P_2O_5	.03	.20	.22	.06	.14	.10
S	0.00	0.00	0.00	0.00	0.00	0.00
Rb	0.000	.005	.004	.006	.003	.005
Zn	.031	.008	.008	.008	.007	.009
Zr	.001	.009	.011	.004	.005	.007
Total	99.4	100.9	99.5	100.0	100.3	100.2

Sample numbers are prefixed with SCB8.

Location of samples		Latitude	Longitude
0-48	Yuzkli Lake	53°12'05''	121°46'42''
0-134	Alces Creek	53°11'43''	121°50'32''
0-135	Alces Creek	53°11'48''	121°50'32''
0-137	Alces Creek	53°11'56''	121°50'32''
0-144a	Yuzkli Creek	53°12'45''	121°51'35''
0-155	Yuzkli Creek	53°12'48''	121°49'07''
0-101	Meadows Creek	53°08'45''	121°45'38''
0-111	Meadows Creek	53°09'28''	121°47'03''
0-130	Alces Creek	53°11'01''	121°48'59''
1-100	Mount Campbell	52°59'41''	121°46'34''
1-101	Font Creek	52°59'41''	121°45'00''
1-278	Cariboo Mountain	52°52'30''	121°34'25''
1-296	Cariboo Mountain	52°52'32''	121°32'17''
1-1027	Cariboo Mountain	52°51'56''	121°37'20''
1-1036	McMartin Creek	52°52'23''	121°30'57''
1-1767	Keithley Creek	52°47'03''	121°27'50''

DIORITIC DYKES AND SILLS INTRUDING SNOWSHOE GROUP

SAMPLE	0-101	0-111	0-130	1-100	1-101	1-278	1-296	1-1027	1-1036	1-1767
SiO_2	51.3	50.3	48.9	50.0	47.7	48.6	49.7	51.4	48.6	53.0
TiO_2	1.04	.92	.73	1.19	1.84	1.13	.91	2.10	1.32	1.56
Al_2O_3	14.3	12.9	16.9	15.3	14.4	15.6	13.4	17.3	14.5	14.9
Cr_2O_3										
Fe_2O_3	2.2	2.2	1.9	2.2	2.2	2.1	1.9	1.7	2.7	2.4
FeO	7.0	6.9	6.3	6.4	7.0	6.9	7.0	8.2	7.1	8.5
MnO	.17	.20	.17	.18	.17	.18	.17	.23	.19	.20
MgO	7.48	9.42	7.33	6.73	8.28	7.57	10.3	3.29	5.00	4.83
CaO	10.3	11.7	10.1	11.3	11.3	10.6	11.2	6.65	8.12	8.42
Na_2O	2.5	1.9	2.9	2.4	2.7	2.4	2.1	5.1	2.9	3.2
K_2O	.12	.14	1.19	.56	.51	.88	.53	.50	.09	.55
H_2O T	2.8	3.1	2.6	2.6	3.0	2.6	3.0	1.9	4.4	2.4
CO_2	.3	.6	.5	.3	1.1	.3	.2	.2	56	.1
P_2O_5	.11	.09	.08	.14	.23	.11	.10	.63	.14	.16
S	0.00	0.00	0.00	0.00	0.00	0.00	.09	0.00	.03	.12
Rb	.004	.002	.005	.004	.003	.005	.003	.006	.002	.005
Zn	.010	.009	.008	.007	.010	.008	.008	.012	.010	.009
Zr	.006	.004	.006	.005	.009	.005	.007	.026	.007	.009
Total	99.7	100.5	99.6	99.3	100.6	99.1	100.7	99.4	100.7	100.4

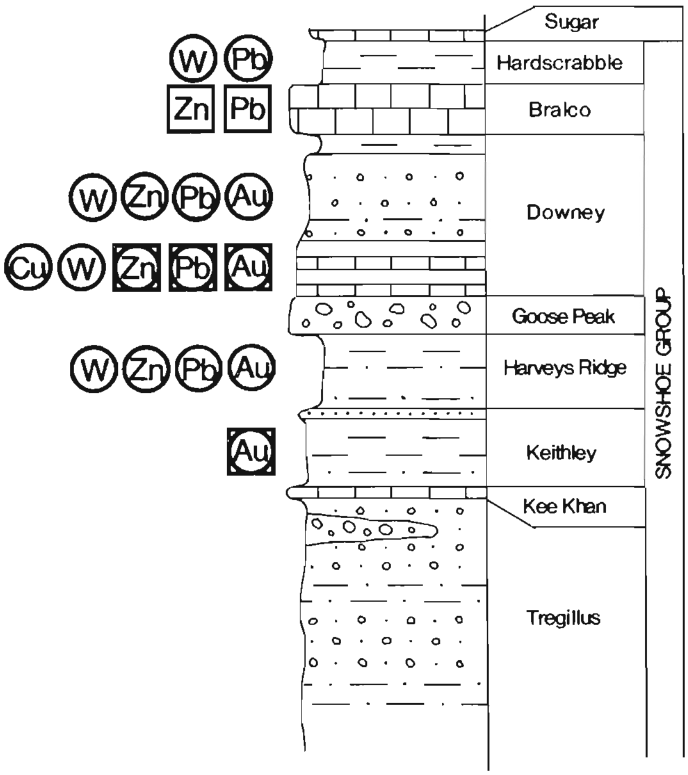
Mountain black phyllite along Antler Creek may be part of the same generation as the Proserpine dykes.

Holland (1954, p. 23) described biscuit-coloured rhyolite porphyry from the head of Luce and Little Snowshoe creeks, which intruded parallel to the country rock foliation but is not itself foliated or altered.

Economic geology

Barkerville Terrane hosts the principal gold occurrences in the area; Mosquito Creek, Island Mountain, Cariboo Gold Quartz, Cariboo Hudson and the Snowshoe and Midas veins. Deposits of less economic importance include those of silver, tungsten, lead, zinc and copper. No attempt is made here to fully describe the economic mineral occurrences within the gold belt as many of them have been well documented by Bowman, (1895), Johnston and Uglow (1926), Holland, (1954), Sutherland Brown (1957) and Alldrick (1983). Recent exploration and mining has concentrated in the areas of established mineralization as described by these authors. The location of gold deposits correlates with elements of 1) the stratigraphy, 2) the structure, and 3) the metamorphism. It is suggested that the interrelationship of these three factors has contributed to the localization of the gold deposits. Stratigraphy, structure, and metamorphic controls of mineralization have previously been noted by Hanson (1935), Benedict (1945), Holland (1954), Sutherland Brown (1957), and Alldrick (1983). The following discussion emphasizes the interrelationship of the geological components and relates the mineralization to the structural synthesis presented in this paper.

Stratigraphic controls: Both lode and placer deposits of gold are associated primarily with the Downey succession. Lode gold deposits are almost entirely confined to the Paleozoic part of the Snowshoe Group (Fig. 47). The



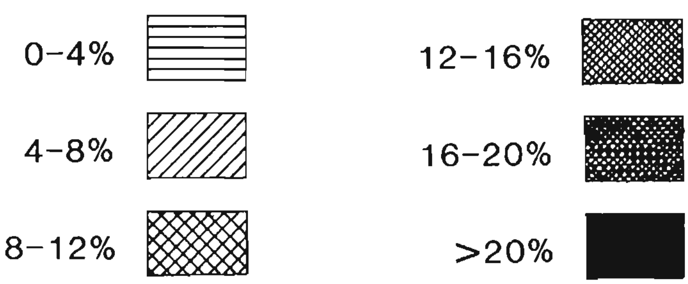


Figure 47. Mineral occurrences within stratigraphic units of Barkerville Terrane. The squares represent replacement deposits and the circles vein deposits of the elements: Au-gold, Pb-lead, Zn-zinc, W-tungsten, and Cu-copper. Silver coexists with gold and lead. The symbols denote existence of a mineral concentration only and have no connotation of deposit size or relative abundance.

most productive lode gold mines of the area (Island Mountain, Cariboo Gold Quartz, and Mosquito Creek; *see* Sutherland Brown, 1957 and Alldrick, 1983) are hosted by the limestone and metabasalt bearing components of the Downey succession. Productive placer gold mines of the area including Williams, Lowhee, Mosquito, Grouse, upper Cunningham and Harvey creeks and Stout, Conklin and Peter gulches (Fig. 48; *see* Bowman, 1895; Johnston and Uglow, 1926; Holland, 1954; and Sutherland Brown, 1957) overlie the same limestone and metabasalt bearing sequence of the Downey succession. Other productive placer deposits (Lightning, Little Snowshoe, French Snowshoe, and Keithley creeks; *see* Bowman, 1895; Lang, 1939; Holland, 1948, 1954) are near areas with high densities of gold bearing quartz veins, mainly in the Paleozoic(?) Harveys Ridge succession of the Snowshoe Group (Fig. 47).

Structural controls: Lode gold concentrations, as auriferous replacement pyrite in limestone, are located in the hinge zones and less commonly along the limbs of regional and minor folds (Benedict, 1945; Sutherland Brown, 1957; Alldrick, 1983). Gold-bearing quartz veins crosscut, and therefore are assumed to be younger than, the regional folds. Examples are common of vein and replacement gold mineralization of the same age and of replacements located in the paths of quartz veins (Benedict, 1945; Sutherland Brown, 1957; and Alldrick, 1983). The auriferous replacement pyrite is therefore considered to have formed after the regional folds which control the distribution of replacement ore. Orientations of veins are in part controlled by the regional fault and fracture pattern and both are consistent with a single regional stress field (Holland, 1954; Sutherland Brown, 1957).

Metamorphic controls: Lode gold concentrations are confined to rocks in the chlorite grade of metamorphism (Fig. 48). It is suggested that the gold mineralization is confined to lower metamorphic grade rocks because the gold precipitated in the cold regime of a circulating meteoric hydrothermal system driven by the heat differential caused by the metamorphic high heat anomally; ie. that the gold precipitation and metamorphism are coeval.

The metamorphism transects stratigraphy such that a single rock unit can vary from chlorite to kyanite grade. If gold concentration was directly related only to stratigraphy there would be gold deposits in the higher grade rocks. The Downey succession, which hosts economic concentrations of lode gold, passes southeast into progressively higher metamorphic grades and at Three Ladies Mountain is involved in kyanite grade. At biotite grade and higher the Downey succession, like the rest of the Snowshoe Group, has no recorded gold showings.

The age of the metamorphism and mineralization are insufficiently understood to test the hypothesis that they are coeval. Andrew et al. (1983) have determined a 179 ± 8 Ma (K-Ar) age for a phyllite near Island Mountain, which is probably the age of metamorphism associated with regional ductile folding. They also reported 141 ± 5 Ma (K-Ar) for sericite derived from a quartz-barite vein in the vicinity of the dated phyllite, confirming the idea that the veining is younger than the ductile fold system. Sedimentary zircons from the Downey micaceous quartzite yield 114 ± 10 Ma (U-Pb) as a lower intercept (P. van der Heyden, written communication, 1983), which is interpreted as a metamorphic age. With this paucity of data, each date could represent a peak in metamorphic heat flow or any other combination of circumstances affecting daughter isotope entrapment. With the information at hand, a regional metamorphic heat source

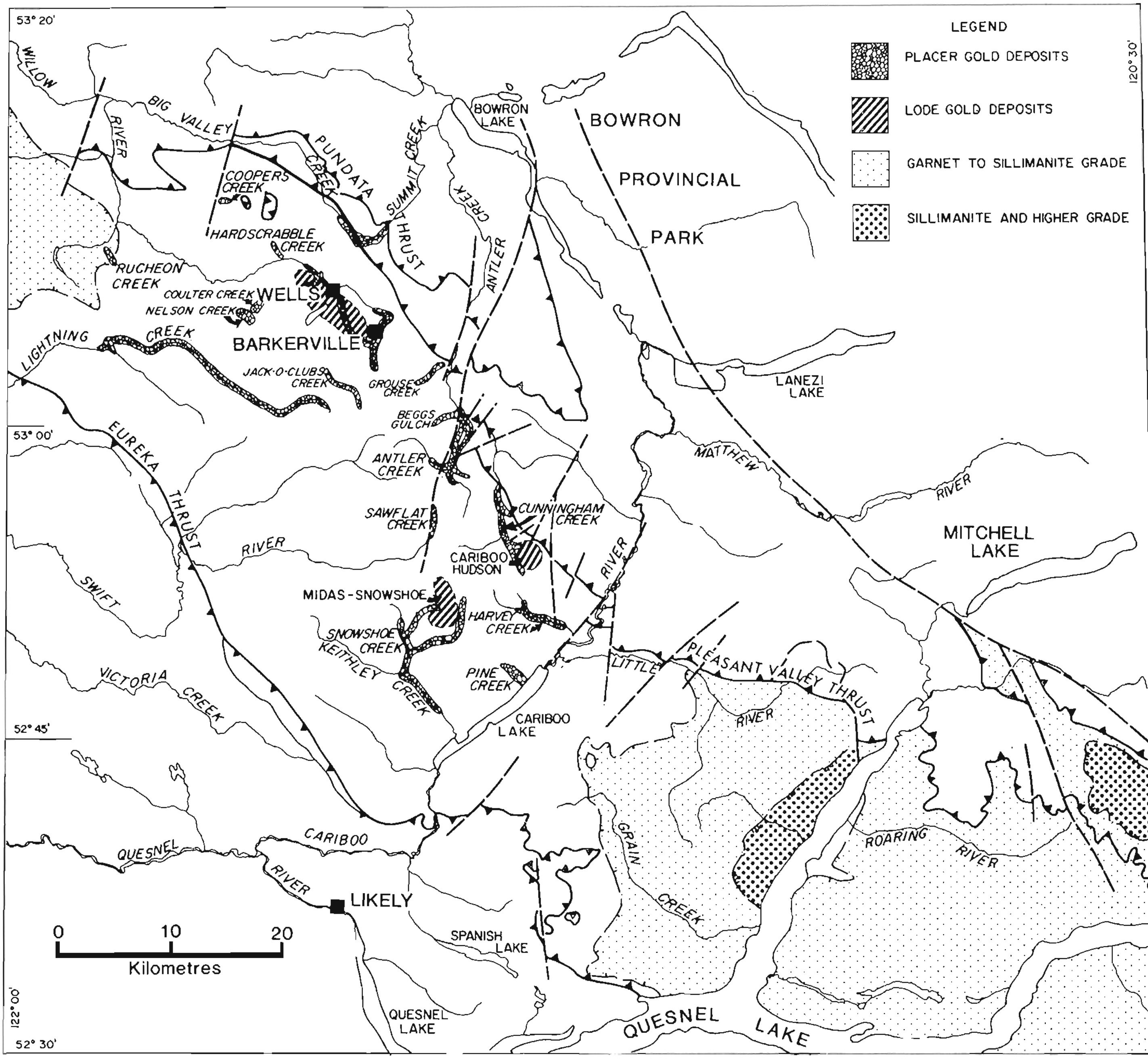


Figure 48. Distribution of the most economically important placer and lode gold deposits, throughout the Cariboo gold belt and southeast to Quesnel Lake. Areas of chlorite (unpatterned), garnet to sillimanite, and sillimanite and greater grades of metamorphism are indicated to show that the gold deposits are confined to the low grade rocks.

could be the same age as a hydrothermal system that pumped gold into the cooler parts of Barkerville Terrane.

In summary, the model of gold precipitation within Barkerville Terrane involves the coincidence of 1) structural traps formed by the thickening of favourable host rock (limestone) in the cores of folds, 2) the host rock overlying or incorporated in rocks able to yield gold to a hydrothermal system, 3) the host rock overlying a high heat anomaly (recorded in the metamorphism), 4) a fracture system above the heat anomaly to act as conduits (now quartz veins) and 5) the high structural position (low temperature) of the host rock relative to a metamorphic temperature gradient. Gold precipitation would occur in the cooler parts of the system both spatially and in time. During peak temperatures of metamorphism the cool parts of the system would be near to the paleosurface and during the waning stages of metamorphism the temperature would drop and the precipitation could proceed at greater depths than previously.

Galena, sphalerite and barite are hosted by carbonates and clastic rocks ascribed to the Downey succession and black siliceous slate, phyllite and argillite ascribed to the Hardscrabble Mountain succession along Cunningham Creek and near Roundtop Mountain (Longe and Hodgson, 1978; 52°55'N, 121°22'W). The barite appears to be stratiform and the galena contains impurities of silver. Some quartz veins in the units contain gold and scheelite. Gossans with anomalous concentrations of zinc, silver and cadmium overlie black fine grained clastic rocks of the Harveys Ridge succession on Harvey's Ridge (Fraser, 1978; 52°51.2'2N, 121°22'W).

Scheelite is found in quartz veins (Mulligan, 1983) in host rocks of Hardscrabble Mountain and Downey succession.

Structure and metamorphism

The Barkerville Terrane consists of essentially one structural package; defined as a deformed sequence of rock separated from others by an angular unconformity. The Snowshoe Group, as the main structural package, may be overlain unconformably by a younger one, the Permian Sugar limestone. Within the Snowshoe Group there may be two structural packages divided by a proposed unconformity beneath the Downey succession. In spite of this possibility the Snowshoe Group is treated as one structural package.

Structures of the Snowshoe Group are divided into 3 categories; from oldest to youngest they are 1) shear, 2) ductile shortening, and 3) brittle shortening and extension. The categories highlight aspects of a spectrum of strain conditions. The spectrum began with shear in the latest Triassic(?) and ended with brittle extension due to uplift in the Eocene. The metamorphism is included in the discussion of the structures because it controls to a large degree the strain parameters. The strain environment is emphasized over the superposition of structural elements. As a consequence there will be little importance attached to designating structures as S_1 or S_2 etc., although for comparison purposes this is done in its simplest form in Table 16.

Shear

This category of structures, formed by shear, consists of mylonitization (as defined by Wise et al., 1984), bedding-cleavage, and isoclinal recumbent, commonly rootless folds. It includes the oldest structures.

Mylonitization

Mylonite of the Snowshoe Group is concentrated in two areas; along the western margin of the Barkerville Terrane, and near the Island Mountain amphibolite.

Table 16. The structures of the Barkerville Terrane organized into a deformation series (D) of surfaces (S), and folds (F).

Event	Structure	Metamorphic mineral
D_5	Extension faults which trend NNE and N, commonly vein filled, generally with oblique displacement (some of these faults may be D_4).	Quartz, pyrite, gold, ankerite, scheelite, galena (some evidence for earlier formation), silver
D_4 (F_4)	Open folds, warps, kinks, and crenulations (F_4) of variable orientation; reverse strike faults, steeply dipping.	Chlorite (retrograde), tourmaline (could also have grown post chlorite). Period of quartz, pyrite, gold, ankerite, scheelite, galena and silver mineralization may have started at this time
D_3 (S_3, F_3)	Spaced crenulation axial surface cleavage (S_3); asymmetric flow to concentric folds (F_3) plunge to NW but may have variable orientation where they are E directed	Quartz, muscovite, biotite? chloritoid
D_2 (S_2, F_2)	Ductile shear Pervasive axial surface cleavage (S_2) asymmetric flow folds (F_2) plunge to the NW, but may have variable orientation where they are E directed	Muscovite, biotite, garnet, quartz
D_1 (S_1, F_1)	Ductile shear, bedding cleavage (S_1); isoclinal recumbent rootless folds (F_1) that plunge NW	Muscovite, quartz, albite?

Sutherland Brown (1957) described flaser quartzite from the Snowshoe Formation where it is presently mapped as the Downey succession.

The most intensely mylonitic rocks, including some ultramylonite, are at the contact of the Crooked and Island Mountain amphibolites. This proximity is assumed to imply that the amphibolites are thrust onto the Snowshoe Group. The sense of shear has not been determined for these rocks, however, Rees and Ferri (1983) suggested that the Crooked Amphibolite is thrust to the east over the Snowshoe Group, as determined from features within the Quesnel Lake orthogneiss near Seller Creek. The study is restricted to one outcrop of gneiss and relies heavily on features of the rock which may be younger than the initial shear fabric of the Snowshoe Group rocks.

The mylonitic fabric is defined by elongated quartz grains supported in a matrix of fine grained recrystallized quartz, sericite, muscovite and variable amounts of sphene and opaques. Where primary feldspar is present the crystals are strained, but elongation of the grains takes place by fracturing. Because of this the feldspar occurs commonly as augen. Foliation of the mylonite is defined by trains of quartz crystals and by moderately anastamosing selvages now filled with fine opague material, sericite and muscovite. Some of the mylonite has aligned micas distributed randomly throughout the matrix of fine grained recrystallized quartz. These mylonitic features are developed to some degree in most rocks of the Snowshoe Group. They are, however, gradational with flow features related to folding and therefore their regional extent depends upon a somewhat arbitrary division of the gradation.

The age of the mylonitization is confined between the age of the Snowshoe Group (pre-Permian) and the unconformably overlying Quaternary deposits. It can be further constrained through various assumptions and inferences. The youngest part of the Snowshoe Group may be Paleozoic and therefore shearing may be post-Cambrian. The peak of metamorphism mainly postdates the shear fabric. The time of freezing of the metamorphic minerals is not known but regionally is considered to be Jurassic. Locally garnet is involved in the shear, but it is not known whether this is due to younger superimposed shear or if the freezing of the garnet was a protracted event covering a considerable period. The most precise age of the mylonite is determined by first assuming it was the result of the emplacement of the Crooked Ampibolite and secondly, by correlating the Crooked Amphibolite with the Antler Formation. The Antler Formation was thrust over the Barkerville and Cariboo terranes after the Permian and prior to the Upper Cretaceous.

The shearing event responsible for the mylonitization may be responsible for the development of a regional bedding-cleavage within the Snowshoe Group. These features are the first phase of the structural sequence of the Snowshoe Group structural package.

Bedding-parallel cleavage

Bedding-parallel cleavage occurs in almost all rocks of the Snowshoe Group and was recognized by all previous workers. It is folded and crosscut by spaced crenulation cleavages. Locally within the Ramos succession near Keithley Creek and Sovereign Mountain there are fold hinges with no bedding-parallel cleavage.

The bedding cleavage may be the regional consequence of the shearing event as portrayed by the mylonite concentrated at the contacts with the Crooked and Island Mountain amphibolites. Such a structural system would likely have isoclinal recumbent nappe-like folds, however, the size of those folds is unconstrained.

Rootless isoclinal folds

Campbell et al. (1973) and Sutherland Brown (1957) suggested the bedding cleavage was the consequence of interbed slip and that it had no associated isoclinal folds. Rootless hand-size isoclinal folds lie parallel to the bedding cleavage and some of them may have formed in the early stages of cleavage development. Known reactivation of the bedding cleavage surfaces leaves some doubt that all rootless isoclinal folds were formed early. Mesoscopic isoclines with the bedding cleavage as axial surfaces exist locally. Regional isoclinal folds associated with the cleavage are not unequivocally known (*see* Struik, 1982a). Shear structures are the first phase of the cleavage and fold progression of the structural package of the Snowshoe Group.

Ductile shortening

Cleavage, folds and faults included in this category represent a period of ductile shortening superimposed on the bedding-parallel cleavage and in part on the shear fabric. The shear fabric itself is a consequence of ductile shortening and the two categories in this way are intermeshed and transitional.

Cleavage and folds

Two or more sets of cleavage and folds are formed under conditions of ductile flow. The earlier, relatively more ductile set may include undifferentiated earlier cleavage and folds. Folds of both sets are asymmetric with moderately to shallowly dipping cleavages. The folds lean towards and the cleavage dips away from an axis (the Lightning Creek Anticlinorium, Fig. 49) that traverses the area from northwest to southeast. The folds are similar, range from open to isoclinal, and trend mostly northwest. Deviations from the northwest trend are most prevalent southwest of the Lightning Creek Anticlinorium (Fig. 50). The highest grade of metamorphism is recorded in, and partly overprints, the cleavage of the first set. Cleavage everywhere is defined by muscovite, although locally it coexists with chlorite or biotite.

Set 1 cleavage. This set is defined everywhere by muscovite and locally by biotite. Garnet grew approximately at the same time as the biotite, and together they represent the highest grade of metamorphism recorded in the area. The cleavage overprints the earlier shear foliation where they are parallel or nearly so. In the hinges of related folds it is a very closely spaced crenulation of the bedding cleavage. Locally it is more widely spaced and less well developed, especially in pure quartzite.

Isoclinal to tight folds are superimposed onto the shear fabric. Their characteristics vary according to the composition of the host rock. Quartzite supports more concentric folds than does pelite although all of this generation are flow folds (similar; *see* Fig. 36).

These folds and their cleavage represent ductile (relatively low viscosity) deformation during early stages of orogenesis.

Set 2 cleavage. The second set is predominantly a widely spaced crenulation in pelite and is poorly to undeveloped in quartzite. It dips steeply and crosscuts the previous generations in areas of chlorite zone metamorphism. Muscovite is partly parallel to the cleavage selvages which

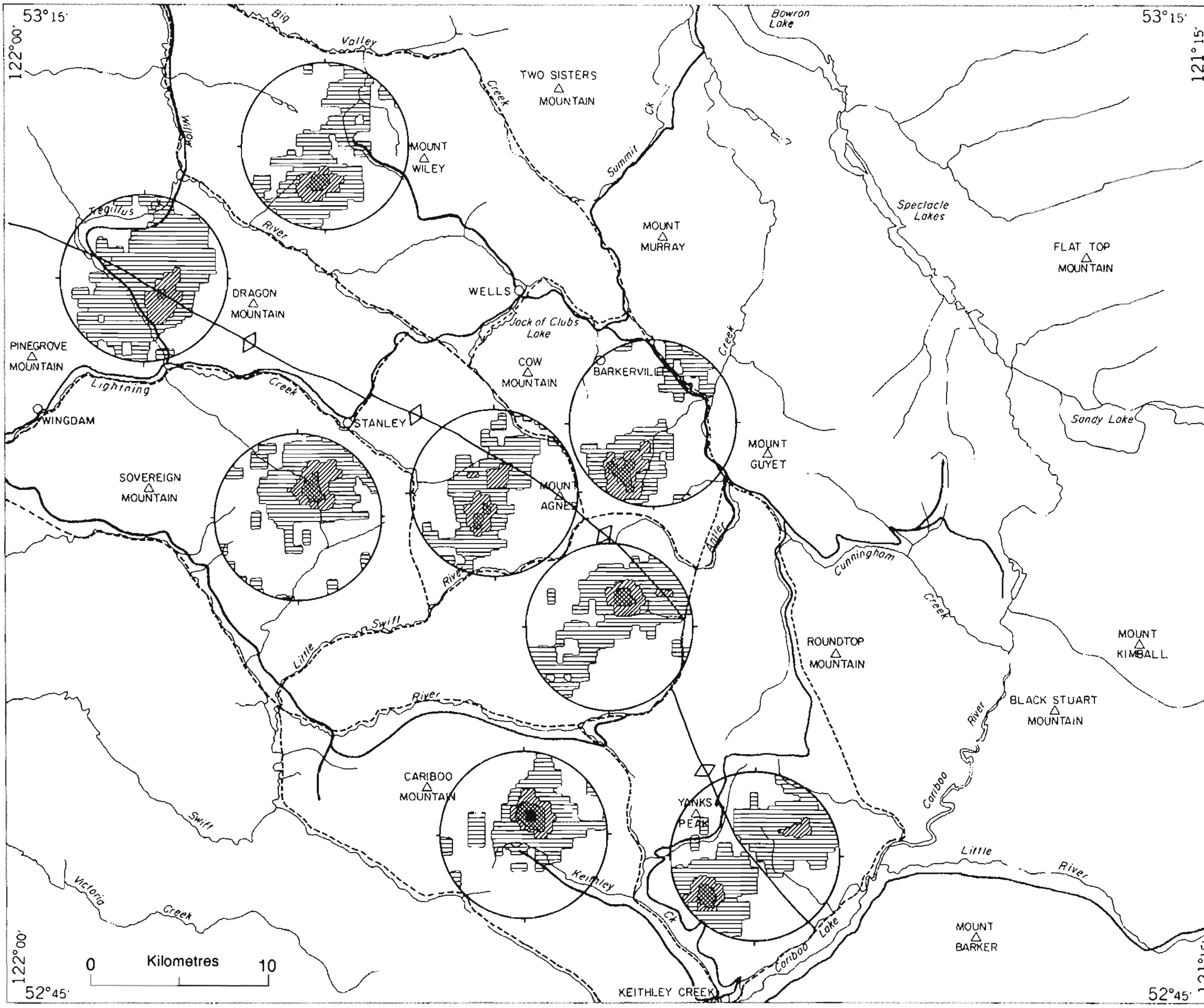


Figure 49. Map of the Barkerville Terrane divided into arbitrary structural domains by the dashed black lines. The equal area nets record the averaged plot of poles to the ductile style cleavage for each domain. The Lightning Creek Anticlinorium (plotted as the black line with open triangles on either side) is defined by the change of dip of the cleavage from southwest to northeast, a feature well displayed by the plots. The diffuse pattern of the plot in the northwest domain records a broad arching of the cleavage as opposed to the most southeasterly plot, which records tighter folding of the cleavage with the presence of a secondary cleavage.

are axial surfaces to folds of earlier muscovite. In zones of garnet grade metamorphism the cleavage may be equivalent to a more ductile flatter foliation parallel to the bedding cleavage. There may, however, be other interceding foliations that are confused because of overprinting during stable metamorphic conditions.

Folds formed along the set 2 cleavage are not always easily differentiated from earlier ones. Where the folds involve quartzite they are approximately concentric. They are open to tight, but locally isoclinal. The Lightning Creek Anticlinorium, in which layer-parallel cleavage is folded, has features of the set 2 type.

The Lightning Creek Anticlinorium. The Lightning Creek Anticlinorium (Fig. 49) is a late antiform superimposed upon a complex of earlier thrusts and folds. It traverses the area from Tregillus Lake southeastward to Cariboo

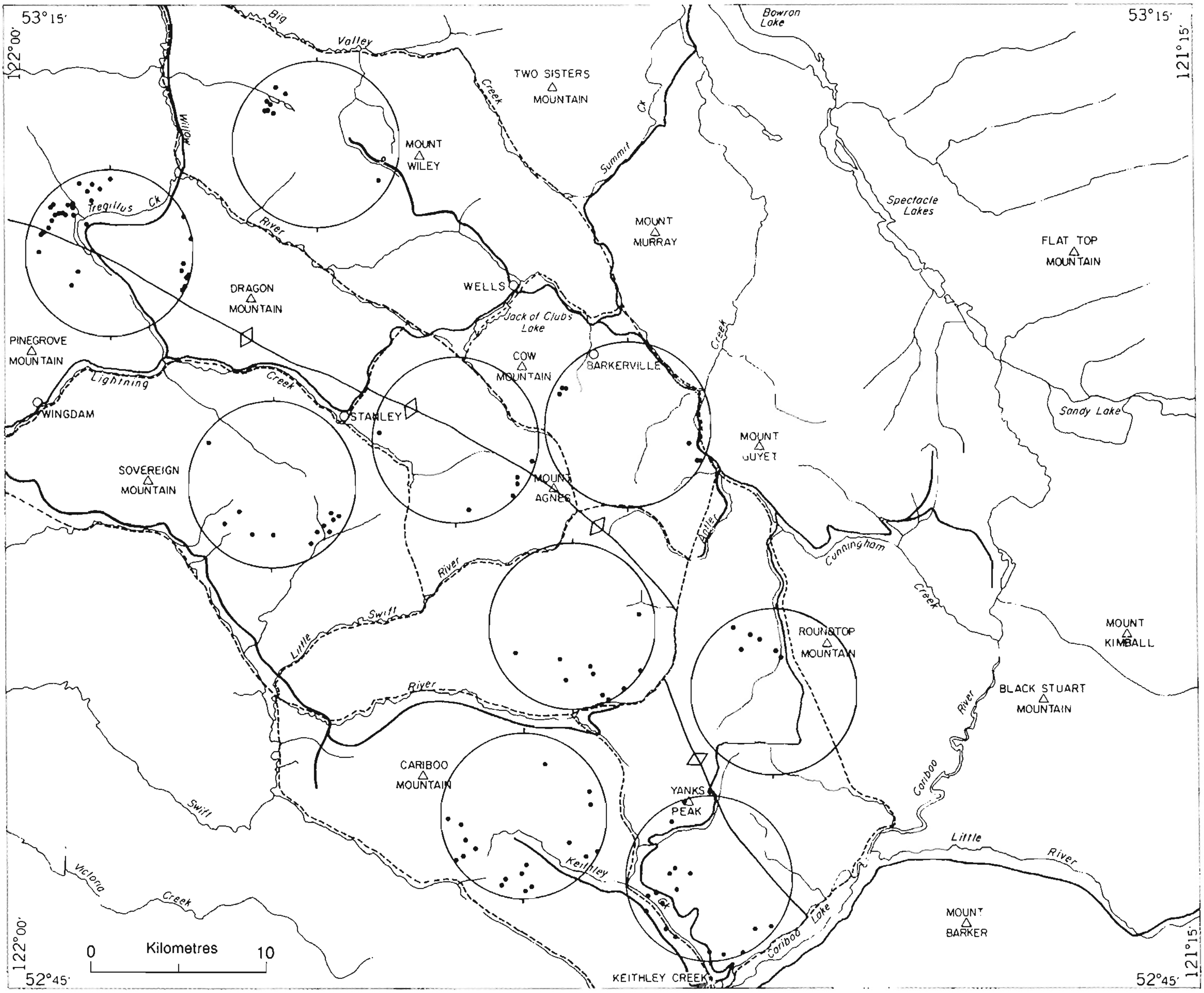


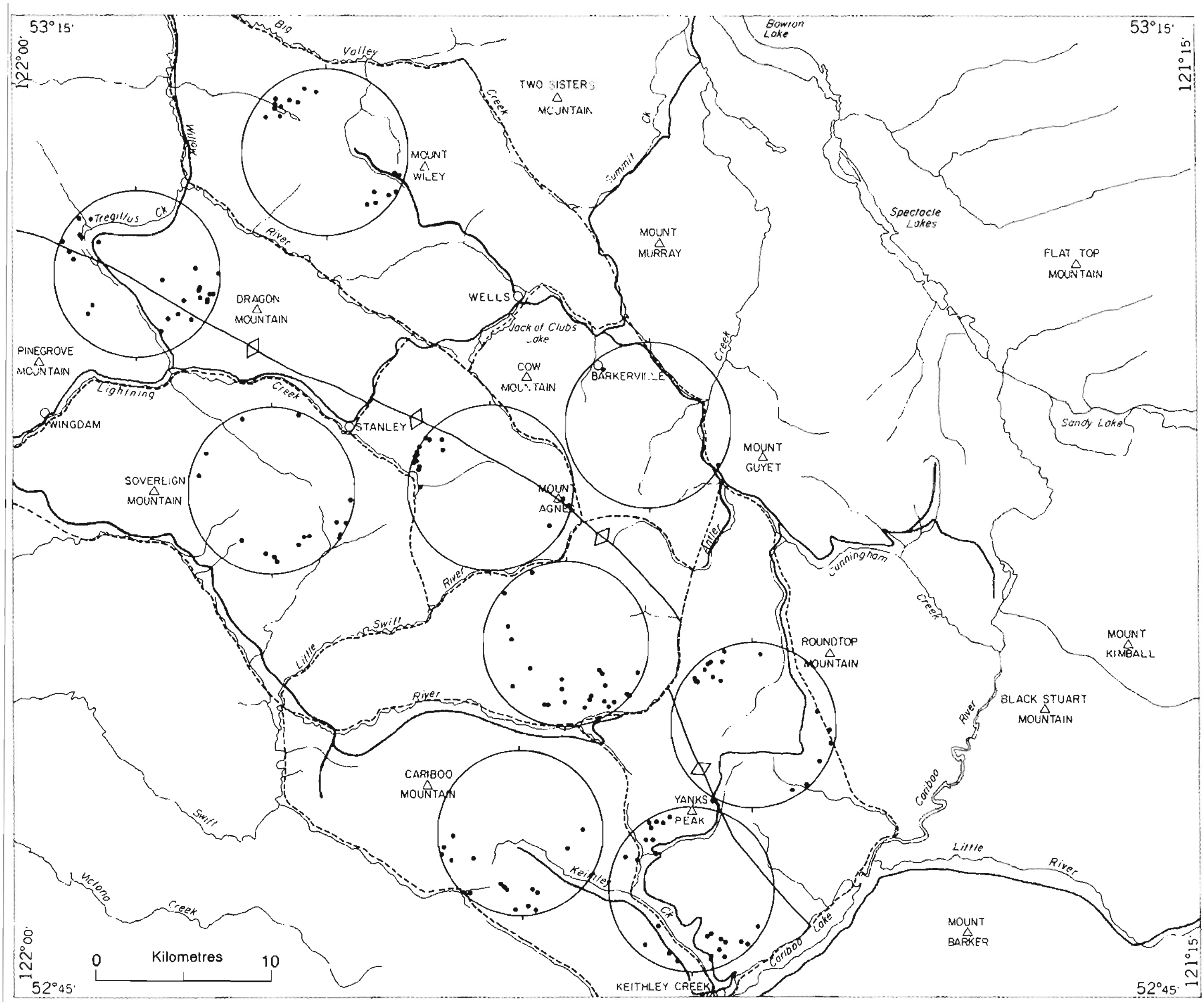
Figure 50. Maps of the Barkerville Terrane divided into arbitrary structural domains marked by black dashed lines. The equal area nets are plots of A) fold axes of the shear folds for each of the domains and B) ductile folds for each of the domains. Note the diversification of the fold axes attitudes to the southwest of the Lightning Creek Anticlinorium.

Lake. Bowman (1889) first described it; and subsequently, Johnston and Uglow (1926), Holland (1948, 1954), Sutherland Brown (1957), Campbell et al. (1973) and Struik (1982a, b). All suggested that the structure was an anticline.

The antiform folds the regional foliation, is symmetric and changes along trend from a broad arch to an open fold. It is intersected with a widely spaced axial surface cleavage where the hinge of the fold is tightest. Folds flanking the anticlinorium verge toward its axis (Fig. 49).

The extent of earlier folds that were folded by the antiform are unknown. Earlier I (1982a) speculated that the antiform may fold recumbent east-verging folds with limbs on the order of 4 to 8 km long, but such speculations depend on a thorough understanding of the stratigraphy, something not yet attained. Because the antiform folds the regional cleavage, it is assumed that it folds the isoclinal folds associated with that cleavage.

Campbell et al. (1973) and Struik (1981) described the cleavage of the antiform as arching through its hinge,



B

whereas Holland (1954) described a cleavage that fanned through the hinge. Near Yanks Peak, where Holland (1954) described the antiform, the hinge is tight and the fanned cleavage is likely the axial planar cleavage to the antiform. The arched cleavages are described from an area of the Lightning Creek Anticlinorium where it is very broad. In this area the axial planar cleavage did not form, and the arched cleavages are those of earlier structures, now folded about the axis of the Lightning Creek Anticlinorium (Fig. 49).

These folds represent deformation in a viscosity regime intermediate between ductile and brittle. They formed during intermediate stages of uplift of the sheared structural package of the Snowshoe Group.

Faults

No strike or cross-faults that can be associated with the period of ductile shortening have been positively identified. A postulated thrust fault (Keithley Thrust) that displaces garnet grade metamorphic rocks onto those of chlorite grade may have been active during the later part of the ductile shortening. Although the trace of it is moderately well defined the terminations of the fault are poorly constrained. The configuration of the metamorphic mineral distributions could also be the result of postmetamorphic folding but the northeastern transition from garnet to chlorite mineral assemblages is sharp and more characteristic of a fault.

Brittle structures

The cleavage, folds and faults of this category transpose, transform and translate, respectively, all earlier structures. Unlike the previous two structure groups, brittle structures include extensional features. A range of brittle structures produced at various times and high structural levels have been amalgamated into this part of the structural spectrum.

Cleavage and folds

The cleavage has an open crenulation style in phyllite but is not visible in quartzite. Earlier muscovite and biotite are crenulated with the formation of stylolitic-type cleavage selvages. Finely crystalline white mica forms part of the selvages. Chlorite, which has randomly overgrown earlier foliations and replaced biotite and garnet in variable amounts, is folded by, and overgrows, the crenulations. The chlorite retrograde metamorphism therefore, occurred after the late open structures of the ductile period of shortening.

Crinkles of phyllite, kinks, and widely spaced hand-size open folds have a more variable attitude than earlier folds. The crinkles and kinks have a single cleavage or joint respectively through their hinges. The kink folds have broken hinge zones. The open folds have a spaced crenulation cleavage.

Compressional

The strike fault through Mount Tom, Island, Cow and Richfield mountains may be a steep reverse fault, and therefore compressional. There is some indication that it is a dextral strike-slip fault (Struik, 1985). The strike faults are cut by northeast-trending extensional faults.

Extensional

Northeast- and north-trending steep faults, important hosts of gold-bearing quartz veins, appear to offset all other structures, and therefore are assumed to be the youngest feature in the Snowshoe Group structural package. It is not clear whether a distinct timing preference exists for the northeast- as opposed to the north-trending faults. Nor is it clear that the sense of movement is consistent in those groups. The more northerly-trending faults have right-lateral strike-slip as well as dip-slip, whereas those trending northeasterly may record more southside-down dip-slip motion than strike-slip. In this manner the northeasterly faults may record the extensional component to a northerly-oriented strike-slip regime. Vein filling within faults of both orientations indicates an overall extension. In combination, strike-slip and extension, the system, may be thought of as transtensional, related to Cretaceous and/or Early Tertiary strike-slip and uplift of the Omineca Belt metamorphic terranes.

Veins that are planar for outcrop-sized distances and are parallel to north-trending faults or northeast-trending joints are considered the latest generation. Locally these apparently young veins are transposed along the regional foliation (Fig. 51). Examples of late movement along the foliation surfaces suggest sporadic slip on the cleavage surfaces throughout the deformation of the Barkerville Terrane.

Figure 51. Flow-folded quartz vein in Snowshoe Group micaceous quartzite near Mount Borland. (GSC 191042)

The crinkles, kinks, small open folds, and extensional faults represent deformation in a high viscosity regime; sufficient to produce brittle features.

Summary

The structures of the Snowshoe Group package are divided into three groups; shear, ductile shortening, and brittle shortening and extension. The groups highlight changes in the strain conditions of protracted deformation (Fig. 52).

Shear, as indicated by mylonite (as defined by Wise et al., 1984), bedding cleavage, and rootless recumbent flow folds, dominates the earliest deformation. It is interpreted as the response to the eastward overthrusting of the Crooked Amphibolite along the Eureka Thrust. The thrusting is reasoned to be Late Triassic or Early Jurassic.

Ductile shortening, as denoted by flow cleavage and folds, and thrust faults, overprints the shear fabric. Conditions of metamorphism were sufficient throughout the shear and ductile shortening to crystallize and successively recrystallize muscovite, albite and quartz. Biotite and garnet crystallized locally and once during the period of ductile shortening. The ductile shortening is a response to compression, possibly representing collision between the two land masses of North America and Quesnellia. The collision began in the Early Jurassic, assuming the metamorphism records the collision and the metamorphism is Early Jurassic (Andrew et al., 1983).

The transition from ductile to less ductile compression may mark the change from Jurassic collision to Cretaceous transpression and strike-slip (*see* Gabrielse,

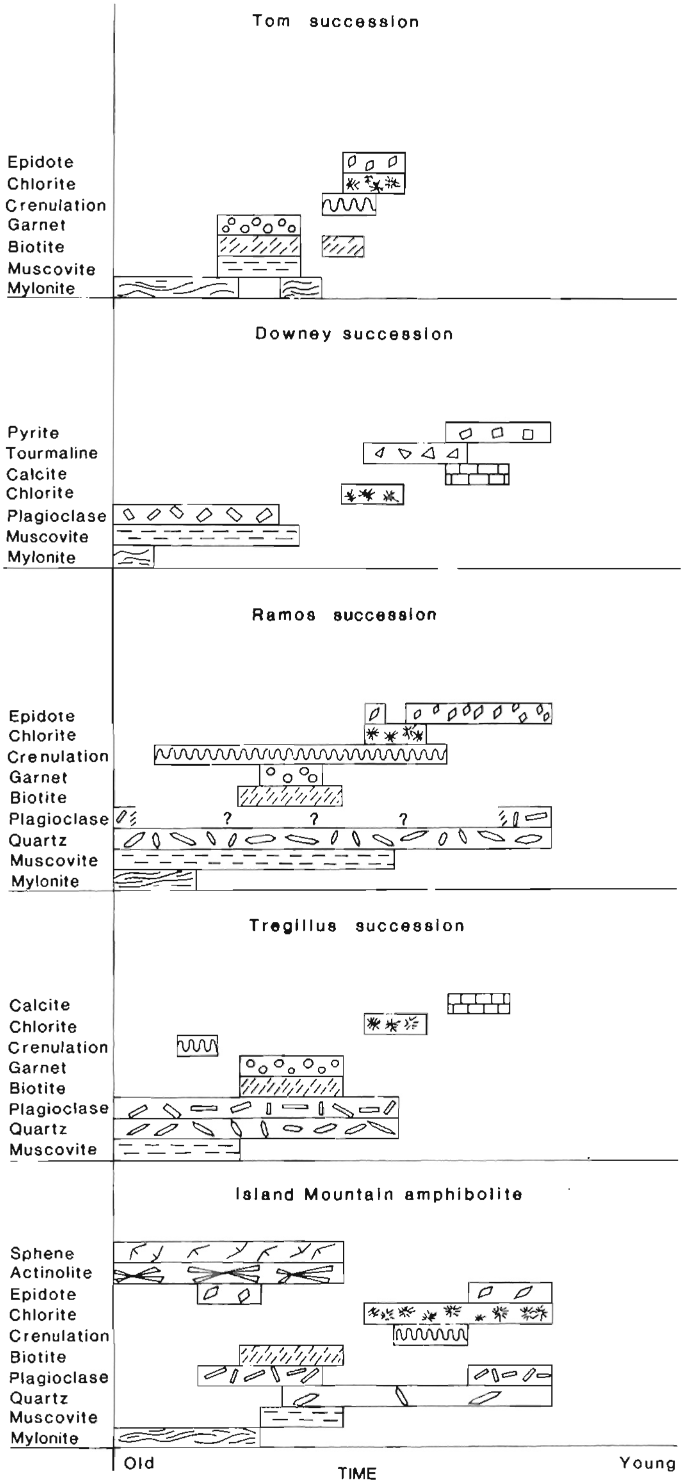


Figure 52. Relative timing of the crystallization of metamorphic minerals and the formation of cataclastic fabric and crenulations for rocks from units of the Snowshoe Group of the Barkerville Terrane.

1985 for documentation of Cretaceous strike-slip in northern British Columbia).

Brittle shortening and extension, as indicated by crenulations, kinks, broad open small-scale folds, and extensional faults, overprint all other structures and mark a period of chlorite retrograde metamorphism. This is thought to be a response to uplift of the structural package, at first remaining in a compressional environment and later changing to one of extension. The transition from compression to extension is inferred to be caused by the change from primarily Cretaceous transpressional tectonics to that of northerly-directed transtensional strike-slip (right-lateral). The initiation of the transtensional strike-slip tectonics is assumed to coincide with that of the Fraser River Fault system and Tertiary extension and uplift of the southern Omineca Crystalline Belt (Price, 1979; Monger, 1985).

ROCKS OF THE SLIDE MOUNTAIN TERRANE

Antler Formation

The Antler Formation as used in this report follows the definition of Sutherland Brown (1957,1963) and Campbell et al. (1973). It is included in the Slide Mountain Group.

The Antler Formation consists predominantly of pillow basalt, cherty argillite, argillaceous chert, cherty siltite, chert and diabase. It also contains lesser amounts of agglomerate, volcanic breccia, gabbro, greywacke, black slate and ultramafic rock. The basalt, chert and mafic igneous rock characterize it from other units of the area. Pillow basalt of the Waverly Formation resembles that of the Antler Formation.

The formation underlies the ridge system northward from Mount Tinsdale to Big Valley Creek. It has been mapped east, north, and west of the area by Sutherland Brown (1957,1963) and Campbell et al. (1973) in McBride map area, and Tipper (1961) in Prince George map area.

Exposures at mounts Tinsdale and Murray and Sliding and Two Sisters mountains are representative of the formation. Its thickness is poorly known because the internal stratigraphy is difficult to trace. Sutherland Brown (1963) estimated the thickness to be 1100 m at Mount Murray and somewhat thicker in the Palmer Range. This estimate may be excessive because the formation is imbricated by thrusting (Struik and Orchard, 1985).

The pillow basalt is dark grey green to grey, aphanitic to finely crystalline and is best developed on Mount Tinsdale and Mount Murray. Pillow structures in these areas range from 0.5 to 2 m across (Fig. 53). Matrix between the pillows is minor. Mineralogically the pillow basalt examined consists of:

Figure 53. Pillow basalt of the Antler Formation near Mount Murray. (GSC 191043)

Plagioclase,	
laths, 0.1 to 1 mm	10-35%
microphenocrysts, 0.4 to 3 mm An(25-35)	3- 8%
Augite,	
crystals, 0.05 to 0.6 mm	10-35%
microphenocrysts, 1 to 3 mm	0- 3%
Chlorite,	
filling 0.1 to 1 mm interstices	7-15%
Semiopaque groundmass,	
sheaths and irregular clumps of very fine chlorite, epidote and minor calcite	15-40%
Opaques,	
pyrite, 0.1 to 0.6 mm	0- 7%
others, 0.1 to 0.7 mm	3- 5%
Epidote, 0.05 to 0.2 mm	0- 5%
Quartz, 0.05 mm	0- 3%
Amygdules,	
0.1 to 0.4 mm from edge to core of quartz, calcite, chlorite, and epidote	0- 2%

The amount of augite decreases with a decrease in the overall crystal size. The finest grained volcanic examined contains less than the 10% augite (cited above), but on average the pillow basalts are more coarsely crystalline than this sample. The groundmass plagioclase laths are highly altered to a fine intergrowth of feldspar, chlorite, epidote, and minor calcite and sericite. Plagioclase phenocrysts are saussuritized. Augite is altered along fractures but is fresh in most specimens. Locally the basalt is picritic containing fine olivine phenocrysts.

Diabase forms sills, dykes and coarse equivalents of the finer pillow basalt. Its mineralogy is similar to that described for the pillow basalt except that the augite is coarser and the rock has an ophitic texture. Alteration products are similar, with the addition of pumpellyite as an interstitial filling and as rims to augite in diabase from the ridge between mounts Guyet and Tinsdale. Coarsely crystalline diabases, which are transitional to gabbro, do not have the fine grained feldspar-lath groundmass and instead have plagioclase crystals up to 4 mm long, interstitial to 0.5 to 4 mm euhedral augite and subhedral hornblende. The plagioclase is saussuritized and the hornblende partly altered to chlorite.

Agglomerates and breccias are mineralogically similar to the pillow basalts and are intimately associated with them.

Ultramafic rocks were found near Mount Murray and these have a mineralogy distinct from the diabase and its associated gabbro. The ultramafics consist of olivine, altered to talc and serpentine; clinopyroxene and/or orthopyroxene; hornblende and the alteration product, chlorite. From the limited number of samples collected they range from lherzolite to harzburgite. The stratigraphic relationship between the ultramafic and other Antler Formation rocks is unknown. The ultramafic rocks form pods and sheets mixed with pillow basalt, diabase, gabbro, and chert on Mount Murray.

Chemical analyses of Antler Formation mafic igneous rocks are listed in Table 15 and others have been presented by Sutherland Brown (1957, p. 44) Campbell (1971, p. 104 and 108), and Hall-Beyer (1976). Sutherland Brown's (1957, p. 44) samples 3 and 4, Table IV, may be from the Waverly Formation. Both Hall-Beyer (1976) and Campbell (1971) considered their chemical analyses to represent tholeiitic basalts and Hall-Beyer also postulated komatiitic affinities for some. Sutherland Brown (1957) reported his analyses as representative of spilites.

The cherty pelite and pelitic chert (Fig. 54) forms 1 to 7 cm even beds varying from grey to light grey, grey green to olive, very light green and maroon. Thicker beds are primarily of pelitic chert or cherty siltite and thinner interbeds are of cherty pelite. These well bedded "ribbon cherts" are best exposed on Sliding Mountain and Mount Tinsdale. Sutherland Brown (1957, p. 37-38) provided a good description of the mineralogical characteristics of the ribbon cherts. They consist of:

Detrital quartz,	
0.01-0.05 mm mostly but some to 0.2 mm	variable
Detrital muscovite and biotite,	
0.01-0.05 mm	0- 3%
Detrital plagioclase,	
0.03-0.07 mm	rare
Detrital accessories,	
zircon, tourmaline, epidote(?) 0.03-0.05 mm	0- 1%
Matrix chert,	
pervasive through various layers	20-90%
Ovoids of chert,	
probably recrystallized radiolaria, .05-.2 mm	5-25%
Secondary micas,	
chlorite and sericite may or may not be aligned, stilpnomelane is not aligned, very fine	0-10%

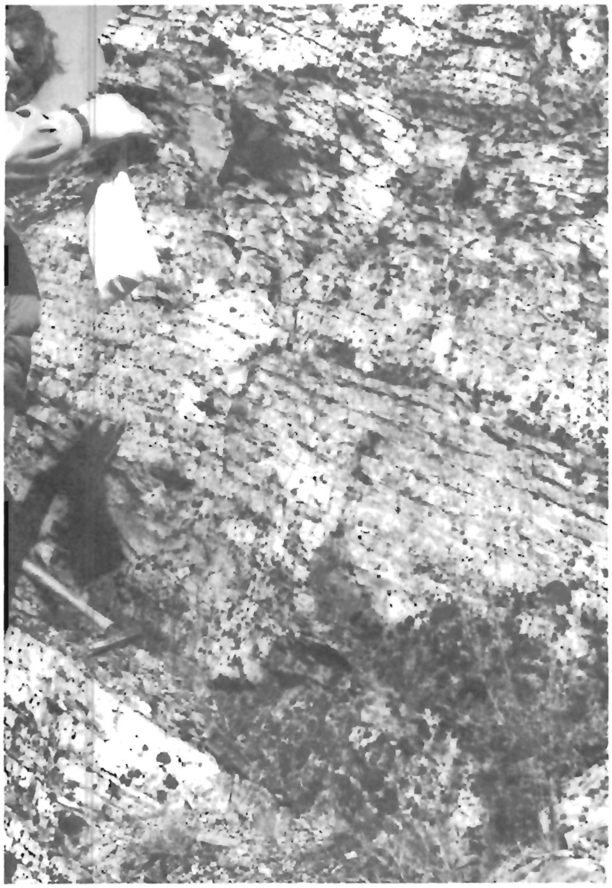


Figure 54. Ribbon chert of the Antler Formation on Sliding Mountain. (GSC 191044)

The white mica and chlorite, both primary and secondary, form concentrations defining fine and coarse layering. Detrital grains of quartz are rimmed by mica and locally deflect foliation where it exists. Pressure solution of detrital quartz grains has produced sutured and cuspate boundaries. Stylolites were found in all cherty sediments examined except the pure chert. Two generations of stylolites were noted in one sample. The prominent direction of stylolites is parallel to the bedding and this is offset by vertical stylolites: the two generations may have been penecontemporaneous. Veining generally at high angles to bedding existed prior to the stylolites along which they are offset. Veinlets are filled with various forms of silica. Cherts with no stylolites are suspected to have also undergone pressure solution, but did so without formation of stylolites because of their purity. Sutherland Brown (1957) reported that detrital mineral trains crossed veinlets showing their nondilational mode of formation.

A thin bed of tan siltstone interbedded with clean grey cherts on the ridge northwest of Mount Murray was seen nowhere else.

Age and correlation

Unsuccessful attempts were made to extract radiolarians from the cherty sediments following the procedure outlined by Pessagno and Newport (1972). Certain cherty pelites, however, yielded conodonts, which, although slightly etched, were sufficiently recognizable to be identified to species level in several cases. The faunal description provided by M.J. Orchard is given in Appendix A.

The described conodonts are the only datable fossils known from the Antler Formation and identify at least part of it as Early Mississippian to Early Permian.

The Antler Formation correlates with other volcanic assemblages of the Slide Mountain Terrane (Struik and Orchard, 1985), the Carboniferous or Permian Anvil Range Group and allochthonous coeval ophiolitic rocks of the Anvil Allochthon (Tempelman-Kluit, 1979), the Sylvester Group in Cassiar Mountains, the Nina Creek succession of Omineca Mountains and the Fennell Formation and Kaslo Group of southern British Columbia. Age relationships between the Antler Formation and underlying units indicate a thrust contact between them. Presently the age data and stratigraphic-structural relationships between the Antler and older formations require a minimum displacement for the Antler Formation of 50 km eastwards.

Crooked Amphibolite

This unit consists of amphibolite, serpentinite, sheared mafic and ultramafic rocks and talc. It is distinguished from the Antler Formation by its shear fabric and amphibolite, but is indistinguishable from the Island Mountain amphibolite of the Snowshoe Group.

The unit is exposed in a thin northwest-trending zone from Wingdam to Cariboo Mountain. It extends beyond the map area northward to near Prince George and southward to Clearwater. There is no type area; the name was derived from Crooked Lake where the unit is involved in a regional antiform-synform pair and was described by Campbell (1971). It varies in rock type along its length, being mostly serpentinite northwest of Cariboo River and amphibolite to the southeast. The Crooked Amphibolite may be up to 300 m thick, but is locally missing.

The contact to the east with Barkerville Terrane is a fault, marked by shear within adjacent rocks. To the west the contact with Triassic and younger rocks of Quesnel Terrane is thought to be stratigraphic because of regional continuity but locally displays shearing.

The amphibolite is dark green and weathers dark olive green. It consists primarily of amphibole, chlorite, plagioclase and epidote. Crystal size ranges from 0.01 to 1 cm on average with epidote being the smallest. The amphibole is mostly actinolite, which is locally mixed with hornblende, and defines the foliation. The plagioclase is oligoclase and the chlorite generally forms mats as part of the matrix. Accessory minerals include sphene and calcite.

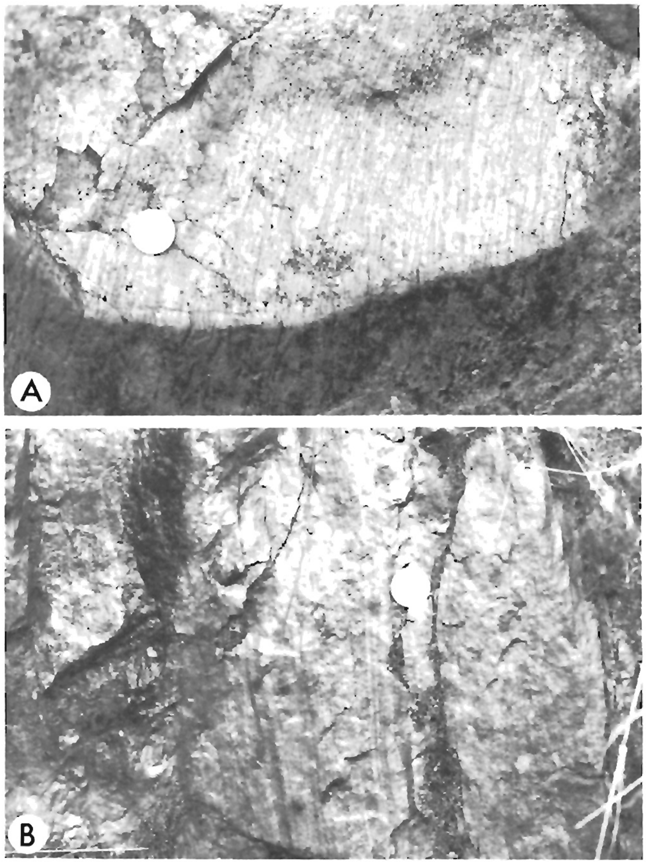


Figure 55. Serpentinite of the Crooked amphibolite, where it is A) foliated near Sovereign Mountain (Canadian nickel for scale; GSC 191045) and B) evenly shear banded near Etheridge Creek. (GSC 191046)

The serpentinite has not been examined microscopically. It varies from massive to brecciated, and local foliation does not parallel the regional trend (Fig. 55). Talc locally is associated with the serpentinite as discontinuous lenses parallel to the regional foliation. Ultramafic rock strongly foliated in the regional trend occurs locally with the serpentinite. It consists of serpentine and remnant pyroxene and plagioclase. Asbestiform serpentine rarely fills uncommon thin fractures. The resistant serpentinite forms irregular and steep topography.

Age and correlation

The Crooked Amphibolite may be the sheared and metamorphic equivalent of the Antler Formation and therefore part of the Slide Mountain Group, but no direct link between the units has been made. If they are equivalent the Crooked Amphibolite would be composed of Mississippian to Permian rocks.

Economic geology

Minor amounts of copper staining in basalt and diabase on Twin Sisters Mountain is the limit of recognized economic minerals in the Antler Formation.

Structure and metamorphism

Discussion of the structure and metamorphism of the Slide Mountain Terrane is divided into two sections; the first dealing with the Antler Formation and the second with the Crooked Amphibolite. Deformation within these rocks may have occurred during and after emplacement against adjacent terranes.

Antler Formation

Structures within the Antler Formation are divided into 3 categories. From oldest to youngest they are 1) soft sediment, 2) semiductile, and 3) brittle. The soft sediment deformation may have formed in a completely different structural regime than the other two categories.

Soft sediment deformation

The thin-bedded chert commonly hosts open to isoclinal folds which predate the cleavage, and are restricted to discrete horizons separated from undeformed chert by bedding-parallel detachment surfaces. Their form varies from recumbent long-limbed isoclines to box and irregular folds. The shear sense of the folds indicates mainly northeastward transport.

The paleoslope along which these rocks moved may have resulted from the same topography differential, causing thrusting of the Antler Formation over Barkerville and Cariboo terranes.

Semiductile shortening

Included within this category are thrusts, and folds and cleavage. The metamorphic minerals grow both parallel to and across the cleavage.

Thrusts. The thrust faults duplicate stratigraphy within the Antler Formation alone. They are determined from repetition of conodont-bearing chert horizons. Because the thrust faults are unseen, it is not known what their intersection angles with bedding are, and in what direction they cut downsection. The thrust sheets determined are on the order of 500 m thick, and contain the entire age range known for the Antler Formation. The faulting postdates Lower Permian rocks included in the thrust sheets. It signifies shortening possibly caused by flow of the Antler Formation down the paleoslope (recorded by the soft sediment folds) and/or represent duplication during the emplacement of the terrane.

Folds and cleavage. Folds involve all rocks of the Antler Formation but are more obvious or prevalent in the sedimentary rocks. They are of all scales, generally open to tight and rarely isoclinal. Their fold axes orientations plot on a great circle and not as a single average direction (Fig. 56). Cleavage is axial planar to the folds (Fig. 56) and parallel to bedding. Sericite and dissolution selvages which define the cleavage surfaces, increase in abundance with the decrease in silica. Pure cherts display no cleavage and pelites the most. Dissolution selvages that form cleavage surfaces may be planar-parallel, in "hourglass" form about pinching points, or stylolitic. The coarsest selvages are irregular. Selvages form along impurity-rich lenses and beds and are generally layer-parallel. Stilpnomelane and pumpellyite randomly grow within the Antler Formation rocks. Conodonts extracted from the chert have alteration indexes (Epstein et al., 1977) of 3 to 5, denoting maximum regional paleotemperatures of 200 to 350°c.

The fold axes lie along the great circle that is the average orientation to the ductile folds within the Cariboo Terrane. They therefore could be related to the same structural history. Such a configuration is caused by folding of nonparallel surfaces, suggesting the rocks of the Antler Formation were deformed prior to folding.

Brittle shortening and extension

The sedimentary rocks of the Antler Formation locally have crenulations and kinks that fold cleavage. The crinkle axes parallel the semiductile fold axes, whereas the orientation of the kinks is highly variable.

Upright faults have northeast and northerly trends, and transect the formation and its boundaries (for example, the Antler Creek fault). Joint sets throughout the volcanic rocks are well developed and are generally perpendicular to the strike of the cleavage (Fig. 57). All upright faults appear to postdate the emplacement of the Antler Formation.

Crooked Amphibolite

Unlike the Antler Formation the Crooked Amphibolite is dominated by shear foliation which parallels the contact with the Barkerville Terrane. It pervades every rock type in the unit except serpentinite. The shear fabric has no associated folds. Metamorphic chlorite, actinolite and hornblende grow with long axes in the shear surface. There are no randomly orientated metamorphic minerals.

The shear foliation formed during thrusting of the Crooked Amphibolite. The thrusting may have had multiple episodes of movement, as deduced from the lack of overgrowing metamorphic minerals.

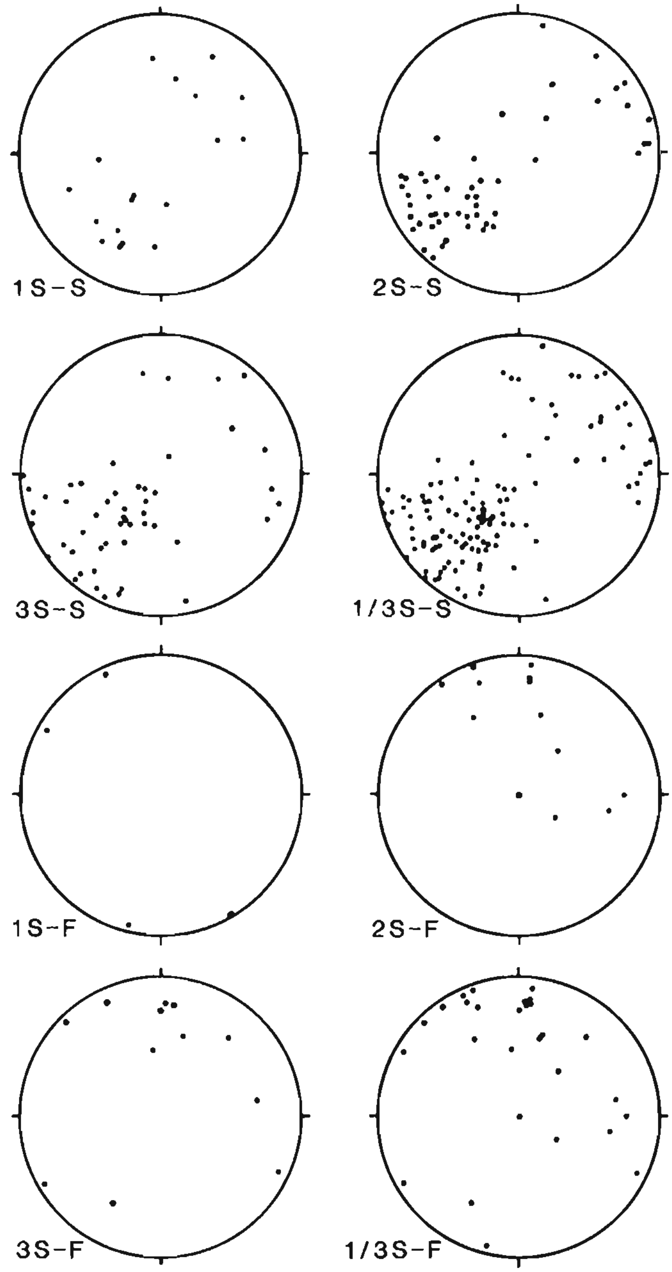


Figure 56. Equal area net plots of the poles to cleavage and the trend of fold axes from the Slide Mountain Terrane Antler Formation. 1S-S, 2S-S and 3S-S are plots of the poles to cleavage from the 1S, 2S and 3S structural domains as shown on Figure 25. These three plots are combined in 1/3S-S. 1S-F, 2S-F and 3S-F are plots of the trends of the folds from the 1S, 2S and 3S structural domains as shown on Figure 25. These three plots are combined in 1/3S-F.

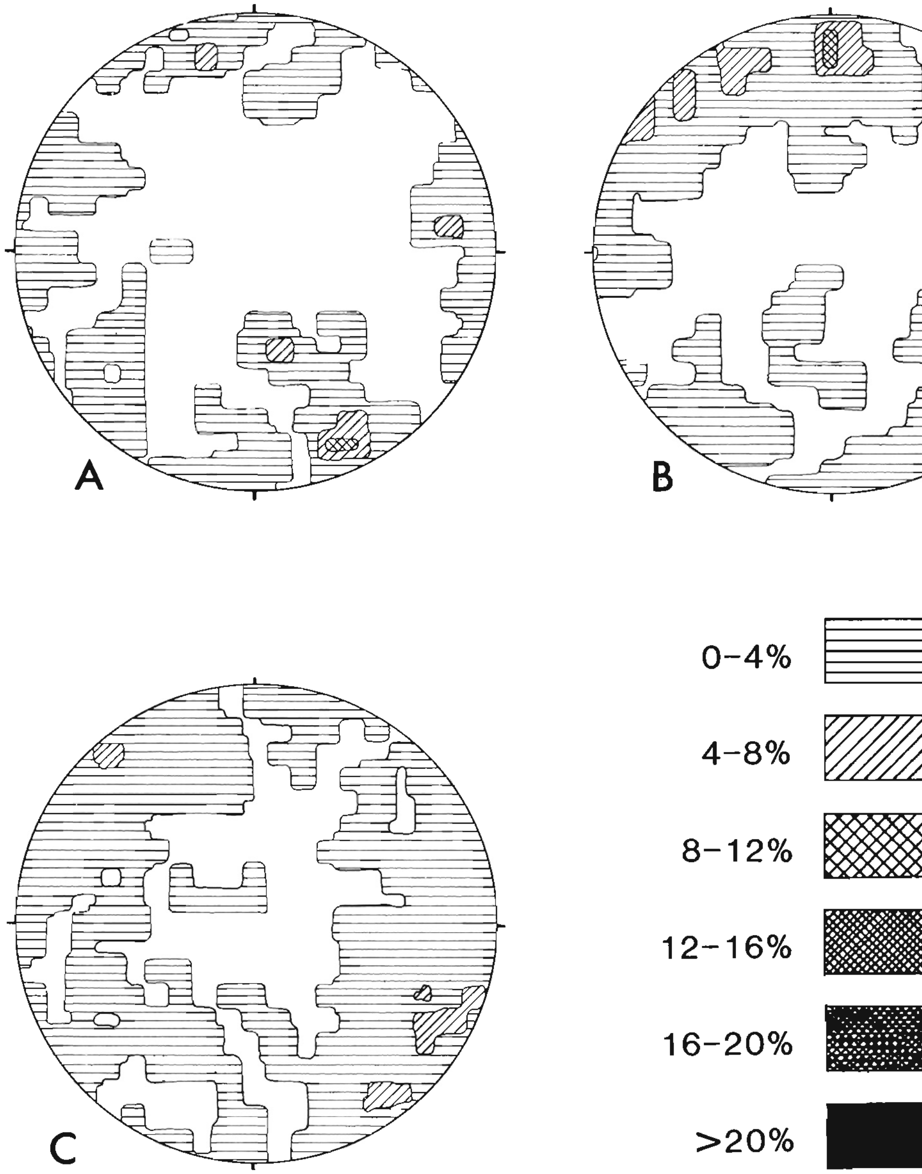


Figure 57. Averaged equal area net plots of the poles to joints in the Antler Formation, where A, B and C are from structural domains 1S, 2S and 3S (Fig. 25) and have 59, 135 and 151 points respectively.

ROCKS OF THE QUESNEL TERRANE

The Quesnel Terrane (Quesnellia) consists of Triassic and Jurassic pelitic and volcanic rock lying west of the Slide Mountain (Crooked Amphibolite) and Barkerville terranes. The boundary with the Slide Mountain Terrane may be depositional, unlike the faults that bound other terranes. Quesnellia extends north and south beyond the limits of the map area (Tipper, 1984). Although it overlies the Crooked Amphibolite it is not known to overlie the Antler Formation.

Stratified rocks

Quesnellia consists of 2 sequences within the map area, older black phyllite and younger mafic volcaniclastics.

Black phyllite

The black phyllite sequence consists of black pelite, siltite, limestone and conglomerate. The pelite is commonly slate or phyllite and may be locally siliceous imposing the character of argillite. The presence of 2 to 40 cm limestone beds dispersed throughout the pelites tends to distinguish it from higher pelites of Quesnellia, but not from units such as the black pelite of the Black Stuart Group of the Cariboo Terrane and the Hardscrabble Mountain or parts of the Harveys Ridge successions of the Barkerville Terrane.

The black phyllite unit overlies the Crooked Amphibolite in most places but is locally missing, as near Cariboo Mountain. No type section has been established for this unit, but for reference the logging area east of Wingdam Lake and the gorge of Fontaine Creek, where it meets the Swift River road, can be used. Thickness of the black phyllite is unknown due to poor exposure and structural complexity, but could be 400 m or more.

The character of the black pelite varies from thinly platy to argillite depending on the concentration of silica. Its 1 to 6 cm thick beds generally are more abundant than interbedded siltite or limestone.

The siltite is generally dark grey to grey and makes up approximately 15% of the interbedded sequence with black pelite.

The limestone is dark grey to black and weathers rusty brown. It is finely crystalline, massive and cut by numerous white calcite and less quartz veinlets. The beds are 2 to 40 cm thick and occupy approximately 5% or less of the black phyllite unit. Locally calcareous siltstone to fine grained quartzite has the weathering characteristics of the limestone.

The conglomerate is light grey consisting of rounded pebbles to small boulders of quartzite and smaller clasts of the same with quartz and feldspar. The matrix is mainly silica and locally dolomite. Channelling and associated scouring characterize the well-bedded conglomerate (Fig. 58). The unit has a maximum thickness of approximately 15 m and lies between Wingdam Lake and Highway 26. The conglomerate is bound by black phyllite of which the overlying rocks contain some calcareous siltstone.

Age and correlation

The age of the black phyllite unit is partly Middle Triassic (Ladinian) as determined from conodonts extracted from limestone near Benny Lake in Spanish Lake map area to the south. It may include older and younger rocks but is assumed to be entirely Triassic. The conodonts (GSC Locality C-102808) as described by M.J. Orchard were *Carinella?* sp. and *Neogondolella* spp. They were collected from a road cut along the Abbot Creek forestry road (Lat. 52°33'16"; Long. 121°18'02"). The black phyllite is correlated to the south with the Slocan Group and Tsalkom Formation. The conglomerate may be related to the Triassic conglomerate of Dragon Mountain near Quesnel (Struik, 1984).

Mafic volcaniclastics

The mafic volcaniclastic rocks are mainly basalt and andesite agglomerate and tuff. These rocks are distinguished from the volcaniclastics of the Downey succession by the coarse clastic detritus.

The mafic volcaniclastic rocks at one locality west of Cariboo Mountain directly overlie Snowshoe Group, but neither the contact relationships or thickness are known.

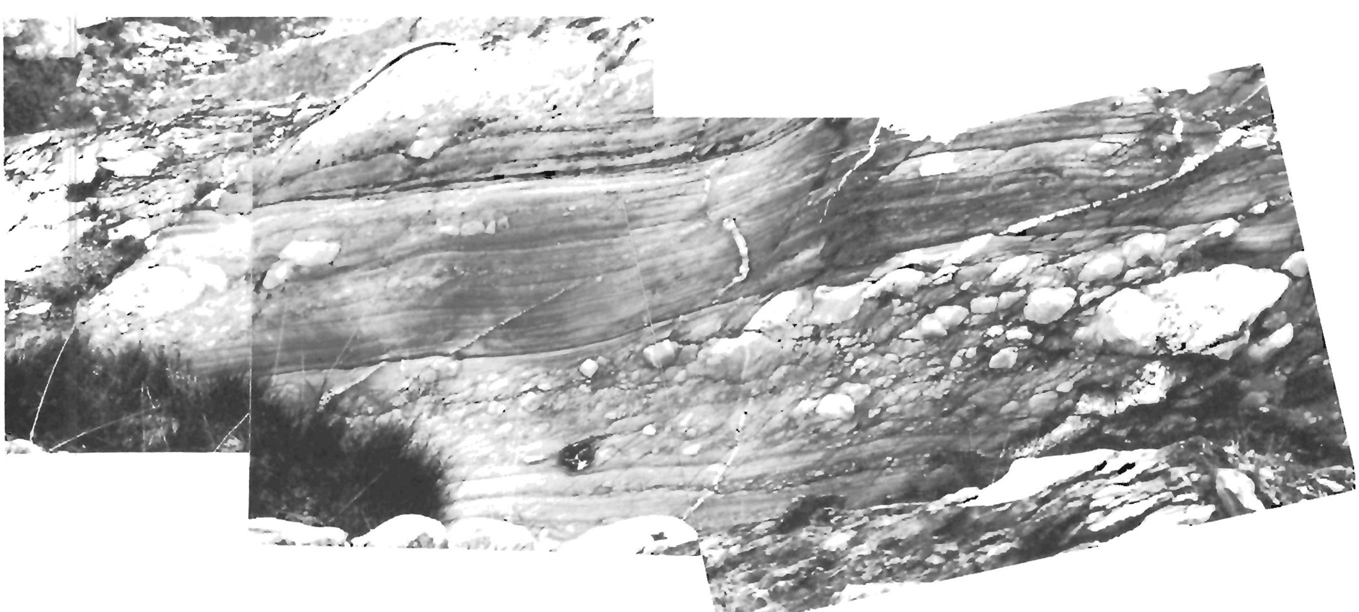


Figure 58. Conglomerate within the Triassic black pelite of the Quesnel Terrane on Highway 26 near Wingdam. Although shown horizontal, the bedding in the outcrop is nearly vertical. For scale the thickness of the beds shown is approximately 3 m. (GSC 191047)

Age and correlation

These rocks were mapped by Campbell (1978) as Upper Triassic or Lower Jurassic (unit TJa) of the Quesnel River Group and are accepted as such here.

Economic geology

Within the confines of the detailed mapping, Quesnel Terrane hosts no economic minerals in lode deposits. Placer gold is found near Wingdam in gravels overlying the black phyllite unit, but there appears to be no relationship to bedrock, because gravels overlying Snowshoe Group rocks also contain some gold in the vicinity.

Both the black phyllite and Triassic volcaniclastic units host concentrations of copper, gold and lead to the southwest of the area along a trend from Quesnel River southeast to Eureka Mountain (Saleken and Simpson, 1984). The minerals occupy quartz veins and country rock in seemingly stratabound replacements or as disseminations in alkalic plutonic rocks (ibid).

Structure and metamorphism

The structure and metamorphism were insufficiently studied because of the limited mapping of the Quesnel Terrane. Ductile to semiductile folds and cleavage are overprinted by local shear and crenulations.

The cleavage is southwest dipping, penetrative, commonly parallel to bedding and the axial planes of open, tight and isoclinal folds. It is defined by pressure solution selvages and accumulations of parallel, finely crystalline white mica. Locally it defines the shear foliation in phyllonite. The fold style varies with lithology; it is parallel in pelite and concentric in quartzite or siltite. Folds trend west-northwest and verge north-northeast. Stretched clasts in the Wingdam conglomerate are elongate in the same direction as the trend of the fold axes.

Shear, as recorded by zones of phyllonite, appears as bands some 1 to 15 m thick and parallels the contact with the Crooked amphibolite. The shear obliterates folds and is itself affected by crenulations.

Crenulations trend at 120?and have axial surfaces that dip to the southwest. They may be related to regional folds such as those marked by the Crooked amphibolite to the southeast.

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APPENDIX A
FOSSIL LOCALITY DESCRIPTIONS

The following conodonts have been identified by M.J. Orchard of the Geological Survey of Canada.

Greenberry Formation

C-86270 and C-86271
From grey crinoidal limestone on the small creek just east of where Sugar Creek enters Big Valley Creek (NTS N589075 E58910):
'Bispathodus' stabilis (Branson & Mehl) s. l.
Indeterminate platform and ramiform elements.
Age: Late Devonian-Early Mississippian

C-86265 and C-86266
From crinoidal limestone 1 km WSW of the small lake draining into Two Bit Creek (NTS N5896515 E58981):
Polygnathus communis communis Branson & Mehl
Indeterminate platform and ramiform elements
Age: Late Devonian-Early Mississippian

C-86260 and C-86261
From grey crinoidal limestone at the upper fork of Two Bit Creek (NTS N589536 and E5914):
Bispathodus aculeatus aculeatus (?) (Branson & Mehl)
Ramiform elements
Age: Late Devonian-Early Mississippian

C-86272
From grey crinoidal limestone on a logging road 800 m NW of the small lake draining into Two Bit Creek (NTS N589718 E591025):
Eotaphrus burlingtonensis (?) Pierce & Langenheim
A lot of fish teeth
Age: Early Mississippian, l. Tournaisian

C-101811
From sheared grey crinoidal limestone near the mouth of Stewart Creek directly overlain by Pennsylvanian black micritic limestone of the Alex Allan Formation (NTS N589394 E5944):
'Bispathodus' stabilis (Branson & Mehl) s. l.(?)
Polygnathus communis communis Branson & Mehl
Age: Late Devonian-Early Mississippian

C-86254/5/6
From grey crinoidal limestone 400 m east of the junction of Cafe and Big Valley creeks (NTS N5893325 E59495):
Polygnathus communis communis Branson & Mehl
Age: Late Devonian-Early Mississippian

C-102686
From grey crinoidal limestone in the creek emptying into Big Valley Creek east of Cafe Creek (NTS N589323 E59527):
'Bispathodus' stabilis (Branson & Mehl) s. l.
Hindeodus cristulus (Youngquist & Miller)
Patrognathus? sp. A
Polygnathus sp. B. (?)
Age: Tournaisian

C-102691
From grey crinoidal limestone in the creek emptying into Big Valley Creek east of Cafe Creek (NTS N589323 E59527):
Clydagnathus n. sp. A.
Mestognathus n. sp.
Polygnathus communis communis Branson & Mehl
Age: l. Tournaisian-e. Visean

C-102562
From grey crinoidal limestone in the creek emptying into Big Valley Creek east of Cafe Creek (NTS N589323 E59527):
'Bispathodus' stabilis (Branson & Mehl) s. l.?
Hindeodus cristulus (Youngquist & Miller)
Mestognathus n. sp.
Polygnathus communis communis Branson & Mehl
Age: l. Tournaisian-e. Visean

C-102687/101910/86251
From grey crinoidal limestone on a logging road south of the creek emptying into Big Valley Creek east of Cafe Creek (NTS N5893 E59556):
'Bispathodus' stabilis (Branson & Mehl) s. l.
Mestognathus n. sp.
Polygnathus communis communis Branson & Mehl
Age: l. Tournaisian-e. Visean

C-86252/3
From grey crinoidal limestone on the north bank of the creek 2 km due north of Nine Mile Lake (NTS N5893425 E5963):
Mestognathus n. sp.
Polygnathus communis communis Branson & Mehl
Age: l. Tournaisian-e. Visean

C-101925
From grey crinoidal limestone 4 m below C-101926 at 5000 feet elevation north of Eight Mile Lake (53°10'N, 120°32'W). This is believed to be the same locality as described by Campbell et al. (1973, p. 58-59):
'Bispathodus' stabilis (Branson & Mehl) s. l.
Hindeodus cristulus (Youngquist & Miller)
Age: Early Mississippian

C-101926
From grey crinoidal limestone 4 m above C-101925 (53°10'N, 120°32'W):
'Bispathodus' stabilis (Branson & Mehl) s. l.
Clydagnathus n. sp. A.
Gnathodus punctatus (Cooper) group
Age: l. Tournaisian-e. Visean

C-101927
From grey crinoidal limestone in a pod just north of C-101925/6 (NTS N589242 E5984):
'Bispathodus' stabilis (Branson & Mehl) s. l.
Hindeodus cristulus (Youngquist & Miller)
Age: Early Mississippian

C-86282
From crinoidal limestone directly overlying Guyet Formation conglomerate east of Alex Allen Creek (NTS N58884 E60045):
Polygnathus communis communis ? Branson & Mehl
Age: Late Devonian-Early Mississippian

C-86274
From grey partly silicified and sheared crinoidal limestone ESE of Greenberry Mountain (NTS N5885175 E60465):
'Bispathodus' stabilis (Branson & Mehl) s. l.
Polygnathus communis communis Branson & Mehl
Fish teeth
Age: Late Devonian-Early Mississippian

C-86275
From grey crinoidal limestone SE of Greenberry Mountain (NTS N58525 E60420):
Polygnathus communis communis Branson & Mehl
Fish teeth
Age: Late Devonian-Early Mississippian

C-101925
From grey crinoidal limestone at the head of a tributary of Greenberry Creek (NTS N588516 E60595):
'Bispathodus' stabilis (Branson & Mehl) s. l.
Hindeodus cristulus (Youngquist & Miller)
Age: Early Mississippian

C-101924
From grey gastropod-crinoid limestone 1 km NNE of Waverly Mountain (NTS N58869 E60635):
Bispathodus aculeatus aculeatus (Branson & Mehl)
Gnathnodus cuneiformis Mehl & Thomas
Hindeodus cristulus (Youngquist & Miller)
Polygnathus communis communis Branson & Mehl
Polygnathus sp. A.
Pseudopolygnathus nudus Pierce & Langenheim
Age: late Tournaisian

C-86276
From partly silicified bedded grey crinoidal limestone 2 km SSW of Mount Tinsdale (NTS N58743 E615375):
'Bispathodus' stabilis (Branson & Mehl) s. l.
Eotaphrus burlingtonensis Pierce & Langenheim
Mestognathus n. sp. ?
Polygnathus communis communis Branson & Mehl
Polygnathus mehli Thompson
Polygnathus sp. B.
Protognathus cf. *P. praedelicatus* Lane, Sandberg, & Ziegler
Age: late Tournaisian

C-102688
From partly silicified bedded grey crinoidal limestone of the same outcrop as C-86276, 2 km SSW of Mount Tinsdale (NTS N58743 E615375):
'Bispathodus' stabilis (Branson & Mehl) s. l.
Clydagnathus n. sp. A
Doliognathus sp. A?
Eotaphrus burlingtonensis Pierce & Langenheim
'Hindeodella' segaformis Bischoff
Mestognathus n. sp.
Polygnathus communis communis Branson & Mehl
Polygnathus mehli Thompson
Polygnathus sp. B
Age: late Tournaisian

C-101912
From partly silicified bedded grey crinoidal limestone of the same outcrop as C-86276/102688 but 1 m lower in the section, 2 km SSW of Mount Tinsdale (NTS N58743 E615375):
'Bispathodus' stabilis (Branson & Mehl) s. l.
Doliognathus sp. A.?
Eotaphrus burlingtonensis Pierce & Langenheim
'Hindeodella' segaformis Bischoff
Polygnathus communis communis Branson & Mehl
Age: late Tournaisian

C-101911
From partly silicified bedded grey crinoidal limestone of the same outcrop as C-86276/102688/ 101912, 1 m above C-101912, 2 km SSW of Mount Tinsdale (NTS N58743 E615375):
'Bispathodus' stabilis (Branson & Mehl) s. l.
Clydagnathus n. sp. A.
Doliognathus sp. A?
'Hindeodella' segaformis Bischoff
Polygnathus communis communis Branson & Mehl
Age: late Tournaisian

C-102689/101894/101913/101914
From grey bedded crinoidal limestone sampled at approximately 1 m intervals, where C-102689 is a mixture of all levels and C-101894 is the lowest in the sequence, 2.5 km S of Mount Tinsdale (NTS N587315 E6168):

C-102689
'Bispathodus' stabilis (Branson & Mehl) s. l.
Clydagnathus n. sp. A.
Polygnathus bischoffi Rhodes, Austin, & Druce
Polygnathus communis communis Branson & Mehl
Polygnathus mehli? Thompson

C-101894
'Bispathodus' stabilis (Branson & Mehl) s. l.
Clydagnathus n. sp. A
Hindeodus cristulus (Youngquist & Miller)
Hindeodus n. sp. A
Polygnathus communis communis Branson & Mehl
Polygnathus mehli Thompson

C-101913
'Bispathodus' stabilis (Branson & Mehl) s. l.
Clydagnathus n. sp. A
Eotaphrus burlingtonensis Pierce & Langenheim
Hindeodus cristulus (Youngquist & Miller)
Polygnathus bischoffi Rhodes, Austin, & Druce
Polygnathus communis communis Branson & Mehl
Polygnathus sp. B

C-101914
'Bispathodus' stabilis (Branson & Mehl) s. l.
Gnathodus sp. indet.
Hindeodus cristulus (Youngquist & Miller)
Age: late Tournaisian

C-101917
From dark grey limestone upsection structurally from C-102689 series, 2.5 km south of Mount Tinsdale (NTS 58731 E61725).
'Bispathodus' stabilis (Branson & Mehl) s. l.
Clydagnathus n. sp. A
Gnathodus punctatus (Cooper) group
Gnathodus sp. indet.
Polygnathus communis communis? Branson & Mehl
Age: late Tournaisian

Alex Allan Formation

The age of the conodont faunules from the Alex Allan Formation ranges from Late Mississippian? to Early to Middle and possibly Late Pennsylvanian. The faunules are composed of conodonts representing distinctly different ages and are assumed to be a product of sedimentary mixing by reworking of older elements into a Middle or Late? Pennsylvanian limestone.

C-86264
From dark grey micritic limestone interbedded with dark grey shale 1.6 km west of Two Bit Creek (NTS N589697 E58632):
Gnathodus bilineatus (Roundy)
Gondolella laevis Kosenko & Kozitskaya
Idiognathodus — Streptognathodus plexus
Idioprioniodus sp.
Neognathodus sp. A

C-86257
From dark grey micritic limestone as an isolated outcrop in the creek 3.5 km due west of Two Sisters Mountain (NTS N589594 E59392):
Idiognathodus — Streptognathodus plexus

C-86259
From black micritic limestone pods in black shale overlying sheared crinoidal limestone in Stewart Creek near its junction with Big Valley Creek (NTS N589394 E5944):
Idiognathodus — Streptognathodus plexus

C-86273
From a 10 cm bed of grey weathering black micritic limestone directly overlying black shale of the Guyet Formation or Black Stuart Group, in Summit Creek (NTS N589015 E599925):
Gnathodus bilineatus (Roundy)
Idiognathodus — Streptognathodus plexus
Idiognathoides sinuatus Harris & Hollingsworth s. l.
Idioprioniodus sp.
Neognathodus sp. A

C-101121/101123/101124/10101/116301/116302/116303/116304/116305/116306
From individual beds of black micritic limestone separated by black shale of the type locality on the road from Wells to Bowron Lake, approximately 250 m north of its intersection with Alex Allan Creek (NTS N58936 E6031):
C-101121
Gnathodus bilineatus (Roundy)
Gnathodus punctatus (Cooper) group
Gondolella laevis Kosenko & Kozitskaya
Gondolella magna Stauffer & Plummer s. l.
Gondolella? sp. A
Idiognathodus — Streptognathodus plexus
Idioprioniodus sp.
Neognathodus sp. A
Neogondolella clarki (Koike)
Rhachistognathus primus Dunn
Streptognathodus elegantulus Stauffer & Plummer ?

C-101123
Gnathodus bilineatus (Roundy)
Gnathodus punctatus (Cooper) group
Gondolella laevis Kosenko & Kozitskaya
Gondolella magna Stauffer & Plummer s. l.
Gondolella? sp. A
Hindeodus sp.
Idiognathodus — Streptognathodus plexus
Idiognathoides sinuatus Harris & Hollingsworth s. l.
Idioprioniodus sp.
Neognathodus medadultimus Merrill?
Neognathodus sp. A
Neogondolella clarki (Koike)
Paragnathodus commutatus (Branson & Mehl)
Streptognathodus elegantulus Stauffer & Plummer

C-101124
Adetognathus sp.
Gnathodus bilineatus (Roundy)
Gondolella laevis Kosenko & Kozitskaya
Gondolella magna Stauffer & Plummer s. l.
Gondolella? sp. A
Idiognathodus — Streptognathodus plexus
Idiognathoides sinuatus Harris & Hollingsworth s. l.
Idioprioniodus sp.
Neognathodus sp. A
Neogondolella clarki (Koike)
Streptognathodus elegantulus Stauffer & Plummer

C-101101
Adetognathus sp.
Gondolella? sp. A
Idiognathodus — Streptognathodus plexus
Idioprioniodus sp.
Neogondolella clarki (Koike)

C-116301
Adetognathus sp.
Gnathodus bilineatus (Roundy)
Gondolella magna Stauffer & Plummer s. l.
Gondolella? sp. A
Idiognathodus — Streptognathodus plexus
Idiognathoides sinuatus Harris & Hollingsworth s. l.
Idioprioniodus sp.
Neognathodus sp. A
Neogondolella donbassica Kosenko
Rhachistognathus primus Dunn

C-116302
Geniculatus sp.
Idiognathodus — Streptognathodus plexus
Idioprioniodus sp.
Streptognathodus elegantulus Stauffer & Plummer

C-116303
Gondolella magna Stauffer & Plummer s. l.
Gondolella? sp. A
Idiognathodus — Streptognathodus plexus
Idiognathoides sinuatus Harris & Hollingsworth s. l.

C-116304
Gondolella? sp. A
Idiognathodus — Streptognathodus plexus
Idiognathoides sinuatus Harris & Hollingsworth s. l.
Idioprioniodus sp.
Neognathodus sp. A
Neogondolella clarki (Koike)
Streptognathodus elegantulus Stauffer & Plummer

C-116305
Gondolella magna Stauffer & Plummer s. l.
Gondolella? sp. A
Hindeodus sp.
Idiognathodus — Streptognathodus plexus
Idiognathoides sinuatus Harris & Hollingsworth s. l.
Idioprioniodus sp.
Neognathodus sp. A
Paragnathodus commutatus (Branson & Mehl)

C-116306
Gnathodus bilineatus (Roundy)
Gondolella magna Stauffer & Plummer s. l.
Idiognathodus — Streptognathodus plexus
Idiognathoides sinuatus Harris & Hollingsworth s. l.
Idioprioniodus sp. ?
Neognathodus sp. A
Rhachistognathus primus Dunn

Unnamed limestone of Spectacle Lakes

C-102692
From grey fusulinid limestone on the southern end of the island at the north end of Spectacle Lakes (NTS N58942 E61335):
Adetognathus sp.
Declinognathodus noduliferus (Ellison & Graves)
Gondolella magna Stauffer & Plummer s. l.
Hindeodus sp.
Idiognathodus — Streptognathodus plexus

Idioprioniodus sp. ?
Neognathodus sp. A?
Neogondolella clarki (Koike)

Sugar limestone

C-102696
From grey crinoidal limestone at the pass between Sugar and Hardscrabble creeks (NTS N58909 E58866):
Neogondolella cf. *N. bisselli* (Clark & Behnken)
Sweetognathus cf. *S. whitei* Rhodes

C-102695
From grey crinoidal limestone, poorly exposed, along trend of C-102696, at the head of Sugar Creek (NTS N589085 E58873):
Idiognathodus sp. indet.
Neogondolella cf. *N. bisselli* (Clark & Behnken)
Sweetognathus cf. *S. whitei* Rhodes

C-102694
From grey crinoidal limestone along trend of C-102695 at the head of Sugar Creek (NTS N589082 E588925):
Neognathodus? sp.
Neogondolella cf. *N. bisselli* (Clark & Behnken)
Streptognathodus cf. *S. elongatus* Gunnell
Sweetognathus cf. *S. whitei* Rhodes
Gen. et sp. nov.

C-86268/C-86269
From grey crinoidal limestone, partly silicified, sheared, at the head of Sugar Creek (NTS N589082 E5889):
Hindeodus sp.
Neogondolella cf. *N. bisselli* (Clark & Behnken)

C-102693
From grey crinoidal limestone, partly silicified resting with sharp contact on black siltite of the Snowshoe Group on the west slope of Hardscrabble Mountain (NTS N58907 E59025):
Gondolella magna Stauffer & Plummer s. l.
Neogondolella cf. *N. bisselli* (Clark & Behnken)
Streptognathodus cf. *S. elongatus* Gunnell
Streptognathodus n. sp. A

Slide Mountain Group — Antler Formation

All conodonts from the Antler Formation are from pelitic chert on Sliding Mountain. They are listed in structural order from the top in the northeast (NTS N588965 E60154) to the bottom in the southwest (NTS N588907 E60044). From the age progression the conodonts suggest that there is a thrust fault duplicating a Permo-Carboniferous sequence.

C-86281
Same location as C-86277
Neogondolella cf. *N. bisselli* (Clark & Behnken)
Age: Early Permian

C-86277
From green cherty argillites directly overlain by pillow basalt on Sliding Mountain (NTS N588965 E60154):
Idiognathoides sp.
Idiognathodus — Streptognathodus plexus
Age: Early-Middle Pennsylvanian

C-86278
Same location as C-86277
Idiognathoides sp.
Idiognathodus — Steptognathodus plexus
Age: Early-Middle Pennsylvanian

C-86279
Same location as C-86277
Idiognathoides sp.
Neogondolella cf. *G. clarki* (Koike)
Gondolella cf. *G. laevis* Kosenko & Kozitskaya
Age: Early-Middle Pennsylvanian

C-86280
Same location as C-86277
Idiognathoides sp.
Idiognathodus — Streptognathodus plexus
Polygnathus sp.
Rhachistognathus? N. sp.
Streptognathodus expansus Igo & Koike
Age: Probably Early Pennsylvanian. *Polygnathus* is presumably reworked.

C-102664
Taken 2 m above C-102665
Gondolella? sp.
Age: Probably Pennsylvanian

C-102665
NTS N58974 E60155
Idiognathodus — Streptognathodus plexus
Gondolella cf. *G. laevis* Kosenko & Kozitskaya
Age: Early — Middle Pennsylvanian

C-102666
Same stratigraphic level as C-102665
Idiognathodus — Streptognathodus plexus
Age: Pennsylvanian — Early Permian

C-102667
A 1 m chip sample 1 m below C-102666.
Idiognathodus — Streptognathodus plexus
Streptognathodus sp. A
Idiognathoides sp.
Gondolella cf. *G. laevis* Kosenko & Kozitskaya
Gondolella magna Stauffer & Plummer s. l.
Age: Probably Middle Pennsylvanian

C-102668
Taken 1 m below C-102667
Idiognathodus — Streptognathodus plexus
Age: Pennsylvanian — Early Permian

C-102669
Taken 2.5 m below C-102667
Neogondolella cf. *N. clarki* (Koike)
Idiognathodus — Streptognathodus plexus
Idiognathoides sp.
Age: Early — Middle Pennsylvanian

C-102670
Taken 5 m below C-102669
Neogondolella cf. *N. clarki* (Koike)
Idiognathodus — Streptognathodus plexus
Idiognathoides sp.
Age: Early — Middle Pennsylvanian

C-102671
Taken 6 m below C-102670
Neogondolella cf. *N. clarki* (Koike)
Idiognathoidus — Streptognathodus plexus
Idiognathoides sp.
Rhachistognathus? sp.
Age: Early — Middle Pennsylvanian

C-102672
Taken 10.5 m below C-102669.
Idiognathoides sp.
Idiognathodus — Streptognathodus plexus
"Streptognathodus" expansus Igo & Koike
Gondolella cf. *G. laevis* Kosenko & Kozitskaya
Age: Early — Middle Pennsylvanian

C-102673
Taken 4 m below C-102672
Idiognathoides sp.
Age: Early — Middle Pennsylvanian

C-102674
Taken 24.5 m below C-102672.
Idiognathoidus — Streptognathodus plexus
Idiognathoides sp.
Gondolella sp.
Age: Early — Middle Pennsylvanian

C-102678
Taken 55.5 m below C-102674 (NTS N588443 E60146).
Paragnathodus ex. gr. *commutatus* (Branson & Mehl)
Age: Late Mississippian — Early Pennsylvanian

C-102679
(NTS N588928 E60082)
Polygnathus cf. *P. communis* Branson & Mehl
Pseudopolygnathus sp(p).
Siphonodella sp.
Age: Early Mississippian

C-102680
(NTS N58892 E6007)
Neogondolella sp.
Age: Permian

C-102681
(NTS N588909 E60055)
Neogondolella sp.
Age: Permian

C-102682
(NTS N588908 E60052)
Idiognathoides sp.
Idiognathodus — Streptognathodus plexus
Streptognathodus sp. A.
Gondolella cf. *G. laevis* Kosenko & Kozitskaya
Neogondolella cf. *N. clarki* (Koike)
Age: Probably Middle Pennsylvanian

C-102683
(NTS N588907 E60047)
Streptognathoidus expansus Igo & Koike
Idiognathodus — Streptognathodus plexus
Age: Early — Middle Pennsylvanian

APPENDIX B
STRAIN ANALYSIS OF GUYET FORMATION CONGLOMERATE

Introduction

Approximate strains were determined for 11 oriented specimens from the Guyet Formation conglomerate. The locations of these samples are shown in Figure 59. The specimens consist of sandy pebble to granule conglomerate and muddy pebble conglomerate to greywacke siltstones.

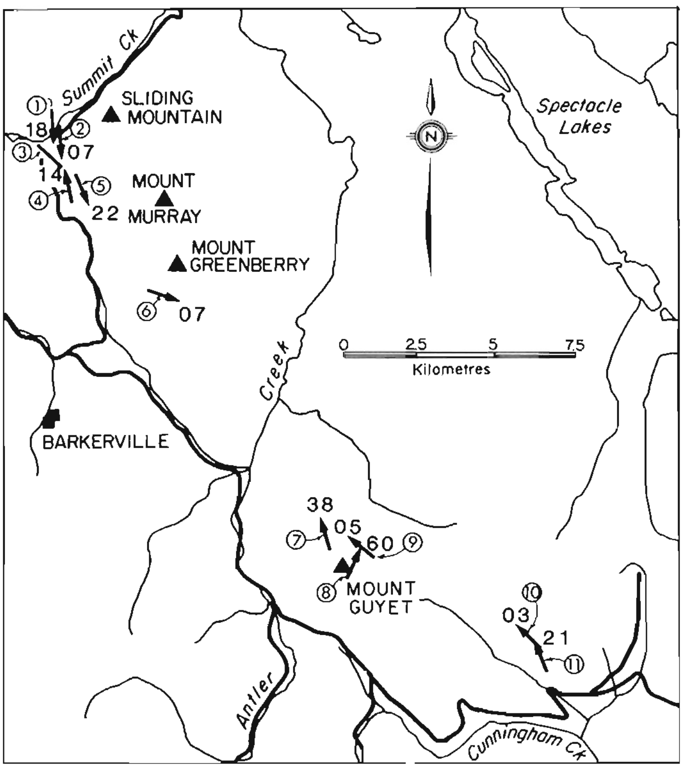

Figure 59. Location map of the Guyet Formation conglomerate samples analyzed for their amount of strain. Attitudes are of the calculated long axes of the strain ellipsoids.

The sampling reflects the uneven distribution of outcrop and the depositional irregularity of the Guyet Formation. The specimens were oriented using the technique described by Ramsay (1967, p. 191).

Method of strain analysis

The strain magnitude and direction were determined with the shape factor grid method of Elliott (1970), described fully by Elliott (1970) and Tobisch et al. (1977).

The procedure is a geometric analysis of the shape of a marker, the composition of which is not considered. For analysis the shape is required to be an ellipse defined by a perpendicular long and short axis fitted to the marker (Griffiths, 1967, p. 122). None of the markers are truly elliptic.

The procedure involves computer-generated polar plots of the shape and orientation of marker ellipses for three perpendicular sections of a rock. The position and shape of a marker upon the polar plots represents its final ellipse[1] as produced by some strain (represented as strain ellipse) upon an initial ellipse. The inter-relationship between these three ellipses is clearly illustrated by Ramsay (1967, p. 204-205).

The marker ellipses measured on one cut section of a strained rock will plot as a distribution on the polar plot. This distribution represents the effect of strain on an initial distribution of ellipse shapes.

Elliott (1970) describes 4 types of initial distributions of elliptic markers: 1) circular shapes, 2) randomly oriented elliptic shapes, 3) unimodal elliptic shapes, and 4) bimodal elliptic shapes. These distributions are assessed and redefined by Boulter (1976). Elliot's method requires choosing a strain ellipse that best destrains the distribution of final ellipse shapes to a distribution of initial ellipse shapes that conforms to one of the 4 theoretical initial distributions. The destraining procedure is done by computer within this study and the program is available through the archives of the Geological Survey of Canada.

Determination of the best-fit ellipsoid

The three-dimensional strain ellipsoid is calculated from the three two-dimensional strain ellipses determined for the perpendicular sections through the rock. The theory for this procedure is outlined by Ramsay (1967, p. 142-147) and is discussed more completely in Struik (1980) and Siddans (1980). Ramsay (1967, p. 145) states that the equation for determining the axes of the ellipsoid can be established by finding the nontrivial solution where the determinant

$$\begin{vmatrix} \lambda_x^{\prime} - \lambda^{\prime} & -\gamma_{xy}^{\prime} & -\gamma_{zx}^{\prime} \\ -\gamma_{xy}^{\prime} & \lambda_y^{\prime} - \lambda^{\prime} & -\gamma_{zy}^{\prime} \\ -\gamma_{zx}^{\prime} & -\gamma_{zy}^{\prime} & \lambda_z^{\prime} - \lambda^{\prime} \end{vmatrix} = 0 \qquad (1$$

the determinant can be solved using standard computer programs for solving eigenvalue-eigenvector matrix problems. The λ' are the unknown ellipsoid axes sizes and the other variables are measures of size and position of the ellipses on each perpendicular section. The knowns, $\lambda_i^{\prime}$ (i = x,y,z), are the square of the inverse of the intersections of the ellipses with the co-ordinate axes. The knowns, $\gamma_{ij}^{\prime}$ (i = x,y,z; j = x,y,z; i≠j; $\gamma_{ij}^{\prime} = \gamma_{ij}^{\prime}$), are the shear strains for each ellipse divided by the $\lambda_{ij}^{\prime}$ (i = x,y,z) value associated with the same ellipse.

Features of the ellipse associated with each of these variables is given by Ramsay (1967, p. 65-68).

Because the ellipses, recorded from the sample cuts, are sections of the same ellipsoid they should have the same values for their coincident axes' intercepts. For example the ellipse on the XY plane has a Y intercept (when X = O) and this should

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[1]Circles are included as special cases of an ellipse.

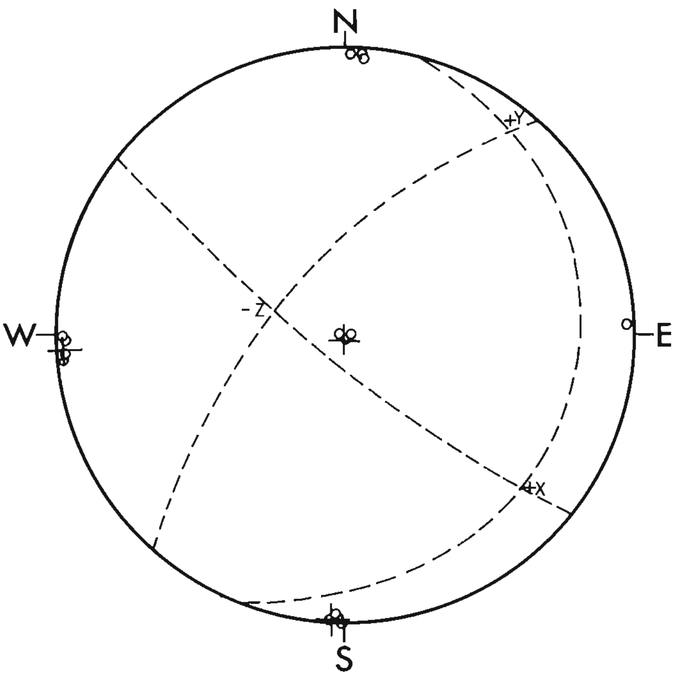

Figure 60. Equal area net plot of the axes to the ellipsoid used as a standard to which calculated strain ellipsoids for the Guyet Formation conglomerate can be compared. The dashed great circles are arbitrary perpendicular planes on which the ellipse intersections of the standard ellipsoid were graphically plotted. The X, Y and Z are the co-ordinate axes required in the determination of the position and shape of the ellipses of the three sections. The dots are the positions of six trial calculations for reconstruction of the standard ellipsoid. The crosses are the position of the average trend of the three ellipsoid axes as determined from the six trials. The parameters of the reconstructed ellipsoid are given in Table 17 under the label 'Standard'.

equal the Y intercept (when $Z=O$) for the ellipse on the YZ plane. The strain ellipses from each section of a rock can be standardized so that they intersect each other at the same place on the co-ordinate axes. The measured ellipses, however, are only approximations to the true strain ellipses and therefore there will be a discrepancy in the determination of one of the co-ordinant intercepts. The magnitude of this discrepancy would be an indication of how good the ellipses are at representing a single ellipsoid.

Because the disparity in the intercept is bound to exist for most cases, there will also be an error in the determination of the ellipsoid. There are three orders in which to calculate variables for the three ellipses and each one yields two possible intercepts for one of the co-ordinate axes. Therefore, because each of these orders of calculation and dual possibilities of intercepts will give a slightly different ellipsoid solution, there will be six possible ellipsoid determinations, none of which will be correct. The best possible solution for an ellipsoid to fit the three strain ellipses of the cut sections will be close to each of these six possible solutions. To determine one possible ellipsoid, the six solutions are averaged.

To average the six possible ellipsoids, the directions and magnitudes of the axes are considered separately. The magnitudes of the six long axes are arithmetically averaged giving the long axis of the average ellipsoid. The same is done for the short and intermediate axes. A standard deviation is determined for each average axis size and is used as a goodness of fit indicator (GOFI) of the two-dimensional strain ellipses to the three-dimensional strain ellipsoid. The standard deviation for the shortest axis is zero because all the short axis lengths are normalized to one.

The mean direction of the six long axes is the eigenvector determined from the eigenvalue-eigenvector technique (Shuenmeyer, Kock and Link, 1972) and is the long axis direction of the average ellipsoid. The short and intermediate axes are calculated similarly. The computer programme devised for this study (SOID) uses the eigenvalue-eigenvector subroutine supplied by Davis (1973, p. 166-167). Eigenvalues calculated with each of the three eingenvectors (mean axis directions) provide a GOFI. The greater the ratio between the first and second eigenvalue, the closer the six axes are to the mean direction.

Data collection

Guyet Formation conglomerate contains a variety of clast types. Because of this polymictic nature, no simple relationship exists between total strain and the geometric strain displayed by any individual grain type. Clasts were measured indiscriminantly to avoid possible bias introduced by measuring only one clast composition, size and/or shape. Clast compositions were recorded with their axial ratio determinations so that composite and specific strain determination could be performed. Comparisons between the determinations for the various clast types within the same rock were made to establish trends of differences of behaviour between the clast types.

The total number of clasts measured for each section varied. Attempts were made to collect at least 60 measurements per section. This number was sufficient in most cases to satisfy Elliott's (1970, p. 2223) criteria of enough measurements such that "a reproducible and clearly defined pattern appears". For example, a minimum of 15 measurements was found adequate to define the best strain estimate for one clast type from one of the rocks.

Clast measurements were made directly on oriented rock slabs where the grain size was sufficiently large and from oriented thin sections where the maximum grain size was granule or finer. Thin sections were used in most cases because of the bimodality in grain size. The quartz grains which are generally finer could not be measured with sufficient accuracy on rock slabs which did allow measurement of many larger rock fragment types.

Measurements from the slabs were taken with a protractor and a millimetre scale in 0.5 mm divisions. The measurements from the thin sections were taken while point counting with a mechanical stage.

Axial lengths were recorded with a calibrated ocular and the angles were measured with the polarizing microscope's rotating stage. During recording of these data observations were made on deformational features of the clasts and matrix, such as, strain lamellae, recrystallization, development of subgrains, pressure solution effects, optical extinction characteristics and boundary offsets.

Results

The results of the strain analysis are tabulated in Table 17. The dimensions of the strain ellipsoid are recorded with the small-

Table 17. Components of strain ellipsoid determinations of Guyet Formation conglomerate.

SAMPLE NUMBER	RELATIVE AXES SIZES	STANDARD DEVIATION OF AXES	AXES ATTITUDES	EIGENVALUES 1	EIGENVALUES 2	ERROR IN CO-ORDINATE INTERCEPTS	COMMENTS ON THE DESTRAINING: SECTION CLASTS USED	DISTRIBUTION	AXIAL RATIO
$ 1	1		264/39	52	21	Z- .111	I all	deltoid	2.18
	2.05	.17	050/46	21	16	Y- .574	II all	deltoid	2.95
	3.39	.36	158/18	50	16	X- .345	III all	random	1.65
$ 2	1		064/01	35	1	Z- .184	I all	random	1.0
	1.08	.10	331/86	10^6	1	Y- .221	II all	random	2.46
	2.81	.28	154/07	10^6	10^6	X- .492	III rock fragments	random	2.72
2	1		060/39	38	2	Z- .230	I all	random	1.08
	1.13	.07	251/51	62	2	Y- .265	II all	random	2.46
	2.89	.33	155/15	416	34	X- .627	III rock fragments	random	2.72
2	1		245/12	42	1	Z- .149	I all	random	1.0
	1.06	.07	040/77	308	1	Z-1.720	II all	random	2.46
	2.73	.21	155/06	377	47	X- .383	III quartz & chert	random	2.61
2	1		065/48	27	2	Z- .252	I all	random	1.0
	1.12	.15	242/41	39	2	Y- .332	II all	random	2.46
	3.12	.43	155/01	39	30	X- .736	III rock fragments	deltoid	3.07
$ 3	1		206/20	127	12	Z- .109	I all	random	1.27
	1.22	.04	026/70	16	11	Y- .083	II all	random	1.38
	1.49	.06	296/00	60	13	X- .138	III all	random	1.27
$ 4	1		232/38	332	24	Z- .101	I all	random	1.65
	1.60	.05	079/50	34	24	Y- .194	II all	random	2.01
	2.39	.09	333/14	142	34	X- .071	III all	random	1.82
4	1		132/80	353,706	28	Z- .195	I all	deltoid	1.0
	1.10	.03	135/02	10^6	347,529	Y- .487	II all	random	2.46
	2.56	.04	326/10	10^6	28	X- .245	III all	deltoid	2.72
$ 5	1		238/11	134	12	Z- .071	I all	random	1.11
	1.10	.03	355/66	22	11	Y- .082	II all	random	1.29
	1.32	.04	143/22	54	19	X- .060	III all	random	1.17
$ 6	1		212/42	47	21	Z- .270	I all	random	1.50
	1.35	.12	010/50	27	5	Y- .326	II all	random	1.50
	1.69	.15	111/07	40	5	X- .153	III all	random	1.27
6	1		219/38	3897	100	X- .110	I chert	random	1.65
	1.60	.06	342/40	113	58	Y- .162	II chert	random	2.23
	2.32	.08	103/27	1376	60	X- .045	III chert	random	1.49
6	1		208/36	58	12	Z- .213	I quartz	random	1.41
	1.27	.09	016/57	14	5	Y- .247	II quartz	random	1.35
	1.54	.11	114/03	32	5	X- .138	III quartz	random	1.22

est axis normalized to 1. Each mean relative axis size has a standard deviation. The standard deviation for the shortest axis is zero because all the short axes lengths were normalized to 1. The smaller the standard deviation, the closer the ellipses are to intersecting each other on the co-ordinate system and to representing an ellipsoid.

Each mean attitude of the strain ellipsoid is given with its associated eigenvalues. The difference in magnitude of these eigenvalues indicates how scattered the six axes trial attitudes are. The greater the ratio between the first and second eigenvalue, the better the ellipsoid represents the 3 ellipses derived from the sections. The third eigenvalue is equal to 1 due to normalization.

The errors in the co-ordinate intercepts for each of the 3 determination (Δ X, Δ Y and Δ Z) are given as a GOFI of the combination of ellipses to an ellipsoid. The smaller the values, the greater the probability that the ellipses from the three sections represent the determined ellipsoid.

A strain ellipsoid determination standard is listed in Table 17 to which the GOFI can be compared. The standard was made from an ellipsoid of known orientation and size from which its ellipse intersections with three prependicular planes were measured graphically. The only errors introduced in the determinations of these ellipses are those from the graphic determinations. The variation in orientation of the ellipsoid axes are plotted in Figure 60 to show the relationship between the scatter of orientation and the eigenvalue GOFI. The calculated direction of the intermediate axis and magnitudes of the axes show the greatest deviation from the original ellipsoid.

Several of the strain determinations show GOFIs compara-

Table 17. (cont.)

SAMPLE NUMBER	RELATIVE AXES SIZES	STANDARD DEVIATION OF AXES	AXES ATTITUDES	EIGENVALUES 1	EIGENVALUES 2	ERROR IN CO-ORDINATE INTERCEPTS	COMMENTS ON THE DESTRAINING: SECTION CLASTS USED	DISTRIBUTION	AXIAL RATIO
$ 7	1		227/05	1981	22	Z- .381	I all	random	1.75
	1.68	.11	109/50	63	22	Y- .228	II all	random	2.32
	2.50	.19	320/38	622	60	X- .009	III all	random	1.65
$ 8	1		252/22	373	98	Z- .237	I all	random	1.38
	1.46	.12	096/70	478	6	Y- .107	II all	deltoid	1.62
	1.54	.12	348/06	101	6	X- .237	III all	random	1.00
8	1		252/22	799	281	Z- .084	I all	random	1.38
	1.41	.04	088/70	1270	5	Y- .042	II all	random	1.46
	1.44	.04	348/05	297	5	X- .844	III all	random	1.00
$ 9	1		197/02	19	2	Z- .212	I all	random	1.50
	1.11	.06	106/30	12	2	Y- 168	II all	random	1.35
	1.47	.09	291/60	20	11	X- .151	III all	random	1.32
9	1		202/09	196	16	Z- .044	I chert	random	1.11
	1.10	.01	108/24	78	16	Y- .034	II chert	random	1.35
	1.43	.02	309/64	149	71	X- .030	III chert	random	1.32
9	1		086/40	29	9	Z- .059	I quartz	random	1.00
	1.09	.01	192/19	74	8	Y- .057	II quartz	random	1.20
	1.38	.02	305/44	41	26	X- .057	III quartz	random	1.32
$10	1		051/26	58	26	Z- .090	I all	random	1.75
	1.44	.03	186/56	35	22	Y- .112	II all	random	1.38
	1.91	.05	306/21	60	25	X- .061	III all	random	1.46
10	1		050/29	64	33	Z- .098	I chert	random	1.75
	1.73	.04	173/45	34	13	Y- .062	II chert	random	1.75
	2.12	.06	300/32	66	13	X- .128	III chert	random	1.58
10	1		057/33	138	53	Z- .026	I quartz	random	1.49
	1.30	.01	194/29	100	48	Y- .022	II quartz	random	1.25
	1.68	.01	312/22	156	75	X- .032	III quartz	random	1.43
$11	1		202/22	18	5	Z- .096	I all	random	1.00
	1.05	.04	294/05	66	5	Y- .087	II all	random	1.27
	1.26	.04	033/68	29	17	X- .080	III all	random	1.15
Standard	1		102/89	206	52	Z- .038	True axial ratio 1-1.5-2.5		
	1 511	.03	267/02	54	39	Y- .097	True orientation vertical-270/00-180/00		
	2.518	.07	180/00	163	39	X- .081			

$ This determination is chosen as representative of the strain for that sample

ble to the test case. These are samples 5, 9 (trial 2), and 10 (trial 3). Within the suite of rocks examined these have moderate to low strains.

The GOFIs give an estimate of the geometric fit of the determined ellipses to an ellipsoid but do not necessarily provide an estimate of how good the ellipsoid approximates the strain ellipsoid.

Deviations from geometric deformation

Deviations from the ideal case of geometric strain are in part a result of irregular grain shape, viscosity contrasts and clast composition.

Grain shape

The strain measurement technique requires that markers (clasts) be treated as ellipsoids. The sections through the clasts are then treated as ellipses. The clasts measured were never truly elliptical. They exhibit a good to very poor fit to ellipses. Many of the clasts of all compositions are angular within the Guyet Formation conglomerate. Most angular are the chert and pelite clasts. Volcanic fragments are the most rounded and they are the best approximations to ellipses, but are not abundant. Most abundant are sand size quartz grains which are subangular to rounded. Larger quartz grains are generally more angular. Chert grains which are very flat and elongate display a moderately good fit to elliptic shapes. In general, the greater the deformation, the easier it is to fit an ellipse to the clasts.

An irregularly shaped clast will not follow the deformation path of an ellipse. Deformation will be accommodated in irregular and unstable parts of the grain. This can be thought of as attainment of a shape most amenable to straining. This concept is discussed further in the section on "Clast composition and mechanisms of deformation", in relation to the strain behaviour of quartz and chert. Interaction or noninteraction with neighbouring clasts will partly govern how irregularities affect the strained shape. Mukhophadhyay (1973) argued that quartz grains in the Ardennes are good markers because they are separated enough not to interact with each other. This is intuitively reasonable. Markers used here, however, all show some interaction.

Impingement of grains upon each other produces elongation of the grains or one of the grains. Elongation is always in the plane of stylolitization which corresponds with the direction of secondary mica growth (ie. the cleavage plane). Grain impingement causes deviation from the strain path of an ellipse by imposing additional grain shape changes. These grain interaction effects would be small if there were no grain boundary sliding, or viscosity contrasts, because bounding areas would deform with the clasts but would always retain their configuration. Two touching clasts would remain just touching throughout the deformation. Within the studied suite of rocks viscosities differ between the clasts, and there is probably grain boundary sliding. Grain interactions do occur and, along with the initially irregular grain shapes, cause final strain marker distributions to have features which are not predicted by geometric deformation.

Viscosity contrasts

Clasts of all compositions were measured and included together in most of the strain determinations.

For discussion, let us assume a perfectly random distribution of heterogeneous clasts. Clasts that are more viscous than the matrix will display a component of rigid body rotation when strained (Gay, 1968a, p. 219). Gay (1968a) pointed out that clasts, with viscosities much greater than the matrix, will exhibit little shape change, but will undergo rotation such that their long axis will coincide with that of the strain ellipse. Clasts, which are nearly equivalent to the matrix in viscosity, will deform geometrically and not rotate rigidly, thereby approximating more closely the deformation predicted by geometric strain analyses.

Clasts with moderate to low viscosity ratio will approximate more closely the deformation predicted by geometric analysis of the strain. Their ellipticities will generally be greater than for the high viscosity ratio clasts. On the polar plot the distribution in a strained state would be directionally uniform, but display a spread in magnitude of final ellipse ratios. This would have the effect of inducing bimodality in the final ellipse distribution on the polar plot, if the strain ellipse were taken to be the central part of the distribution. A characteristic of this type of induced biomodality is that the strain ellipse axes will parallel the two directional modes of the unstrained distribution. Gay (1968a) assumed that the boundaries between matrix and clast are fixed (ie. there is no creep or slippage). In the presence of pressure solution phenomenon this assumption would not hold.

A consequence of this assumption being invalid is that the rigid body rotation component of competent clasts need not occur. Gay (1968a) pointed out that clast interference has a tendency to lessen the control of viscosity contrast and therefore presumably decrease the effects of rotation.

Figure 61 illustrates bimodality of a distribution produced by viscosity contrast between clast and matrix. The distribution tends to be bimodal where only quartz clasts are involved and even more accentuated where less viscous clasts are also present.

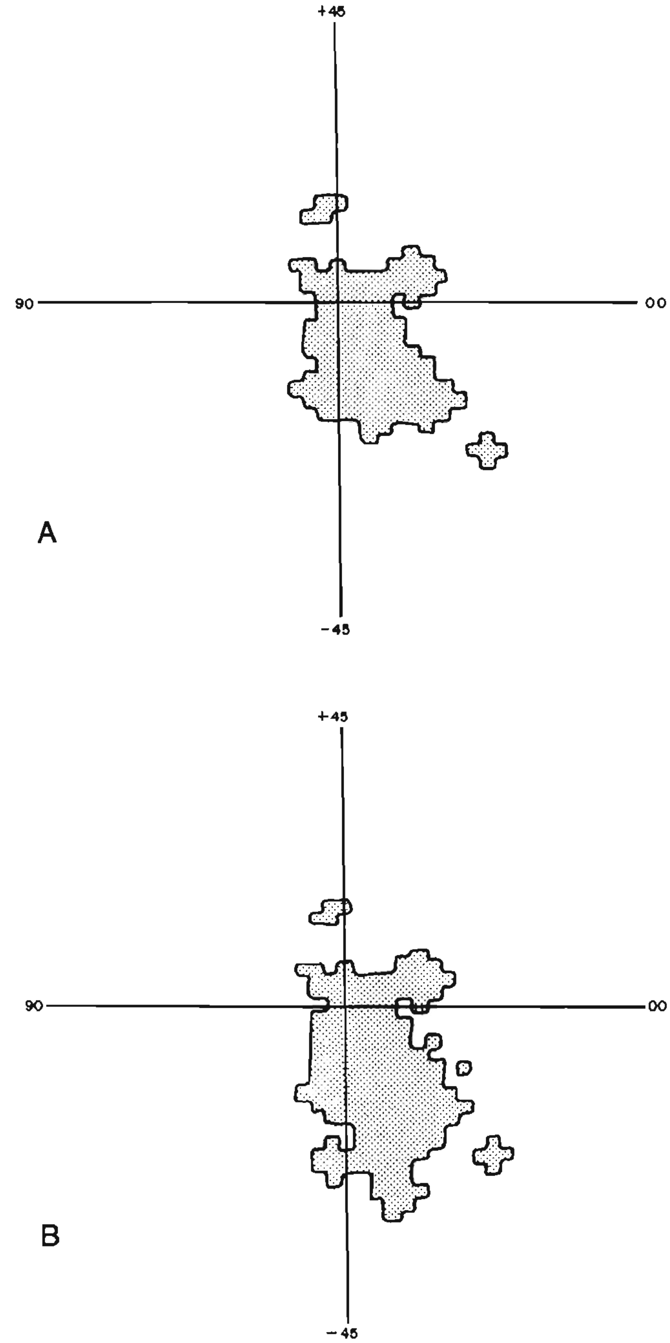

Figure 61. These are polar plots of the natural log of the axial ratio and orientation of the long axis of the final ellipses from sample 1. Plot A contains data from quartz grains and plot B combines these with data from chert grains. The difference in dispersion of the pattern from the origin is because the chert shows more deformation than the quartz. These differences are due to viscosity contrasts between the grains, and between the grains and the matrix.

The bimodal characteristic of initial clast-long-axis directions for the suite of rocks examined in this study cannot be demonstrated as a characteristic of deposition. It is more likely a phenomenon induced by applying to heterogeneous material, the geometric strain theory for homogeneous material.

It is demonstrated from Figure 61 and from the previous discussion that clasts show varying degrees of strain and that this variation is partly related to viscosity differences. Gay (1968a) outlined a method by which approximations of the strain can be determined by knowing the clast concentration and the viscosity ratio between clast and matrix. In a demonstration of his technique Gay (1968b, p. 301) uses a chert clast-shale matrix combination which is approximated by the muddy conglomerate of the Guyet Formation. He showed that the chert clasts will act as the best strain markers. For a conglomerate with a quartz sand matrix the chert grains will overestimate the strain of the whole rock. Pelite clasts within the same rock will overestimate the whole rock strain even more.

Viscosity differences may compensate for each other in the Guyet conglomerate, so that the strain determinations are reasonable estimates when the initial sedimentary distributions are considered random or nearly so. For the muddy conglomerate, the chert particles will form the core of an induced bimodal distribution. By assuming random initial distributions, the core would be chosen to represent the strain ellipsoid. This is satisfactory because the chert would represent the best strain marker in this case. For the sandy conglomerate some combination between quartz and chert would best represent the strain ellipsoid. Because of the abundance of quartz in these rocks, the final distribution will be governed primarily by the quartz. Therefore choosing the strain ellipsoid to be represented by the core of such a distribution would lead to the best approximation of the true strain ellipsoid.

For the studied samples, the effects of viscosity contrast are accounted for by treating the final distributions as random and including all the grain types. This was found impractical for only one rock (1) which contained predominantly chert and cherty rock fragments in a shale matrix. The final distribution for this rock appeared to be bimodal, and this could not be related to clast differences. Treating this rock as having a truly bimodal initial distribution of grain shapes is reasonably valid, because the clasts have similar viscosities.

Clast composition and mechanisms of deformation

Recent writings in tectonics allow for four mechanisms of producing strain (*see* Elliott, 1973; Nicholas and Poirier, 1976; Rutter, 1976; White, 1976, 1977; Borradaile, 1979; and Etheridge and Wilkie, 1979 for reviews and current hypotheses). Of these four mechanisms, only three appear to be accepted by all. These are dislocation creep and glide, diffusion creep (includes Coble creep and Nabarro-Herring creep) and grain boundary sliding. Pressure solution is commonly accepted though it is in some cases equated with diffusion creep (*see* Elliott, 1973).

Elliott (1973) and Etheridge and Wilkie (1979) implied that grain boundary sliding must accompany diffusion or dislocation flow. Borradaile (1979), however, suggested a grain boundary sliding which involves clastic particles that move past each other without being themselves deformed. This process he called independent grain boundary sliding. The grain boundary sliding of Elliott (1973) and others was called dependent grain boundary sliding by Borradaile (1979). An intermediate style was referred to as controlled grain boundary sliding (*see* Borradaile, 1979, p. 87).

Deformation characteristics of the four major clast types from the studied rock suite are now described and related to these four deformation mechanisms. Assumptions required for the geometric strain analysis technique are discussed in relation to the observed deformation features.

Quartz. Most of the old quartz grains (those not recrystallized or crystallized during deformation) have undulatory extinction. White (1976) attributed undulatory extinction to concentrations of strain dislocations causing lattice distortion. This suggests dislocation creep as one of the strain mechanisms affecting quartz grains. Other evidence for dislocation creep according to White (1976) is the presence of strain lamellae and bands. Lamellae were only rarely seen. Sutherland Brown (1957) reported lamellae and strain shadows in quartz from the Guyet Formation and considered the phenomenon to be inherited from the source rock of the clasts. Strain features in the quartz clasts of Guyet Formation conglomerate can be accounted for by deformation after Guyet deposition.

Quartz subgrain development can be attributed to two causes, 1) precipitation of quartz in strain shadows due to diffusion flow, and 2) recovery-recrystallization in response to dislocation overloads. Both of these forms of subgrain development have been observed. Figure 62 shows type 1 development. Quartz fibre and granule growth at the ends of grains are oriented in the maximum elongation direction. Because this orientation occurs consistently for many grains the subgrain development is part of the strain of the Guyet Formation. This can also be said for subgrains produced as in Figure 63 by dislocation overloads (crushing). Consistently the elongation of the subgrains and the larger crosscutting boundaries (cracks, fractures) are parallel to the maximum elongation direction of the Guyet strain, or can be directly related to that strain direction.

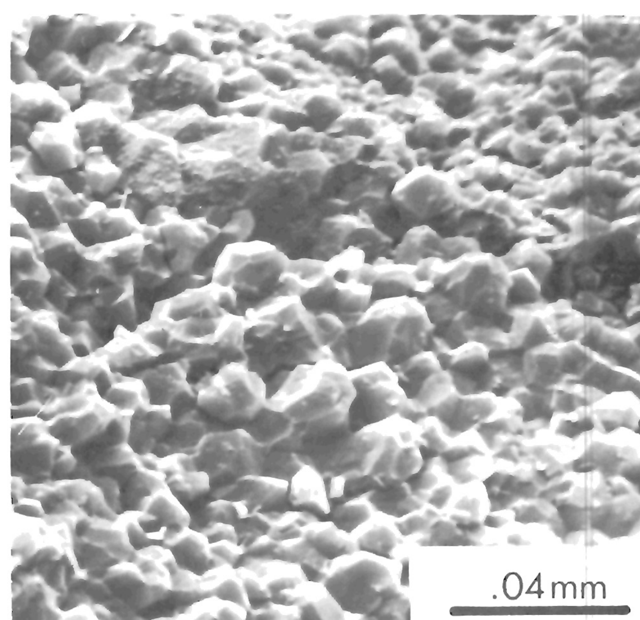

Figure 62. Scanning electron micrograph of a deformed chert clast within the Guyet Formation conglomerate. Compare this to the micaeous chert shown in Figure 65.

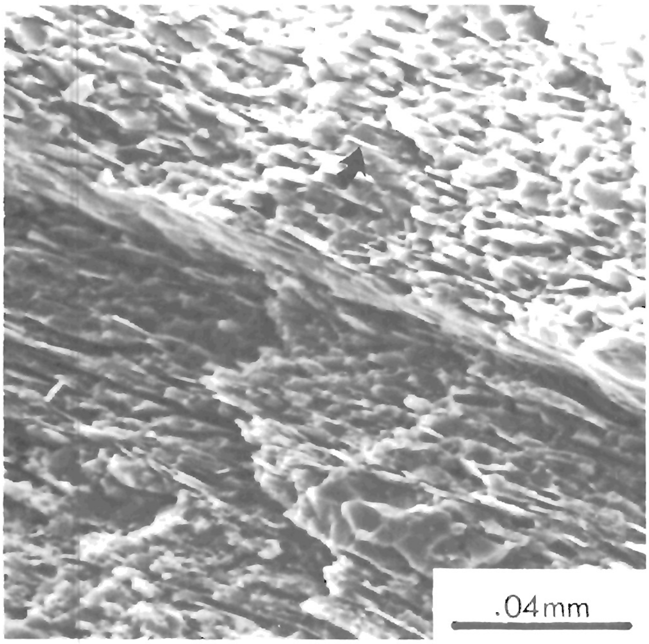

Figure 63. Scanning electron micrograph of a deformed micaceous chert grain from Guyet Formation conglomerate. Note that the individual microcrystals of the chert grain are elongate between the mica plates (sample of which is noted by arrow) within the chert. Compare this to the clean chert shown in Figure 64.

Some quartz grains were ruptured due to extension parallel to the cleavage trace. Intervening spaces were filled with recrystalline fibrous and granular subgrains of quartz. The size of the subgrains varies from 0.01 mm to 0.25 mm.

All quartz grains were measured. Grains that showed subgrain development were measured so as to include the subgrains. Grains that have changed shape by solution and redeposition upon the same grain will represent reasonable approximations to geometric strain. Pressure solutions may transport material from a grain or grains to deposit it in the pressure shadow of another grain. This may cause the strain for that grain to be overestimated, however, the area which donated the material will be underestimated and a partial balance will occur.

Crushed quartz grains with their induced shape irregularities are not good geometric strain markers. The error introduced, however, by including them in the measurements can be no more than the errors associated with controlled grain boundary sliding. Grain boundary sliding is suspected to have occurred in the muddy conglomerate of the Guyet Formation. Compaction by pressure solution shortening and general dewatering of the clay matrix will have rotated quartz (and other more viscous) grains towards the plane of flattening. This rotation would be grain boundary sliding as opposed to the rotation due to viscosity contrasts as outlined by Gay (1968a). Rotations due to grain boundary sliding will alter the relationship between the initial and final distribution, required as part of the Elliott (1970) strain technique.

Chert. Although also composed of quartz, chert grains display a behaviour markedly different from grains of coarsely crystalline quartz. Because of the small size of the quartz crystals composing the chert grains it is not known whether the individual crystals exhibit undulatory extinction. Dislocation creep therefore is hard to establish or refute with the aid of an optical microscope.

Chert grains show a flow-like behaviour, particularly where quartz has impinged upon chert grains. The quartz grains become partly enveloped by the chert. A difference between chert and quartz grains is the high crystal boundary area of the chert. To account for the behaviour of the chert it is postulated that the high boundary area encourages diffusion flow. This concept was discussed by Etheridge and Wilkie (1979) in relation to mylonites. Grain size reduction is argued to enhance superplastic behaviour by increasing grain boundary area and promoting diffusion flow and/or dislocation creep with dependent grain boundary sliding (*see* also White, 1977).

Deformed chert (Fig. 62) from the Guyet Formation conglomerate has regular shaped individual crystals. Micaceous chert (Fig. 63) from the same rock as Figure 62 contains individual crystals that are elongate. The sheet silicates of the micaceous chert appear to have insulated the quartz crystals, permitting them to deform individually.

The regularity of the quartz crystals in deformed pure chert (Fig. 62) suggests that the individual crystals are continually recrystallizing, either internally or by mass transfer between grains. Material transported between grains is either deposited as new grains in areas opened by grain boundary sliding or becomes incorporated into the crystal structure of the preexisting grains, or both. Throughout the process, the individual crystals did not become elongate as would be expected if they were deforming by the grain boundary diffusion and flow method outlined by Elliott (1973). His model involves shortening of grains parallel to the compression direction and elongation perpendicular to compression. Elongation takes place by deposition of material removed from areas of the grain under compression. This process may be operating in the micaceous cherts, where individual quartz crystals are elongated preferentially in the maximum extension direction.

Impure cherts with varying contents of mica exhibit stylolite development under pinching conditions. The stylolites are marked by concentrations of sheet silicates and minor fine opaque matter.

Quartz grains that exhibit crushing phenomena at their edges may be promoting similar behaviour as described for clean cherts. The finer crushed fragments are forming more boundary area and therefore, enhanced boundary diffusion. This argument suggests that within the area of a grain, both dislocation and diffusion mechanisms can occur simultaneously. Only grain size and possible impurity content would be the governing factor. As the quartz grain breaks down to form subgrains, diffusion would increase as the grain boundary area increased. This process may have operated in forming the chert-like accumulation at the end of the large grain in Figure 64. Compare this type of accumulation to that which fills pressure shadows, as in Figure 65. Pressure shadow material is characteristically elongate.

How does the diffusion process of the cherts affect its use as a geometric strain marker? They would react very much as theoretically required were it not for interference from grains of different composition. Their originally angular to subrounded shapes also do not accord well with theory, but that is less of a problem than clast interference which induces new shape changes.

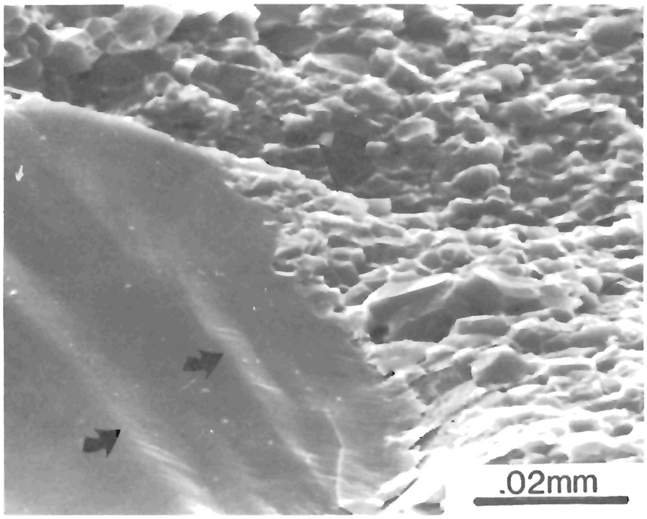

Figure 64. Scanning electron micrograph of quartz subgrain formation (large arrow) at the ends of a crushed quartz grain from Guyet Formation conglomerate. Strain lamellae are located by small arrows.

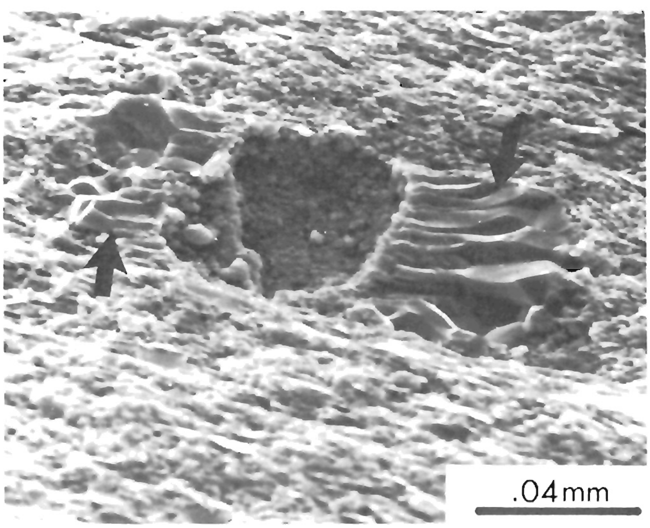

Figure 65. Scanning electron micrograph of quartz crystal fibrous growth (noted by arrows) in the pressure shadows of a quartz grain from conglomerate of the Guyet Formation. Compare this to the type of quartz deposits at the end of a recrystallizing crushed quartz grain as shown in Figure 63.

Pelite. The deformation mechanism observed in pelite is pressure solution. In rocks of low strain, where the preferred orientation of sericite is not yet prominent, the pelites have flowed into the interstices of surrounding grains (*see* Sutherland Brown, 1957, p. 34) but have not yet developed stylolitic striping. Local areas of pelites in such rocks do have this striping due to the impingement of other grains. The tendency for pelite grains to enter available spaces introduces irregularities to its form which are unaccounted for by the geometric strain technique. For rocks with low strain the irregularities are no worse in inducing errors than those previously described for chert. Problems do arise for those rocks that have undergone moderate to substantial amounts of strain. This is at the onset of the preferential orientation of micas.

Pelite clasts in these rocks show substantial amounts of pressure solution as shown in Figure 62 where quartz stringers are displaced along stylolites. Gray (1978) described similar dissolution of quartz veins and made a good case for this phenomenon not being lateral offset of the quartz vein. The pelite clast in Figure 62 was shortened perpendicular to the stylolites by approximately 60%. The material removed to produce this amount of shortening is not traceable. The elliptic shape ascribed to this pelite clast would be a bare minimum estimate of the deformation involved.

Intense deformation of pelite clasts results in their complete conversion to stylolitic residue consisting of fine dusty opaques and well oriented micas. Material of this nature is useless as an elliptic strain marker. In many cases it is undifferentiable from matrix pelite that has undergone the same process; also being reduced to stylolitic stripes.

Volcanic. Volcanic clasts are found in Guyet Formation muddy conglomerate. Deformational characteristics of these clasts are governed by mineral content and the size and distribution of these components within the clast. Feldspar crystals are commonly layered and show straight boundaries. Deformational twins are rare. Feldspar laths are aligned preferentially with the cleavage direction. Fine dissolution surfaces are locally highlighted by quartz stringers that terminate against and are offset by them.

The predominant deformation mechanism is reasoned to be grain boundary sliding of competent feldspar laths in a matrix of sheet silicates (predominantly chlorite and sericite) that are recrystallizing preferentially into the cleavage plane. This is accompanied by development of pressure solution stylolites.

The grain shapes are subrounded and therefore are moderately good approximations to ellipsoidal markers. Interference from other clasts has caused shape distortions and impeded the volcanic grains from closely following the geometric deformation path.

Summary

Final distribution patterns for the samples investigated do not represent an initial distribution which has been deformed along the path prescribed by the Elliott (1970) shape-factor grid. The deviations from geometric strain will have altered the path of deformation for each marker grain. A strain ellipsoid that best destrains the distribution geometrically will not be the true strain ellipsoid. It will be the strain ellipsoid which best forces the distribution to follow a geometric strain path into a preconceived initial distribution.

Because of this ambiguity the final distributions were assumed, in most cases, to represent initially random distributions. They may not have been and probably were not originally random but this is not unambiguously determinable. Further arguments for using the final distributions as if they represented random initial distributions was given by Seymour and Boulter (1979). They tested two strain determination methods which relied on knowledge of the initial distribution of markers. Mathematically strained distributions of real undeformed rocks were destrained using these two methods. Errors for these determinations ranged from 10 to 91% depending on the symmetry of the distribution about the bedding trace.

These errors exist for cases which are perfectly geometrically deformed. The bedding trace plays an important role in these determinations. Bedding traces are nonexistent for much

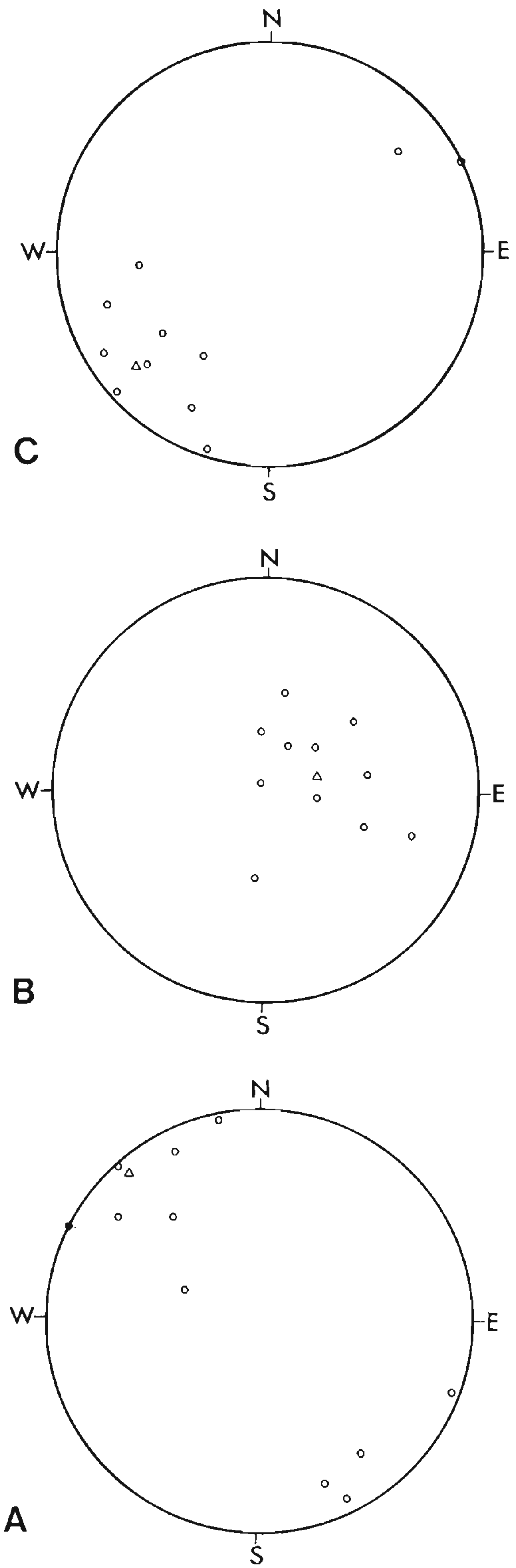

Figure 66. Equal area net plots of the trends of the three axes of the determined strain ellipsoids for the 11 samples from the Guyet Formation conglomerate. The average strain ellipsoid orientation determined from the 11 samples is represented by the triangle. A) long axis, B) intermediate axis and C) short axis.

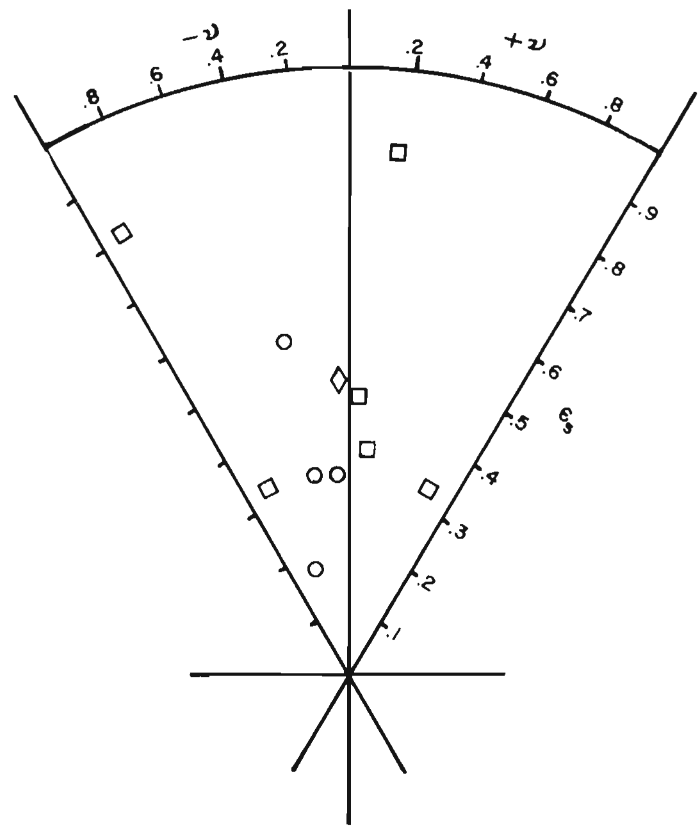

Figure 67. Nadai deformation plot showing the magnitude (epsilon) and shape (lambda) of the strain ellipsoid for 11 samples of Guyet Formation conglomerate. The squares are muddy conglomerates and the circles are sandy conglomerates. The average strain ellipsoid is marked by the diamond. Points plotting in the negative field are in constriction (the strain ellipsoid is shaped like a cigar), and those in the positive field are in flattening (the strain ellipsoid is shaped like a pancake). Any volume loss, as would normally be the case with the compaction of wet marine sediments, would shift the field of flattening to the left, such that the average strain ellipsoid would plot on or just to the right of the plane strain line (the division between constriction and flattening).

of the Guyet Formation conglomerate. Any errors accumulated from analyzing nongeometrically deformed markers must be added to those that Seymour and Boulter (1979) attributed to the strain analysis.

Assuming a random initial distribution can only introduce small percentage errors for any but low strains. Assuming non-random distributions can only introduce larger possible errors by misinterpretation of distribution patterns and the inherent error in the final distribution.

Strain orientation and magnitude

The strain ellipsoids which were chosen to represent a conservative strain estimate for each rock are marked in Table 17. The axes of these strain ellipsoids are plotted on equal area nets in Figure 66. The mean axes' directions are also plotted.

The average long axes of the strain ellipsoid (X, maximum extension direction) closely parallels the mean orientation of the ductile fold axes of the Cariboo Terrane (Fig. 59, 66, 24). The mean maximum compression direction (Z) is toward the southwest. This agrees with the predominant eastward dip of the cleavage and yields a Z direction approximately perpendicular to the cleavage. This relationship is confirmed by hand sample and thin section examination. The intermediate axis (Y) lies in the cleavage plane.

The average deformation ellipsoid approximates plane strain (Fig. 67) assuming there has been no volume change. Any volume decrease, resulting from compaction of the rock, would increase the field of constriction and include the average strain ellipsoid and some of those in the apparent flattening field. The distribution of strain forms is not related to rock type (Fig. 67) and does not follow an obvious geographic pattern. Tobisch et al. (1977) summarized average strain ellipsoids for some large orogenic belts, and all but two of these fall within the apparent field of flattening.

The elongation in the fold axis direction for the mean strain ellipsoid is 42.5%, assuming no volume change during deformation. The elongation is accompanied by a 28.8% shortening perpendicular to the eastward-dipping cleavage and a 1.7% shortening in the dip direction of the cleavage. Cloos (1947) suggested that elongation of this sort in the Appalachian fold belt could be attributed to arcing of the orogenic front. Any arcing present for the Barkerville area is insufficient to account for a 42% elongation parallel to the orogen. No other structures in the area take up the elongation by compression perpendicular to the orogen. The strain ellipsoid is thought to record a component of shear that parallels the orogenic trend.

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