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LEAD-ZINC AND COPPER-ZINC DEPOSITS IN SOUTHEASTERN BRITISH COLUMBIA

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GEOLOGICAL FRAMEWORK

Trygve Höy

Introduction

A large lead-zinc metallogenic province extends from northern Idaho and Montana through southeastern British Columbia to north of Revelstoke in the Selkirk Mountains. Included in this province are stratabound and disconformable base metal deposits in both clastic and carbonate rocks that range in age from Proterozoic to Devonian. This excursion will visit the operating Sullivan mine, the mined-out St. Eugene and Bluebell deposits, and the Duncan and Goldstream prospects. Emphasis will be placed on the stratigraphic, structural and tectonic setting and controls of these deposits.

Tectonic and Structural Setting

The lead-zinc province is within the Columbian Orogen, the eastern of two orogens in the Canadian Cordillera (Wheeler, *et al.*, 1972). The Columbian Orogen comprises a central core zone of variably deformed and metamorphosed sedimentary and volcanic rocks mainly in the Omineca Crystalline Belt, an eastern belt called the Foreland Thrust and Fold Belt that consists dominantly of miogeoclinal sedimentary rocks thrust northeastward, and a western belt in which the dominant tectonic transport was southwesterly (Wheeler, *et al.*, op. cit.). The excursion commences in the Purcell Anticlinorium in the western part of the Thrust and Fold Belt, crosses into the Kootenay Arc on the eastern edge of the Omineca Crystalline Belt on Day 2, and follows the trend of the Arc and the eastern edge of the Omineca Belt northward for the remainder of the trip (Fig. 1). Formations through which the

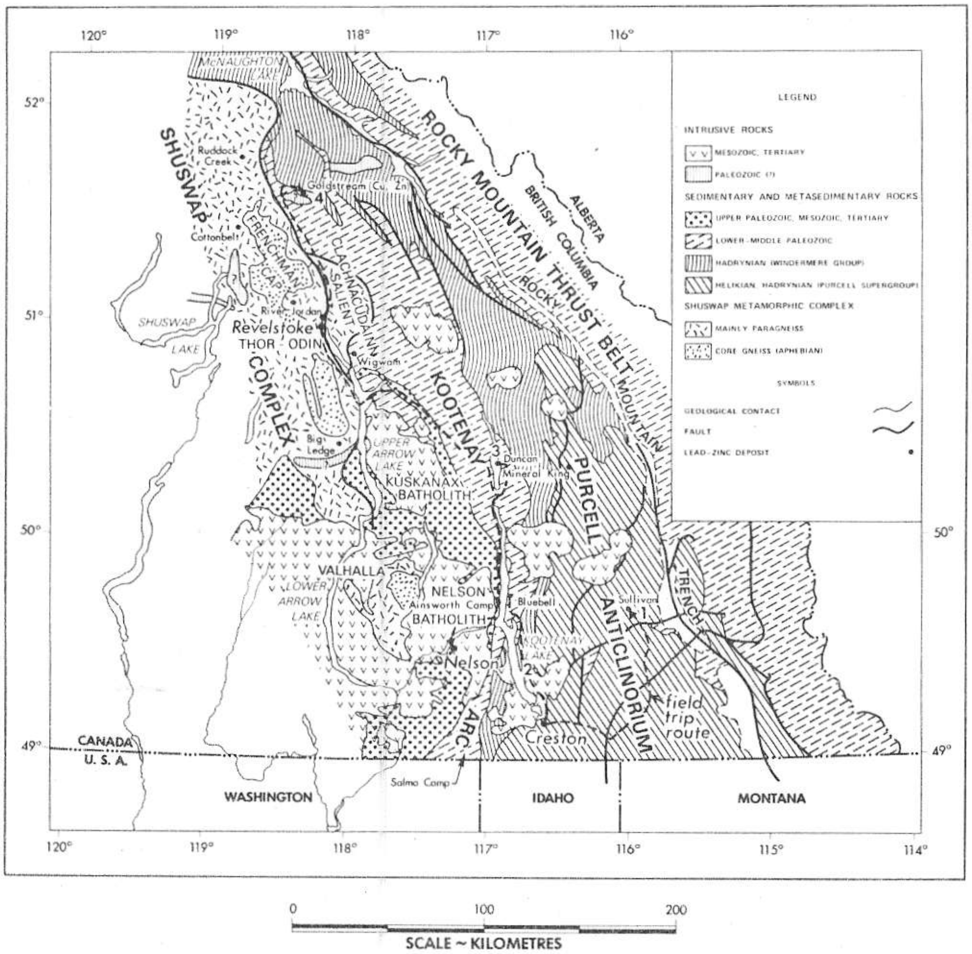

Figure 1. Geological map of southeastern British Columbia showing field trip route.

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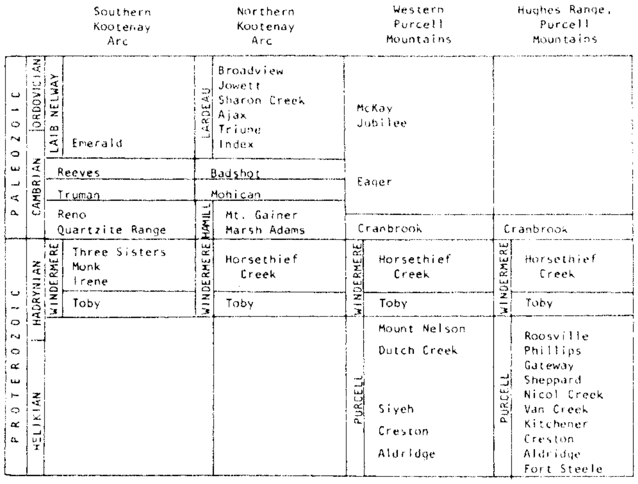

Figure 2. Correlation chart.

excursion passes and their correlation are illustrated in Figure 2.

The Purcell Anticlinorium is a broad north-plunging structure in Helikian and Hadrynian age rocks between the Rocky Mountains to the east and the Kootenay Arc to the west. It is transected by a number of steep longitudinal and transverse faults (Fig. 1). The transverse faults appear to have been active intermittently since at least Hadrynian time and played an important role in controlling the type, distribution and thickness of late Proterozoic and early Paleozoic sediments (Lis and Price, 1976). The Sullivan ore body (Fig. 1) occurs within Helikian clastic rocks on the eastern flank of the Purcell Anticlinorium.

The Kootenay Arc (Fig. 1) is a north-trending arcuate structural zone that developed in a succession of rocks ranging in age from Hadrynian to early Mesozoic. The Bluebell deposits at Riondel are replacement deposits in lower Cambrian Badshot Formation marble in the central part of the Arc, and the Duncan deposit to the north comprises a number of sulphide zones in deformed, dolomotized Badshot marble.

The Goldstream area is near the eastern edge of the Omineca Crystalline Belt, separated from the Shuswap Metamorphic Complex to the west by the Columbia River fault (Fig. 1). The area is dominated by tight, generally north-trending folds with horizontal to steeply east-dipping axial surfaces that developed in a late Proterozoic to lower Paleozoic succession of pelitic and calcareous schist, quartzite, marble and basic metavolcanic rocks. Spatially associated with the basic volcanic rocks or within them are a number of massive copper-zinc sulphide deposits. One of these, the Goldstream deposit, will be visited on Day 4.

The Shuswap Metamorphic Complex is a belt of high grade and intensely deformed metamorphic rocks, intruded by granitic rocks, that comprises the core of the Columbian Orogen. Its eastern margin, in fault contact with the Kootenay Arc and the Selkirk terrain, is characterized by a series of domal structures that expose mixed paragneiss, granitic gneiss and migmatite that recently have been dated as Proterozoic Aphebian (R. L. Armstrong, U.B.C., personal communication, 1980). Core gneisses are overlain by a heterogeneous assemblage of calc-silicate gneiss, pelitic gneiss, quartzite and marble that has been dated as Helikian. This succession of paragneisses hosts a number of stratabound lead-zinc deposits (Fyles, 1970b). The deposits consist of either a single layer of massive to irregularly banded sulphides or a series of discontinuous lenses within a thin calcareous schist unit. They are folded and metamorphosed along with the country rocks. Because of the inaccessibility of the Shuswap deposits they will not be visited on this excursion.

THE PROTEROZOIC BELT PURCELL SUPERGROUP

Trygve Höy

Introduction

A number of lead-zinc deposits occur in the Proterozoic Purcell Supergroup. These include the stratabound Sullivan mine, Kootenay King and North Star deposits, and the Vulcan occurrence, in clastic rocks of the Aldridge Formation. Transgressive deposits include the Stemwinder and the St. Eugene. Replacement deposits in younger Purcell platformal carbonates include the Mineral King and Paradise deposits 80 km to the north of the Sullivan. The Sullivan mine will be visited on Day 1 of the excursion, and the Purcell succession in the vicinity of Moyie Lake 30 kilometres south of the Sullivan will be examined in the morning of Day 2.

Stratigraphic and Structural Setting

Purcell rocks are exposed in three tectonically distinct areas in southeastern British Columbia and southwestern Alberta. From east to west, these are the Clark Range within the Lewis Thrust sheet in the southeast corner of B.C. and southwest corner of Alberta, the Hughes, Lizard and Galton Ranges on the east side of the Rocky Mountain Trench, and the Purcell Mountains within the Purcell Anticlinorium. In the Clark Range, arenaceous and carbonate facies are relatively more important and formations are thinner and lithologically diverse (Price, 1964). Coarse-grained fluvial and shallow-water sediments host a number of widespread, low-grade stratabound copper occurrences in the Clark Range (Morton, *et al.*, 1973; Collins and Smith, 1977). In the Purcell Mountains, formations are thicker and rocks somewhat more argillaceous and less carbonate-rich, whereas rocks just east of the trench are transitional between the area to the west and the Clark Range to the southeast. Purcell rocks east of the Rocky Mountain Trench have been described by Rice (1937), Leech (1958 and 1960), Price (1962 and 1964), McMechan (1979) and Höy (1979a). Some of the best exposures are in the Kootenay King area. The Kootenay King is a laminated stratiform Pb-Zn deposit in siltstone correlative with the Middle Aldridge turbidites to the south and west. The Aldridge in the Kootenay King area is underlain by coarse fluvial sands of the Fort Steele Formation (Fig. 2). A succession of turbidite wacke beds is in the central part of the Middle Aldridge, stratigraphically above the Kootenay King horizon. The Middle Aldridge is overlain by dominantly shallow-water tidal flat, flood plain and deltaic deposits of the Creston Formation and platformal carbonates of the Kitchener Formation (Fig. 2). Shallow-water siltstone and quartzite of the Van Creek and andesitic volcanic rocks of the Nicol Creek Formation (McMechan, Höy and Price, in press) overlie the Kitchener east of the trench and along the southeast margin of the Purcell Anticlinorium. They in turn are overlain by shallow-water clastic shelf deposits of the Gateway, Phillips and Roosville Formations.

The oldest rocks exposed in the Purcell Anticlinorium to the west, including the area crossed on the morning of Day 2 of the excursion, are quartzites, siltstones and argillites of the Aldridge Formation (Fig. 3). The Lower Aldridge comprises at least 1500 m (base not exposed) of rusty-weathering argillite, siltstone and quartzite. The contact between the Lower and Middle Aldridge is gradational over a few tens to several hundred metres. Locally an intraformational conglomerate occurs at or near the top of the Lower Aldridge.

The Middle Aldridge comprises thick, grey quartz-wacke beds and interlayered laminated siltstone layers, intruded by a number of regionally extensive metagabbro sills (Fig. 3). In general, quartz-wacke beds become thinner, less pure and volumetrically less important higher in the Middle Aldridge section (Höy and Diakow, 1981). The quartz-wacke beds "possess sole markings, including grooves, poor flutes, and longitudinal ridges, and internal sedimentary structures of cross- and convolute bedding, that

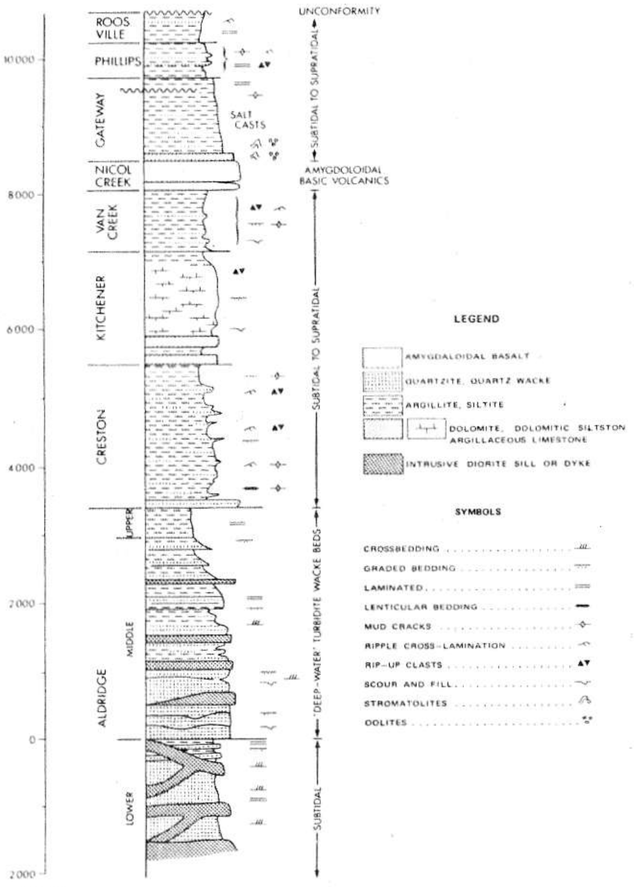

Figure 3. A composite stratigraphic column of the Purcell Supergroup in the Moyie Lake area.

suggest their similarity to units ascribed elsewhere to turbidity currents (Bishop, Morris and Edmunds, 1970). Internal stratification of these units permits recognition of the standard Bouma (Bouma, 1962) intervals, although Bouma's B, C, and D Layers do not appear to fit a regular pattern of super-position in the Aldridge Formation. Cross-bedding in the C interval indicates a basin axis running in a north-south direction through the centre of the outcrop area" (Edmunds, 1977, p. 22).

Interturbidite argillaceous siltstone in the central part of the Middle Aldridge commonly consists of a sequence several metres thick of alternating dark and light laminations less than a millimetre to a few millimetres thick. Laminations in about a dozen of these laminated siltstone sequences can be matched across distances of up to three hundred kilometres (Edmunds, 1973, 1977; Huebschmann, 1973). They are reliable stratigraphic markers that allow correlation within the 3 kilometre thick Middle Aldridge succession.

The Upper Aldridge includes 300-400 metres of generally rusty-weathering laminated dark grey argillite and lighter grey siltstone. The overlying Creston Formation consists of light green, brown and pale purple argillaceous quartzite, siltstone and argillite that contain numerous shallow-water sedimentary structures. The Creston is overlain by shallow-water carbonates and clastics of the Kitchener and "Siyeh" Formations (Fig. 3). Andesitic lavas of the overlying Nicol Creek Formation along the eastern flank of the Purcell Anticlinorium are not present throughout most of the Purcell Mountains. Rather, grey and green argillite, dolomite and quartzite of the Dutch Creek Formation (Fig. 2) overlies the "Siyeh" Formation (Reesor, 1973) and is overlain by oolitic and stromotalitic dolomite and dolomitic limestone, argillaceous limestone and argillite of the Mount Nelson Formation (Reesor, 1973; Rice, 1941).

In summary, Fort Steele fluvial sandstones in the Kootenay King area on the east side of the Rocky Mountain Trench are supplanted to the west by deep-water siltstone and occasional turbidite beds of the Lower Aldridge in the Purcell Mountains. Middle Aldridge time is marked by the introduction of extensive and thick accumulations of turbidite deposits. Locally, intraformational conglomerate, boron alteration zones (Ethier and Campbell, 1977) and sulphide accumulations developed. The basin of turbidite deposition expanded and, during Middle Aldridge time, overlapped "inner fan" deposits in the Kootenay King area. Post-Aldridge rocks record deposition in an extensive, generally shallow-water marine platform environment.

Lead-Zinc Deposits in the Purcell Supergroup

Clastic-hosted stratiform Pb-Zn deposits are restricted to Lower and Middle Aldridge sedimentary rocks. The Sullivan deposit and the Vulcan occurrence, a thin stratiform pyrrhotite laminated zone containing minor lead and zinc (Gifford, 1971), occur at the Lower-Middle Aldridge boundary. North Star and the transgressive Stemwinder deposit, just south of Sullivan, are in Lower Aldridge siltstone, and the Kootenay King deposit is in the Middle Aldridge siltstone.

Carbonate-hosted deposits occur in dolomite of the Mount Nelson Formation on the northern edge of the Purcell Anticlinorium. They include the Mineral King (Fyles, 1959), Paradise (Hedley, 1950; Atkinson, 1977), and Ptarmigan (Hedley, 1950). Numerous small silver-lead-zinc vein occurrences are common in the Middle Aldridge southwest of the Sullivan area. The St. Eugene is a vein system that cuts across the Middle and Upper Aldridge and the lower part of the Creston Formation. The excursion will stop briefly at the St. Eugene on the morning of Day 2.

Structural and Stratigraphic Control of Clastic-Hosted Proterozoic Deposits

The Purcell Mountains exhibit a pronounced northeast trending structural grain that is delineated by late transverse faults with attendant localization of granitic intrusions. This grain is inherited from older, fundamental faults that were active during sedimentation in late Proterozoic Windermere time and lower Paleozoic time (Leech, 1958; Price and Lis, 1975; Lis and Price, 1976). It has been suggested (Höy, 1979a; and in press) that these faults locally modified the depositional pattern of older Purcell rocks. Rapid lateral facies changes, dramatic local variations in sediment thicknesses, and intraformational conglomerate are the most obvious record of these syndepositional faults (Höy, in press). A pronounced Bouger gravity low and a magnetic lineation trend southwesterly south of Kimberley in the vicinity of the St. Mary - Boulder Creek fault and the Moyie - Dibble Creek fault. This, combined with an anomalous thickness of Middle Aldridge turbidite beds east of the trench suggested to Kanasewich, *et al.* (1969) that a southwest trending Precambriam rift projected under flat lying Paleozoic and Mesozoic rocks from Alberta into southeastern British Columbia.

Areas of tourmalinization (boron concentration) that Ethier and Campbell (1977) suggest are concentrated near the basin fracture zones, and intraformational conglomerate that Höy (in press) relates to syn-sedimentary faults appear to be concentrated in the vicinity of the transverse structural zone south of Kimberley, and in a zone coincident with the Hall Lake fault to the north. The Sullivan, Stemwinder, North Star and Kootenay King deposits are all preferentially located in the vicinity of this transverse zone. Both the Sullivan and the Kootenay King deposits occur in somewhat thickened sections and the Sullivan and a number of other showings are associated with intraformational conglomerate and tourmalinization, suggesting that basin fractures and mineral deposits are genetically linked. Local north-trending breccia zones at Sullivan (Ransom, 1977a; Hamilton, *et al.*, in press) suggest that third order basins cross-cutting a regional northeast-trending second order rift structure were the local control of mineralization.

DAY 1

THE SULLIVAN OREBODY

J. M. Hamilton, R. L. Hauser and P. W. Ransom

Introduction

Day 1 will be spent underground at the Sullivan Mine. The following discussion outlines the important geological features of the mine including basic stratigraphic, structural and mineralogical relationships, as well as a review of data and interpretations that have resulted from the last decade of intensive geological research.

The Sullivan orebody is a large gently dipping iron-lead-zinc sulphide deposit lying conformably in Helikian clastic metasedimentary rock of the Aldridge Formation of the Purcell Supergroup (Fig. 2). The Aldridge Formation is a flysch-like sequence at least 3800 m thick. The orebody is located at the western edge of the Rocky Mountain Trench and on the eastern flank of the Purcell Mountains. Showings of sulphide mineralization were discovered in 1892, but little work was done prior to completion of a rail line to Kimberley in 1899. Early development of the property included a smelter in nearby Marysville in 1903. This venture failed owing to metallurgical difficulties, and the property was acquired in 1909 by the Federal Mining and Smelting Company. The Consolidated Mining and Smelting Company of Canada Ltd. (now Cominco) took a lease and option on the property in late 1909 and completed purchase by 1913.

A main adit at the 3900 foot level, some 200 m below the showings, was started in 1915 and a major effort was directed toward solving the milling and metallurgical problems. In 1920 a differential floatation process was developed that proved successful for obtaining lead and zinc concentrates suitable for smelting. From the date of acquisition by Cominco to the end of 1980 the Sullivan Mine produced approximately 125,500,000 short tons (113,900,000 tonnes) containing 6.7 per cent lead, 5.8 per cent zinc and 2.4 oz. per short ton (81 gms per tonne) silver (Fig. 4). Remaining diluted reserves at the end of 1979 were 54,000,000 short tons (49,000,000 tonnes) containing 4.5 per cent lead, 5.9 per cent zinc and 1.1 oz. per short ton (37 gms per tonne) silver.

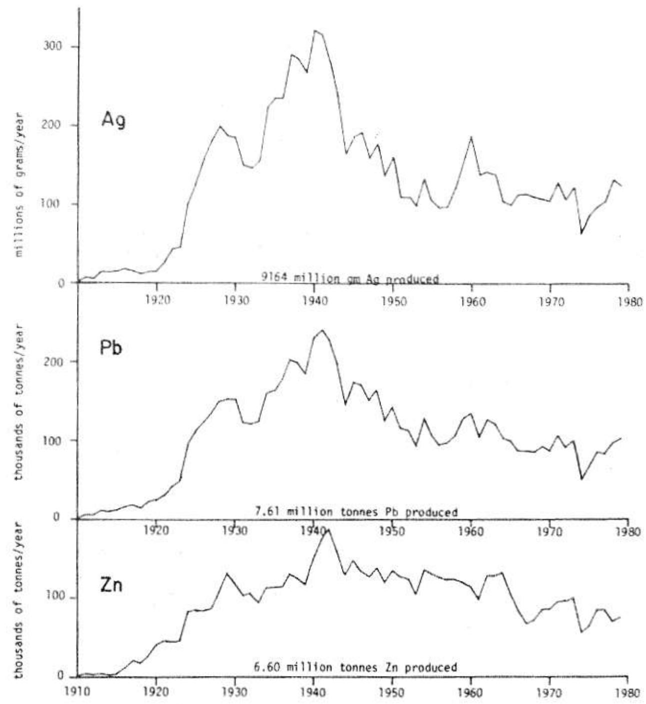

Figure 4. Annual production of silver, lead and zinc from the Sullivan mine, 1910 to 1980.

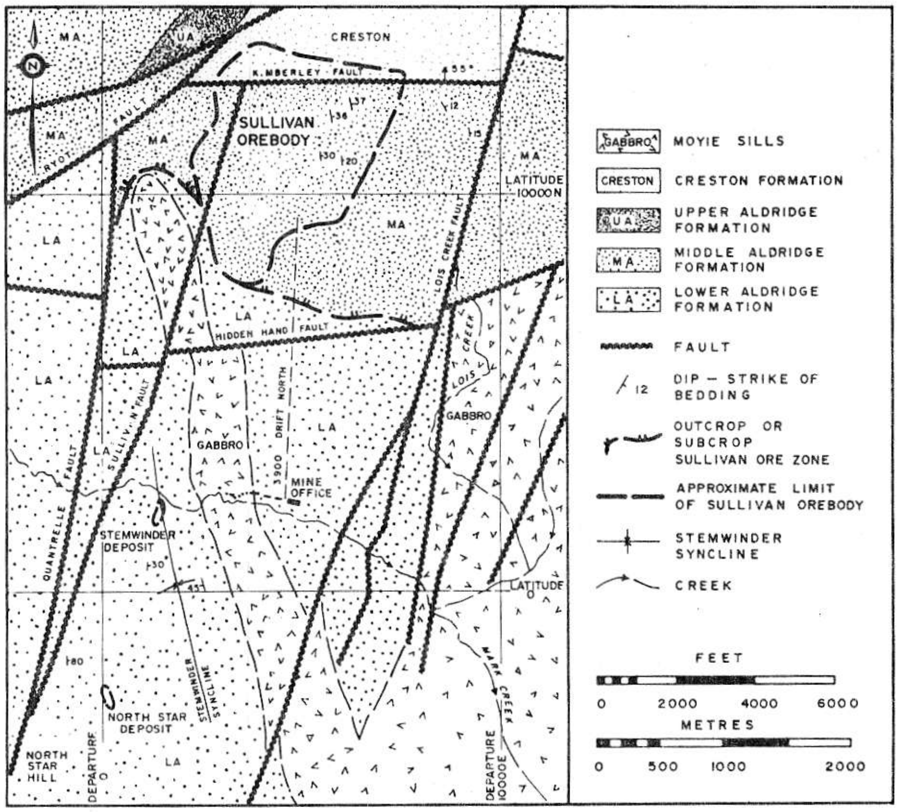

Figure 5. Geology of the Kimberley area.

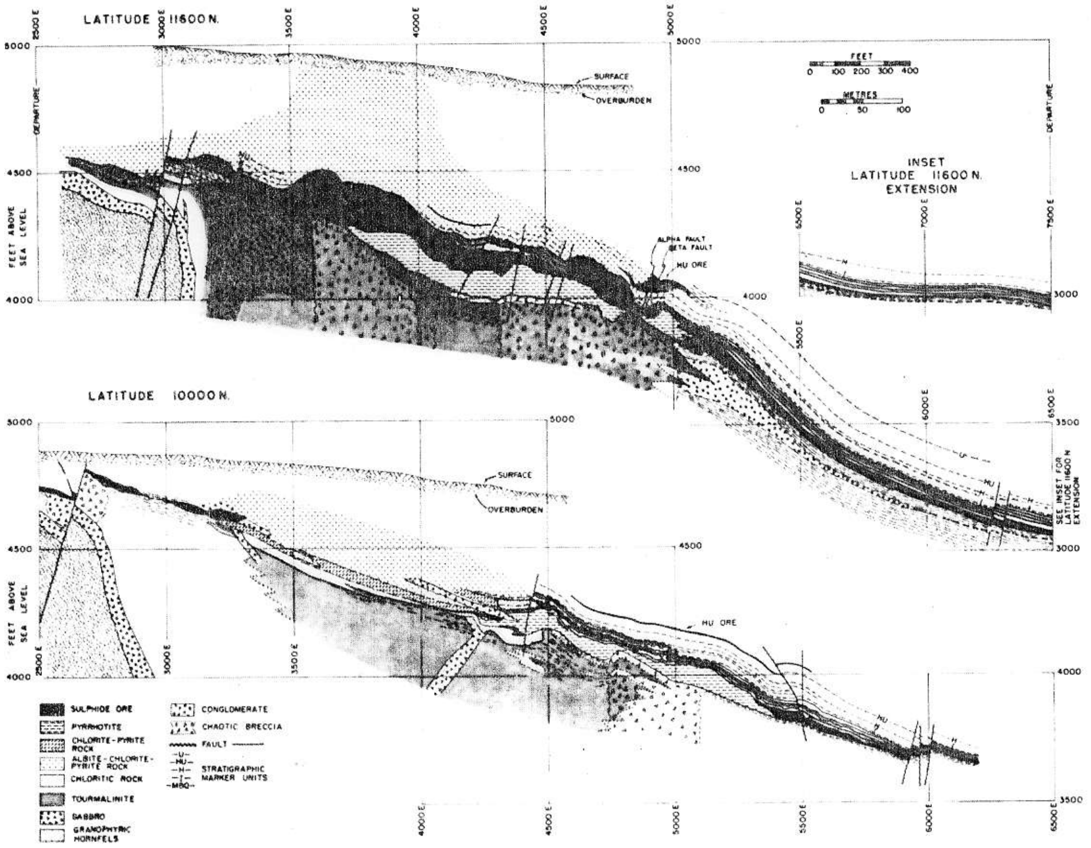

Figure 6. "East-west" vertical sections through the Sullivan orebody.

This discussion draws heavily from earlier works by Cominco geologists, including Freeze (1966), and, in particular, Ransom (1977) and Hamilton, *et al.* (in press).

Geological Setting

The Aldridge Formation in the Purcell Mountains is divided into Lower, Middle and Upper divisions. The Lower division is at least 1500 m thick (base not exposed) and is composed of a rhythmic succession of thin- to medium-bedded very fine grained wacke. The Sullivan orebody is at the top of the Lower division (Fig. 5). The Middle Division is 200 m of medium- to thick-bedded quartzose turbidites. The Upper division is 300 m of thin bedded to laminated mudstones. The Aldridge Formation is metamorphosed to middle to upper greenschist facies.

These rocks have been intruded by the Moyie Sills, gabbroic in composition, which make up 25 per cent of a typical Lower Aldridge section but become less numerous in the Middle Aldridge and are absent in the Upper Aldridge. Zircons from differentiates of a sill near the top of the Lower division provide concordant uranium-lead ages of 1430 ± 20 million years (Zartman, *et al.*, in press).

The Purcell Anticlinorium is the dominant regional structure. It is characterized by open folds plunging gently to the north and to the south. The Sullivan occurs on the east side of this regional structure, on the east limb of an open anticline. The orebody is up to 2000 m north-south by 1600 m east-west. It has flat to gentle easterly dips in the west, moderate easterly to northeasterly dips in the centre, and gentle easterly to northeasterly dips in the east.

The ore zone is truncated on its north side by the Kimberley fault. This east-striking fault dips 55 degrees north, with an apparent stratigraphic displacement of 2000 m, north side down. "Sullivan-type" normal faults also cut the orebody. These faults strike 010°, dip steeply west and exhibit west side down displacements in the ore zone up to 25 m. Larger displacements occur on similarly oriented faults located east of the orebody.

Geology of Footwall Rocks

Bedded Sequence. This unit is at least 150 m thick and is mainly composed of graded quartz wacke beds 10 to 30 cm thick. Cross laminae are rarely observed. The main mineral constituents are quartz, sericite biotite and some carbonate. Pyrrhotite laminations up to 1 mm thick occur in the uppermost 5 to 15 m of the unit, at intervals of about 1 per cm near the base to 5 per cm near the top.

Conglomerate. An intraformational conglomerate occurs near the top of the bedded sequence beneath the northern two-thirds of the orebody (Figs. 6 and 7). Its footwall contact is often disconformable in the west, where conglomerate thickness exceeds 80 m, but often only slightly disconformable or conformable in the east. Its hanging wall contact is generally conformable and is normally about 10 to 15 m below the sulphide footwall.

Clast lithologies and matrix compositions are the same as those of other footwall sedimentary rocks. No foreign clasts have been observed. Clasts are generally 1 to 3 cm in diameter but range from 2 mm to at least 1 m and comprise 10 to 95 per cent of the rock. Pyrrhotite is a typical but variable accessory mineral accentuating either matrix or clasts. Pyrrhotite occurs as laminations in certain clasts and as rims on other clasts. The footwall conglomerate may be extrusive (Shaw and Hodgson, personal communication, 1977) or it may be a slump feature.

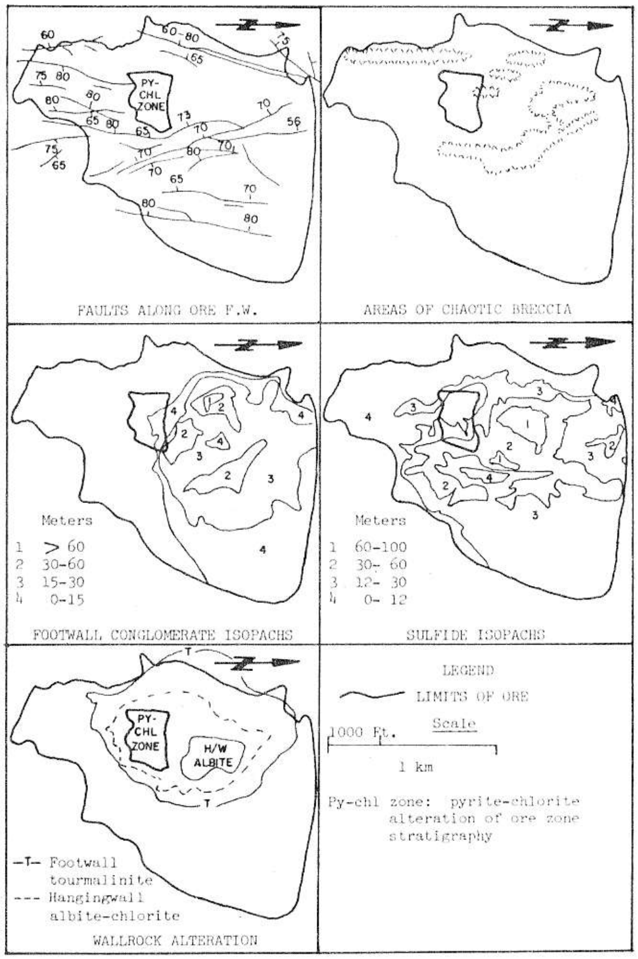

Figure 7. Faults, breccia distribution, conglomerate and sulphide isopachs, and wall-rock alteration, Sullivan deposit.

Chaotic Breccia. Generally north-trending bodies of breccia which have steep dipping contacts are located beneath the western part of the orebody (Jardine, 1966). These extend down unknown distances from the sulphide footwall (Fig. 6). The largest breccia body has dimensions of 120 by 900 m (Fig. 7). Breccia fragments range in size from 1 cm to 10 m, and exhibit a large range of degrees of roundness. Pyrrhotite is a common matrix constituent and galena and sphalerite are locally abundant. Chaotic breccia zones are now interpreted as conduits through which ore-forming fluids passed.

Alteration. The western half of the orebody is underlain by tourmaline rich rock (Figs. 6 and 7). This tourmalinite zone is funnel-shaped and extends at least 450 m below the sulphide footwall. Most tourmalinite appears to be due to metasomatic addition of boron, but the upper 50 m of the zone exhibits some features consistent with syngenetic availability of boron. In hand specimen tourmalinite resembles black or dark brown chert, and it is called chert locally. In thin section, quartz grains with indistinct boundaries are seen to comprise 15 to 50 per cent of the rock and are set in a felted matrix of tourmaline cyrstals 1 to 10 microns long. Pyrrhotite and subordinate sphalérite comprise 0 to 15 per cent of tourmaline rich rock, as disseminated grains or in delicate laminae.

Intensely chloritic rocks occur locally near the footwall of the western part of the orebody. Chloritic rocks occur along the eastern margin of the footwall tourmalinite zone, in crosscutting zones of large vertical extent below the centre of the orebody and in a comformable zone up to 10 m thick immediately underlying the centre of the orebody (Shaw and Hodgson, 1980*b*). Albite-rich rocks are commonly associated with pyrite-rich chloritic rocks in the area underlying the pyrite-chlorite-calcite assemblage of the ore zone (Shaw and Hodgson, 1980*b*).

Vein Mineralization. Sulphide veins are present in the footwall in and adjacent to the zone of tourmalinite and very rare elsewhere. Irregular veins commonly form networks composed dominantly of pyrrhotite, galena and sphalerite. Generally minor amounts of quartz, arsenopyrite, chalcopyrite, cassiterite, tourmaline or scheelite occur in some veins.

Footwall Intrusive Complex. A pair of gabbro sills 15 to 60 m thick separated by about 90 m of granophyre occur about 450 m below the eastern part of the orebody. This intrusive-granophyre package has the form of a gently north-plunging arch several km long which passes below and west of the western margin of the orebody (Figs. 5 and 6). The sides of the arch have steep to moderate dips. The arch is truncated at its north end by the Kimberley fault. Granophyre is interpreted to be highly altered sedimentary rock. Origin of the arch feature in the footwall intrusive complex is not well understood but intrusion into sedimentary layers draped over a syn-depositional horst is one possibility.

Geology of the Ore Zone

Introduction. The Sullivan orebody can be classified into two distinctive parts joined by a transition zone. The up-dip or western part of the orebody (generally that part underlain by tourmaline rich footwall rock, Fig. 7) is much thicker than the eastern part (Fig. 7). The western part is typified by sulphide layers which are relatively internally homogeneous and which contain relatively little intercalated waste layers. The transition zone up to 75 metres wide, is generally coincident with the eastern margin of footwall tourmalinite. Outboard of the transition zone, in the eastern part of the orebody, ore occurs in five laterally persistent "Bands" intercalated with significant clastic sedimentary rock layers (Fig. 8). Ore Bands in the eastern part of the mine are typified by delicate layering of sulphides.

Western Part. The ore zone averages about 50 m thick but relatively abrupt thickness changes within the 10 m to 100 m range occur. Conformable massive pyrite layers up to 5 m thick occur intermittently at the sulphide footwall and hanging wall. Nearly barren pyrrhotite lenses 10 m or more thick occur. The largest of these is at or near the sulphide footwall; it ranges up to 35 m thick, with a length of 350 m. Layering is not well developed in the barren pyrrhotite. Sphalerite and galena occur as wispy concentrations, disseminated grains, fracture fillings and veins. The top of the main pyrrhotite lens grades into an intermittently developed zone in which wispy layering is defined by galena and sphalerite concentrations. Further up section, layering becomes more pronounced and contorted, and discontinuous intercalated layers of clastic sedimentary rock are present. At the top of the western part of the orebody, thin bedded to laminated sulphide rock may be present.

Thick sections of internally homogenous sulphide containing no intercalated layers of clastic sedimentary rock and only a minor component of non-sulphide grains are thought to be features indicative of relatively rapid accumulation of sulphides. Layering of sulphides and presence of some intercalated waste beds near the top of the western part of the ore zone suggest a gradually decreasing rate of sulphide deposition.

An intensely developed pyrite-chorite-calcite alteration assemblage crosscuts the complete ore zone within the area indicated on Figures 6, 7 and 10. Contact zones between it and adjoining pyrrhotite or ore are 1 to 15 m wide. Fluids which led to development of a hanging wall alteration assemblage are regarded as having created the pyrite-chlorite-calcite assemblage in the ore zone (Shaw and Hodgson, 1980a, 1980b).

Eastern Part. The stratigraphy of the ore zone and enclosing rocks in the eastern part of the orebody, down dip from the transition

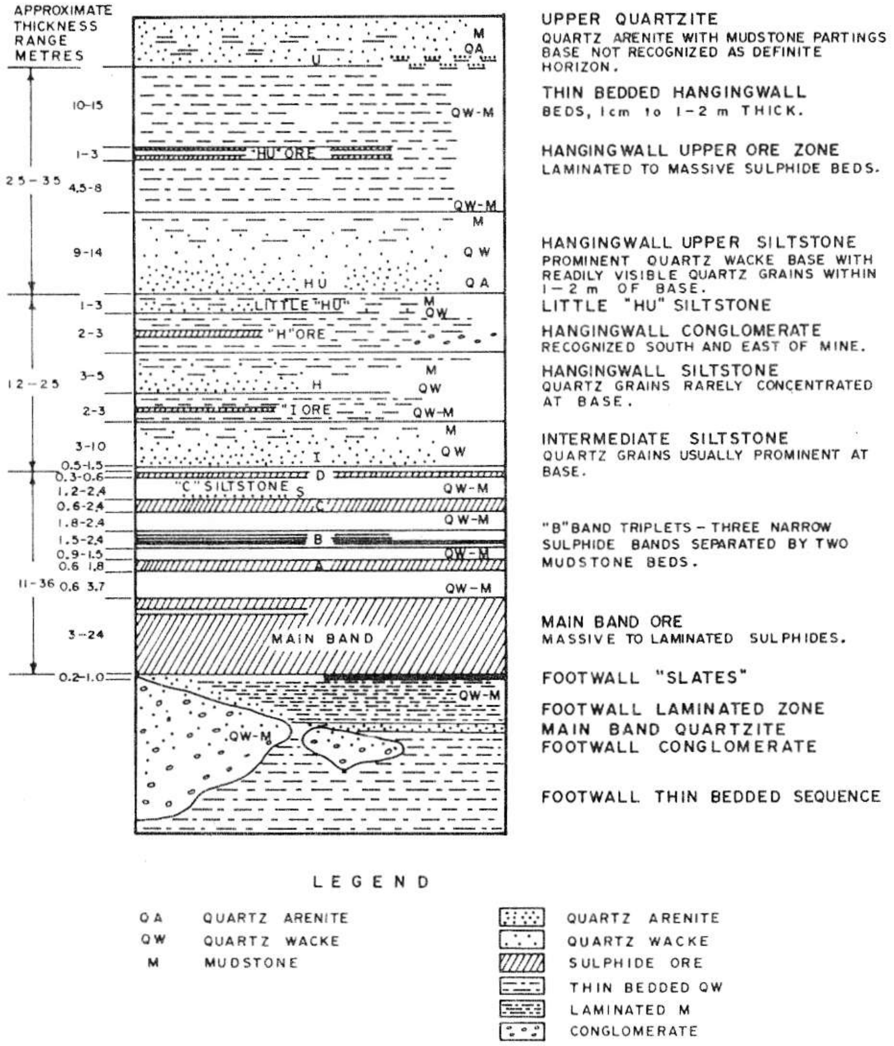

Figure 8. Ideal section of the bedded ore, Sullivan deposit.

zone, is shown in Figure 8. Ore extends from the base of the Main Band to the top of the D Band.

The Main Band rests on the Footwall "Slates" with a very sharp conformable contact. The basal half to two-thirds of the Main Band is different from over-lying sulphides. It consists of a succession of fine grained pyrrhotite, sphalerite and galena beds without intercalated clastics. The sulphide beds contain different proportions of the three sulphide minerals, and while generally less than 3 cm thick, range up to 30 cm thick. In some areas, notably the eastern fringes of the orebody, the basal part of the Main Band contains megascopic fragments and inclusions which together make up to 20 per cent of the rock. These are generally less than 1 cm across, but the fragments may range up to 1 m across. The common fragment types are altered and unaltered clastic sedimentary rock, carbonate rock and granular quartz-rich rock. Monomineralic inclusions of pyrite, calcite, quartz, sphalerite, scapolite and garnet are more numerous but volumetrically not greatly more significant than lithic fragments.

The upper portion of the Main Band, and overlying Bands contain up to 40 per cent of thin, closely spaced interbeds of mudstone and quartz wacke. Grain size of the sulphide layers is coarser but more variable than in the basal part of the Main Band. Sulphide beds are generally thinner and have a greater tendency to be monomineralic. Sections several cm thick composed entirely of delicate nearly monomineralic sulphide lamellae 0.3 to 5 mm thick are common in this part of the ore zone.

Four graded beds of intercalated waste rock separate the five Bands and make up 25 to 40 per cent of the ore zone, or about 50 per cent of that portion of the ore zone which overlies the Main Band. These four beds are similar to but thicker than those in the footwall bedded sequence.

Metal Distribution. Vertical distribution of metals in a portion of the eastern part of the orebody is shown on Figure 9. Iron concentration is roughly constant at 20 to 25 per cent. Zinc concentration ranges from about 20 per cent in the Main Band to about 7 per cent in the D Band. Lead concentration is greater at the top of the Main Band than at the bottom, reaching a maximum of about 14 per cent, decreasing to 1.4 per cent in the D Band. Vertical distribution of silver is similar to that of lead, decreasing from 50 to 5 gms per tonne.

Distribution of metals in plan is shown on Figure 10. Lead and zinc distribution maps show products of thickness times grade for those metals. Not surprisingly, most lead and zinc is located where the orebody is thickest. However, distribution maps for silver and tin show grade only. In general, highest grade areas for these metals are located in the western portion of the orebody. Of particular interest are the silver to lead ratio and lead to zinc ratio

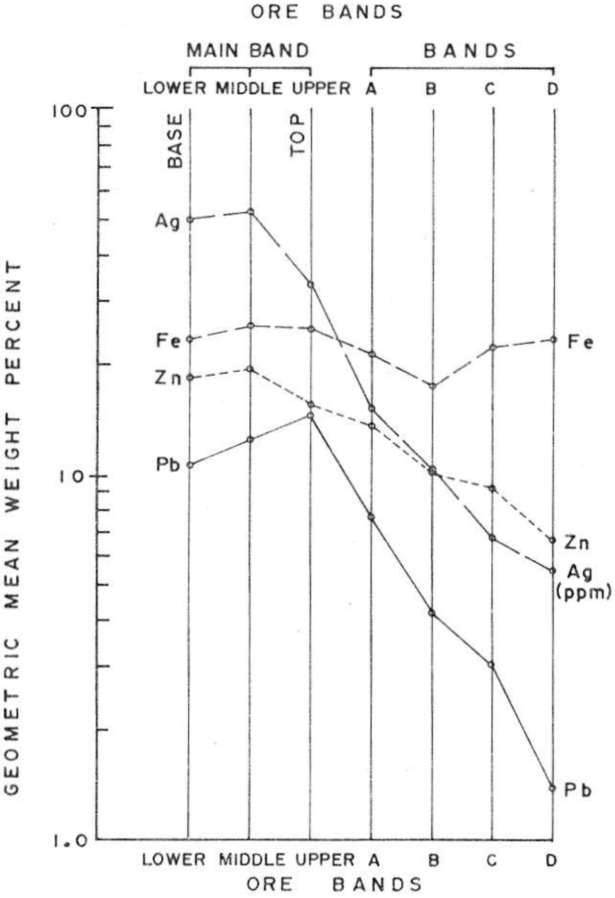

Figure 9. Vertical metal distribution in ore bands, Sullivan deposit.

plans. The silver to lead ratio plan shows that silver is enriched with respect to lead in the western part of the orebody. Similarly, the lead to zinc ratio is highest there.

These data are consistent with deposition of lead, silver and tin preferentially in the area of the orebody proximal to the vent zones, thought to be represented by areas of chaotic breccia (Fig. 7). Zinc is more concentrated than lead in more distal area.

Sulphur Isotopes, Geothermometry and Geobarometry. Detailed sulphur isotope investigations (Campbell, Ethier, Krouse and Both, 1978; Campbell, Ethier and Krouse, 1980) show that within the ore zone there is a general increase in relative abundance of ^{32}S with increasing stratigraphic position. Campbell, Ethier and Krouse (op. cit.) ascribe this variation to changing conditions affecting the supply of H_2S in the basin and temperature gradients existing in the basin, and conclude that Proterozoic seawater was the sulphur source.

Ethier, Campbell, Both and Krouse (1976) studied mineral relationships to obtain values for metamorphic temperatures and pressures. Arsenopyrite data provide a temperature estimate of 400° to 490°C; fractionation of sulphur isotopes suggests temperatures of 300° to 340°C; oxygen isotope fractionation between quartz and magnetite indicates a 400° to 560°C range. Sphalerite geobarometry work suggests pressures of 5 kilobars. As they point out, this figure is at least double that consistent with the known post-ore succession.

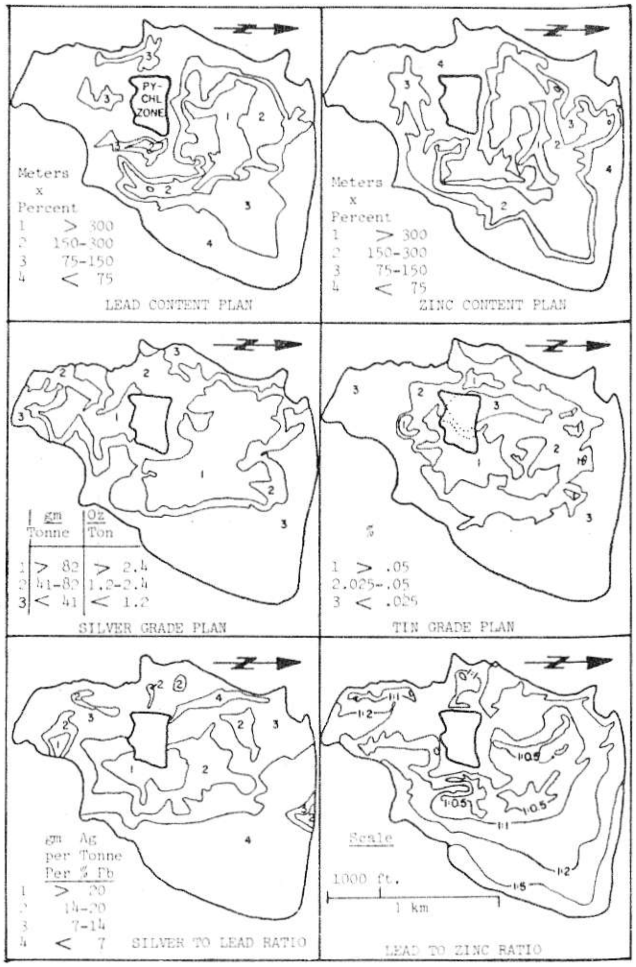

Figure 10. Metal distribution maps, Sullivan deposit.

Structural History. Detailed structural mapping (McClay, in preparation) has revealed three phases of folding. Phase 1 is characterized by isoclinal folds with axial planes parallel to bedding planes and north-trending fold axes. Phase 2 is characterized by relatively open folds with gentle north or south plunges and with moderately west-dipping axial planes. Both Phase 1 and Phase 2 folds indicate easterly vergence. Phase 3 folds are associated with east-dipping thrusts; axial planes have steep dips and folds have variable plunges to NW or SE. Synsedimentary slump features occur but are extremely rare. East-dipping low angle thrusts associated with Phase 3 folds offset early Cretaceous lamprophyre dykes.

Geology of Hanging Wall Rocks

Stratigraphy. The ore zone is overlain by four distincitive graded series: the I, H, Little HU and HU series. Each has a graded quartz wacke bed at the base which grades upward to a thick mudstone top. The top of the mudstone portions are commonly pyrrhotite-laminated. Distinctive pyrrhotite-laminated sections have been matched over distances of up to 1500 m. Up dip and to the south, these pyrrhotite-laminated sections grade into ore. The I, H, Little HU and HU series are not well documented outside the immediate mine area. They are regarded as products of largely post-ore infilling of the sub-basin in which the Sullivan orebody was deposited.

A hanging wall conglomerate averaging 4 m thick occurs stratigraphically beneath the H sulphide laminations. This conglomerate has been traced for about 1000 m along the southern margin of the orebody. Like the footwall conglomerate, the hanging wall conglomerate contains no foreign clasts. However, in contrast, it does contain some ore-grade clasts. It appears to correlate with the top of a largely post-ore breccia located in the southwest part of the orebody. Evidence for an extrusive origin for the hanging wall conglomerate is persuasive.

The HU series is overlain by 10 to 15 m of graded quartz wacke beds, and then by the U Quartzites which are regarded as the base of the Middle division of the Aldridge Formation.

Alteration. Hanging wall rocks above the western portion of the orebody are intensely altered to an albite-chlorite-pyrite assemblage up to 125 m thick (Fig. 6). Massive albitite in which original sedimentary textures have been obliterated forms a core that grades outward to an envelope of albitic and chloritic rock in which original gross sedimentary layering is preserved. The massive albitite core is located about 225 m northeast of the centre of the pyrite-chlorite-calcite alteration assemblage in the ore zone. Structural interpretation suggests that the hanging wall is displaced tó the northeast (McClay, in preparation); crosscutting dykes assumed to be apophesies to the footwall intrusive complex are offset in that direction. Accordingly, it seems reasonable that the hanging wall alteration assemblage was once centred over the pyrite-chlorite-calcite assemblage that crosscuts the ore zone.

Geological History

The Sullivan orebody is interpreted as a hydrothermal synsedimentary deposit which formed in a sub-basin on the Aldridge marine floor. It is located directly over conduits through which mineralizing fluids passed. Cross-strata permeability developed along synsedimentary faults and fractures; fluid escape along these led to development of chaotic breccia zones. Footwall conglomerate was extruded from breccia pipes (Shaw and Hodgson, personal communication, 1977), or was laid down when locally oversteepened sediments collapsed.

Boron-rich fluids percolated up the zones of cross-strata permeability, soaking adjoining footwall sediments and discharging onto the sea floor. Then fluid composition and/or conditions in the sub-basin changed, and sulphides were deposited. Initial sulphide deposition over the vent area was rapid, as evidenced by lack of layering in sulphides and by lack of included clastic sedimentary rock. These features are felt to be consistent with deposition of sulphide particles which issued from the vent area.

Pyrrhotite deposition predominated during initial sulphide accumulation. Later, pyrrhotite interlayered with galena and sphalerite became dominant in the western part of the orebody. Deposition of the basal half of the Main Band may have occurred at this time, although detailed correlation of sulphide stratigraphy across the transition zone is difficult. Deposition of sulphides by chemical precipitation from a brine pool may have become the dominant mechanism during this period.

Waning stages of sulphide deposition were much less violent, and well layered sulphides intercalated with intermittent influxes of clastic sediments became the dominant depositional style. In the upper part of both the eastern and western portions of the orebody, delicate sulphide lamellae consistent with chemical precipitation are widespread.

The I, H, Little HU and HU graded sequences were laid down during filling of the sub-basin. Deposition of clastic sediments predominated but minor deposition of sulphides continued.

Composition of fluids changed again. Sodium-rich fluids percolated through the ore-bearing sequence and created albite-chlorite-pyrite alteration in the footwall, pyrite-chlorite-calcite alteration in the ore zone, and albite-chlorite-pyrite alteration in the hanging wall (Shaw and Hodgson, 1980a, 1980b).

Moyie sills (1430 ± 20 m.y.) were emplaced. Their form and contact relationships are consistent with emplacement under 2 to 4 km of overlying sediment, indicating emplacement during deposition of the overlying Creston Formation (Zartman, *et al.*, in press).

Three phases of folding are recognized (McClay, in preparation). The first two are of pre-early Cretaceous age, and since they suggest a similar stress field, may be close in time. They correlate with regional development of the Purcell Anticlinorium. The third phase is associated with east-dipping thrust faults which displace early Cretaceous lamprophyre dykes.

ACKNOWLEDGEMENTS

People who write about Sullivan geology stand on a lot of shoulders. We say thank you to the dozens of careful observers, recorders and interpreters of geological data at Sullivan who have preceded us. This manuscript was significantly improved during critical review by D. R. Shaw. K. R. McClay's comments are also appreciated.

DAY 2 (A.M.)

THE PROTEROZOIC PURCELL SUPERGROUP IN THE MOYIE LAKE AREA

Trygve Höy

On the morning of Day 2 the excursion crosses the Purcell Anticlinorium, from the Sullivan deposit at Kimberley just west of the Rocky Mountain Trench to Creston near the western edge of the Anticlinorium. Successive stops along the excursion show the main lithologies of the Purcell Supergroup, beginning with a Lower Aldridge exposure a few kilometres south of Kimberley and moving generally up-section to a Kitchener exposure at Moyie Lake. Most stops are in the Moyie Lake area, and as it has been recently remapped in detail (Höy and Diakow, 1981), its geology will be described briefly.

The Moyie Lake area is in the southern Purcell Mountains near the central part of the Purcell Anticlinorium. The area (Fig. 11) is cut by the northeast trending Moyie fault, that continues northeast across the Rocky Mountain Trench and southwest into Montana. The fault appears to follow the locus of an older fault along which the north side was relatively down-dropped (Leech, 1958) in pre-late Devonian time. North of the Moyie fault a thick sequence of Cambrian rocks lies unconformably on Upper Purcell rocks. Latest movement on the Moyie fault has resulted in a net vertical displacement of several kilometres, and at the north end of Moyie Lake, Lower Aldridge rocks on the north hanging wall of the fault are juxtaposed against Kitchener Formation footwall rocks (Fig. 11). As suggested by Leech (1962), a "basal Devonian" unit, comprising a fluvial conglomerate and gypsum horizon, has acted as a glide zone for the Moyie fault. The conglomerate has been recognized in a number of places in the footwall of the fault to the western limit of the area mapped (Höy and Diakow, 1981).

The structure of the area south of the fault, the "Moyie block" of Benvenuto and Price (1979), is dominated by the Moyie anticline, a northeast plunging upright anticlinal fold. The western limb of the Moyie anticline is cut by the Moyie fault. Lower and Middle Aldridge rocks are folded into moderately tight to open north to northeast trending folds that are outlined by metagabbro sills. In the hanging wall immediately adjacent to the Moyie fault, folds are tight and locally overturned. The Lower Aldridge is exposed in two overturned anticlinal folds just west of the north end of Moyie Lake and west of Cranbrook Mountain (Fig. 11).

Road Log:

Kimberley to Creston
Highway 95A to Cranbrook,
Highway 3/95 to Yahk,
and Highway 3 from Yahk to Creston.

Trygve Höy and F. R. Edmunds

The excursion begins in the Lower Aldridge below the Sullivan horizon at Kimberley and continues in stratigraphically higher rocks in the Moyie Lake area. Stops 2, 3 and 4 are in the Middle Aldridge above the Sullivan horizon, and stops 5, 6 and 7 are in stratigraphically overlying Purcell rocks on the west limb or near

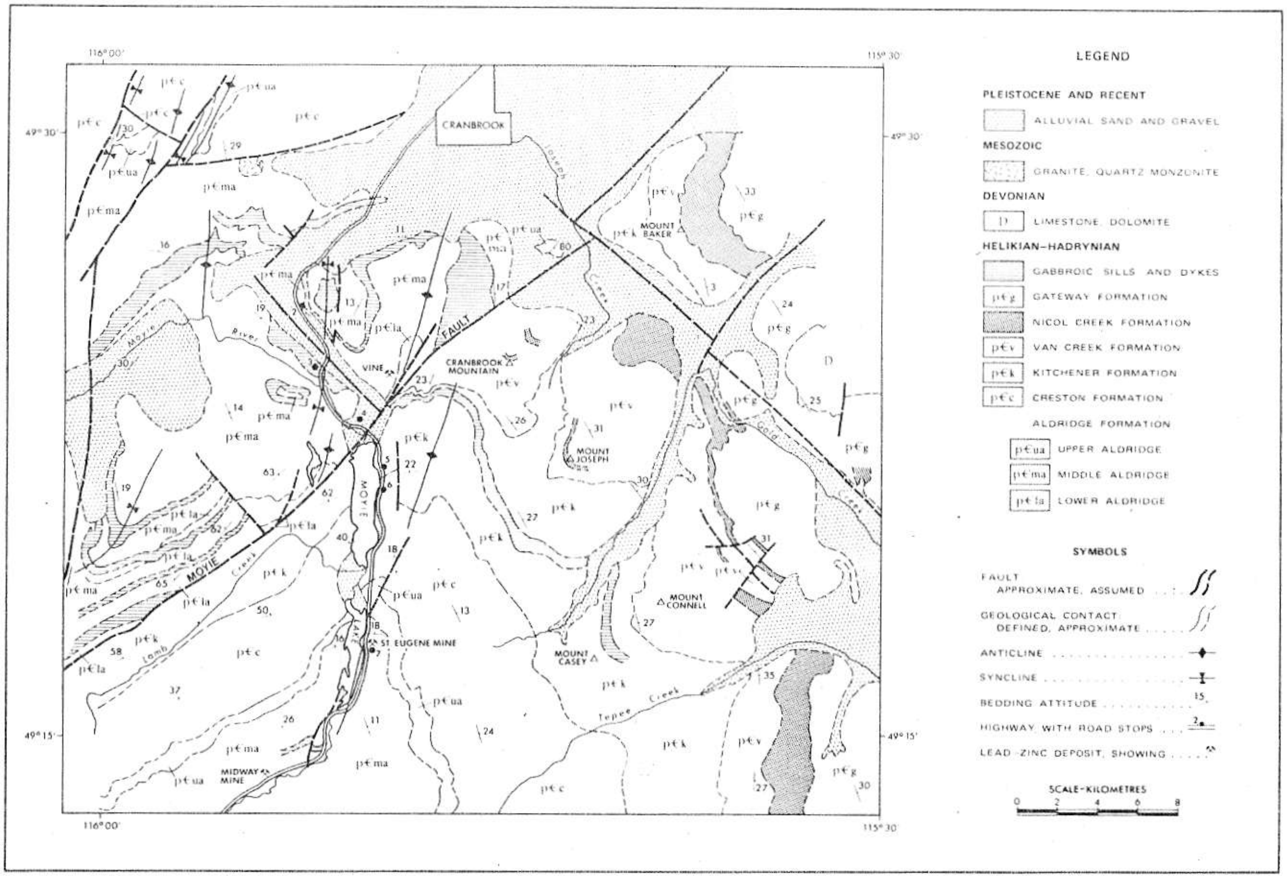

Figure 11. Geological map of the Moyie Lake area, showing excursion route and stops (after Höy and Diakow, 1981).

the crest of the Moyie anticline southeast of the Moyie fault.

The following road log, originally written by Edmunds (1977), has been modified by Höy based on recent mapping in the Moyie Lake area.

0.0 km Kimbrook Inn, Kimberley.

STOP 1-1, 4.0 km, Mark Creek and Marysville. Lower Aldridge sedimentary rocks within a few metres of the Sullivan Mine horizon are exposed in the gorge below the bridge. In general, these are thin-bedded sub wackes and argillites although metasandstone units occur. The Stop permits a comparison with the turbidite-bearing Middle Aldridge Formation at STOP 2. Particularly note the bedding thickness, which is substantially smaller and less well-defined than in the Middle Aldridge.

6.8 km The trace of the St. Mary Fault. Cambrian sedimentary rocks in the treed area on the east lie unconformably on the Kitchener Formation in the upper part of the Purcell Supergroup.

13.0 km Village of Wycliffe. We detour here, but a new bridge, to be operational starting September 1981, is the basis upon which distances are given.

18.4 km St. Mary River. Outcrops of Kitchener Formation occur along the river bank.

18.4 to 26.00 km The route runs on glacial gravels. The underlying units pass down-section through the Kitchener, Creston, and Upper Aldridge Formations, into flat-lying Middle Aldridge west of Cranbrook.

26.4 km Junction with Highway 3/95 from Radium and Fernie.

26.8 to 33.2 km City of Cranbrook.

The route continues through essentially flat-lying, poorly exposed lower Middle Aldridge Formation. Cliffs visible occasionally on the north side of the highway are metagabbro sills.

STOP 1-2, 45.8 km, outcrop on Highway 3/95 just opposite Lumberton turn-off. A sequence of turbidite units and interbedded argillites is exposed on the east side of the highway. It lies in the Middle Aldridge Formation, about 1000 metres above the base. The presence of AE turbidites up to 1 metre thick may be contrasted with the Lower Aldridge lithology of STOP 1 and rocks encountered in the Sullivan Mine.

The turbidite units are generally graded only at the top, but occasionally graded in a reverse manner in their basal part; the central part of coarser grained turbidite beds is commonly massive. Within the finer grained silt and fine sand-sized turbidites, parallel laminations, convolute laminations, ripple bedding, and flame structures occur. Sole markings, including grooves, flute marks, furrows and ridges, tool marks, and load casts, are common. Current indicators in Middle Aldridge turbidite deposits in the Moyie Lake area show a northerly transport direction. Concretionary bodies and rip-up clasts occur in some beds, and large crystal casts with the swallow tail form of selenite are in some of the more argillaceous beds.

STOP 1-3, 48.8 km, Outcrops of a metagabbro sill on both sides of the highway and in a railway cut 50 metres to the north. Metagabbro sills comprise a considerable portion of the Lower and Middle Aldridge succession in the Purcell Mountains (Fig. 3). They are regionally extensive and in the Moyie Lake area have been used as stratigraphic markers (Höy and Diakow, 1981). The sills are correlated with the Crossport "C" sill in Idaho which yielded a concordant U-Pb zircon age of 1430 ± 20 m.y. (Zartman *et al.*, in press).

49.6 km Moyie River Bridge.

51.5 km Moyie River Bridge.

STOP 1-4, 51.5 km, Outcrop on north side of highway just east of Moyie River bridge. Exposures of Middle Aldridge Formation includes a Marker Argillite in the cliffs east of the road and south of the river. The Marker Argillite occurs at the base of the cliff towards the north end. It is about 3.5 metres thick, but includes up to a metre of non-laminated sedimentary rock. The corroded condition of the unit is due to the weathering of a relatively high proportion (±4%) of iron sulphides.

52.8 km Trace of Moyie fault. Lower Aldridge exposed in an anticlinal fold on the northwest hanging wall side of the fault is juxtaposed against Upper Purcell Kitchener Formation on its southeast side (Fig. 11). The Moyie fault is one of a number of northeast trending faults that transect the Purcell Anticlinorium. Repeated movements in the vicinity of these faults since Proterozoic time have had a fundamental control on stratigraphy and tectonics in southeastern British Columbia. The most recent displacement on the Moyie fault is right-lateral and reverse.

55.4 km Cross Peavine Creek and enter the steep north-plunging nose of the Moyie Anticline. Moyie Lake lies on the west and the top of the Kitchener Formation lies on the east.

STOP 1-5, 56.7 km, Highway outcrops east of Moyie Lake. Lower units of the Kitchener Formation occur on both sides of the road. They are interbedded argillaceous dolomites and dolomitic argillites. Of particular interest here is a structure referred to as "molar tooth". Small crenulated veinlets of calcite, approximately perpendicular to the bedding, weather in negative relief. Their form is quite varied, in some places being almost tubular. In addition, ripple-marked surfaces and scour-and-fill structures can be found in the more argillaceous units. Elsewhere in the Kitchener (and Kitchener-equivalent) stromatolites have been identified. The transitional contact with the underlying Creston Formation occurs about 100 metres south of the de-bussing point.

STOP 1-6, 58.5 km, Creston Formation outcrops on the cliffs on the east side of the highway, east side of Moyie Lake. In general, the outcrops are sub-tidal to supratidal argillite, quartzite and wacke beds in different shades of green, dull grey-green, and maroon from near the top part of the Creston Formation. Sedimentary structures are numerous and include desiccation and synereses cracks, mud-chip breccias, scour-and-fill structures, ripple marks, cross-bedding, and graded bedding. The characteristic bed-form is thin-bedded, commonly laminated argillite-siltite couplets.

59.9 km Transition between the Upper Aldridge Formation and the Creston Formation. The Upper Aldridge is a dark, carbonaceous, thin-bedded argillite succession. The Creston in the transition zone is a green colored alternation of thin argillites and wackes, weathering rusty (Aldridge characteristic), but slightly green on relatively fresh surfaces. Some sand lenses can be found, as well as silt "intrusions" and penecontemporaneous pull-apart structures.

For the next 2.4 km the highway runs over recessive weathering Upper Aldridge Formation carbonaceous argillite and siltite.

63.6 km Highest exposures of the Middle Aldridge turbidite units.

63.9 km Town of Moyie.

STOP 1-7, 64.5 km, St. Eugene Mine. The entrance to the St. Eugene mine is on the east side of the highway. Adits and associated dumps can be seen up the hillside to the east. On the far shore of Moyie Lake to the west the Aurora Mine dump is visible.

The St. Eugene Mine lies about 2000 metres above the base of the Middle Aldridge Formation. It was discovered in the late 1890's by a Kootenay Indian and acquired by the owners of the Trail smelter in 1905 to supplement diminishing supplies of concentrates. The mine is a ladder-vein, striking WNW and dipping steeply south. The ore is essentially galena and silver. By 1916, when reserves were exhausted at 620 metres below the headframe, production had amounted to 931,430 tonnes grading 12% Pb, 1% Zn and 200 g/tonne Ag.

The following description of mineralization is taken from a Cominco Report:

"Lead, silver and zinc are the most important metals of the St. Eugene vein system and are mainly derived from argentiferous galena and sphalerite. Tetrahedrite is present in some areas, and small amounts of chalcopyrite are generally common in occurrence. Gangue minerals of the productive veins include quartz, biotite, chlorite, garnet, amphibole, pyrrhotite, pyrite and magnetite. Some epidote, grunerite and fluorite is found locally.

Commercial concentrations of these minerals occur in tabular ore shoots within steep-dipping veins and in multiple orebodies within extensive cymoid structures. The average width of mineable material, in both vein-type and cymoid-type deposits is in the order of 2 to 3.5 km.

The main break in which the vein-type shoots form is generally uniform in trend but irregularly refracted in detail. The break is sometimes wide and other times narrow, and may be either filled with gangue minerals or barren.

The cymoid structure is bounded by two main breaks with parallel veins ('parallels') and cross veins ('avenues') transecting the inter-area. The structure is generally well mineralized throughout and larger deposits sometimes reach 10 m in width with one or more bands of near-massive galena up to 1.3 m thick.

All the shoots tend to be longer down their dip than along their strike. The Lake Shore and St. Eugene shoots of the St. Eugene mine are apparently of the cymoid type and appear to pitch moderately to the east, whereas the Moyie shoot, which occurs in a warped area of the North vein, is of the vein-type with a near-vertical pitch. The Aurora shoots, across the lake at the Aurora mine are in the same vein system and are also related to a warped break, but apparently pitch moderately west.

The overall vein system, including ore shoots and barren sections, can be traced for some 3500 m along its strike and some 1400 m down its dip."

71.4 km Midway Mine lies on the north side of the road. This is a small but high-grade gold-quartz vein deposit. Associated metallic minerals are arsenopyrite, pyrite, galena, sphalerite, and chalcopyrite. The earliest records are from 1933 when the property was acquired by the B.C. Cariboo Goldfields Limited. Only a few ore shipments have been made.

No more stops are planned for this leg of the trip. For the next 40 km the route along Highway 3/95 descends through Middle Aldridge stratigraphy as it passes obliquely through the western limb of the north-plunging Moyie Anticline. By the town of Yahk, the route has reached its lowest elevation in the stratigraphy of the block south of the Moyie Fault — just 100 metres above the top of the Lower Aldridge Formation.

95.3 km Town of Yahk.

99.0 km Junction between Highway 3 and Highway 95. The route continues along Highway 3 to Creston for 42 km.

STRUCTURAL AND STATIGRAPHIC SETTING OF KOOTENAY ARC DEPOSITS

T. Höy and E. W. Muraro

Introduction

Stratabound lead-zinc deposits in the Kootenay Arc are essentially restricted to a "platformal" carbonate unit of Lower Cambrian age. The deposits consist generally of lenticular masses of pyrite, sphalerite and galena in dolomite or chert zones within highly deformed limestones (Fyles, 1970a). The larger deposits generally range in size from 6 to 10 million tonnes and contain 1-2% Pb, 3-4% Zn and trace silver. Although a number of stratabound Kootenay Arc deposits are undergoing development work, there is no present production. The Bluebell deposit at Riondel and the Duncan deposit north of Kootenay Lake will be visited on Days 2 and 3 respectively.

Tectonic and Structural Setting

The Kootenay Arc is a north-trending arcuate structural zone that developed in a succession of rocks ranging in age from Hadrynian to early Mesozoic (Fig. 2). In general, the earliest recognized structures in the Arc are tight to isoclinal, north-trending recumbent folds. In the Lardeau area in the northern Kootenay Arc there is some evidence that indicates that these structures may have developed during the Devono-Mississippian Cariboоan orogeny. Here, a Phase 1 syncline in lower Paleozoic rocks is truncated by an uncomformity at the base of the lower Mississippian Milford Group (Read, 1976). As well, a conglomerate at the base of the Milford Group contains clasts of the underlying Broadview Formation in which the earliest foliation varies from clast to clast (Read, 1975, p. 29). More open but locally isoclinal, north-trending Phase 2 folds with upright to steeply west dipping axial surfaces are superposed on the Phase 1 folds (Fyles, 1964; Höy, 1977, 1980). These folds dominate the structure of the Kootenay Arc and account for the pronounced north-south structural grain. In the Lardeau area, radiometric dates appear to restrict the second phase of deformation to an interval between 178 m.y. and 164 m.y. (Read, *et al.*, 1975). The latest discernible deformation in the Arc caused faulting and gentle folding of the earlier structures.

Stratigraphy

Stratigraphic successions and their regional correlation in the southern and northern parts of the Kootenay Arc are shown in Figure 2. They include dominantly coarse clastics and carbonates of the Hadrynian Windermere Supergroup, overlain by quartzite and pelite of the Hamill Group, interlayered clastics and carbonates of the Mohican or Truman Formation and a regionally extensive lower Cambrian marble called Reeves Formation in the Salmo area and the Badshot Formation to the north. Argillite, shale, calcareous shale, quartzite and basic volcanic rocks of the Laib or Lardeau Groups overlie the Reeves-Badshot marble. The more northern section, used in the areas through which the excursion passes, is described below.

The Toby Formation at the base of the Windermere Supergroup is a conglomeratic mudstone (diamictite) unit generally consisting of boulder to pebble-sized clasts of dolomite, quartzite, and argillite that are derived primarily from the underlying Dutch Creek and Mt. Nelson Formations (Rice, 1941; Lis and Price, 1976). The clasts are supported by a fine argillaceous, dolomitic or rarely, silty matrix. Aalto (1971) has suggested that the formation is a tillite. Elsewhere, the Toby Formation consists of well-sorted and closely packed clasts within a sandy mudstone matrix, suggestive of fluvial deposition (Lis and Price, 1976).

The overlying Horsethief Creek Group consists of up to 8500 metres of pelite, slate and grit, with several quartzite and polymict units (Rice, 1941; Lis and Price, 1976). Lis and Price (1976) suggested that the polymict conglomerate units were deposited as mud flows and gravels, derived from the upper part of the Purcell Supergroup. They suggest that these units are fanglomerates which accumulated adjacent to a fault scarp that separated the uplifted source area south of the scarp from a deep structural basin on the north. The Horsethief Creek Group thins from greater than 9 kilometres to only a few kilometres, northward within the basin. Southeast of the fault scarp (now marked the St. Mary Fault, a northwest dipping reverse fault), lower Cambrian quartzites of the Cranbrook Formation lie unconformably on Upper Purcell rocks. The Upper Proterozoic Windermere Supergroup is missing completely and 4 kilometres of Middle Proterozoic strata, exposed further to the northeast, has been eroded away.

The lower Cambrian Hamill Group unconformably overlies the Horsethief Creek Group (Rice, 1941; Lis, personal communication, 1974). It comprises micaceous quartzite, quartz-rich schist, micaceous schist and a 60 to 200 metre thick, white orthoquartzite marker unit (Fig. 12). In the Riondel area massive amphibolite units in the central part of the Hamill are probably metamorphosed basic volcanic layers (Höy, 1980). The Mohican Formation is a gradational unit between the Hamill Group and the Badshot Formation. It consists of a thin-bedded succession of calcareous schist and quartzite, rusty-weathering micaceous schist, and limestone (Fig. 12). The Badshot Formation is an extensive, relatively pure limestone unit several tens to greater than 100 metres thick. Irregular zones of dolomite and chert within the limestone host the lead-zinc deposits. In the northern part of the Kootenay Arc, the Badshot is overlain by up to several hundred metres of dark argillite or micaceous schist at the base of the Index Formation, calcareous schist of the Upper Index, a succession of dark argillite and argillaceous quartzite of the Triune Formation, the Ajax Formation quartzite, and argillite of the Sharon Creek Formation (Fyles and Eastwood, 1962). Metamorphosed andesitic volcanic rocks of the Jowett Formation overlie the Sharon Creek Formation and are overlain by dominantly coarse clastic rocks of the Broadview Formation (Fig. 2). In the Salmo area at the southern end of the Kootenay Arc, argillite, calcareous phyllite and minor micaceous quartzite of the Laib Formation is overlain by a thick limestone-dolomite unit of Middle Cambrian age called the Nelway Formation (Fyles and Hewlett, 1959). The Nelway Formation dies out to the north, supplanted by more argillaceous rocks of the Lardeau Group.

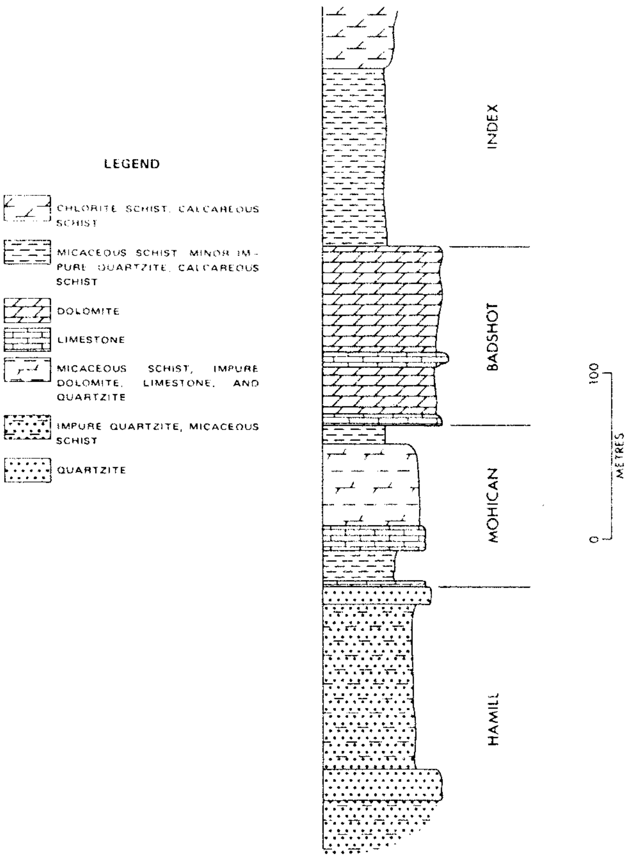

Figure 12. Upper Hamill, Mohican, Badshot and lower Lardeau composite section, Duncan area, central Kootenay Arc (after Muraro, 1962 and Fyles, 1964).

Lead-Zinc Deposits

Lead-zinc deposits in the Kootenay Arc are concentrated in two areas, the Salmo camp at the southern end of the Arc (Fyles

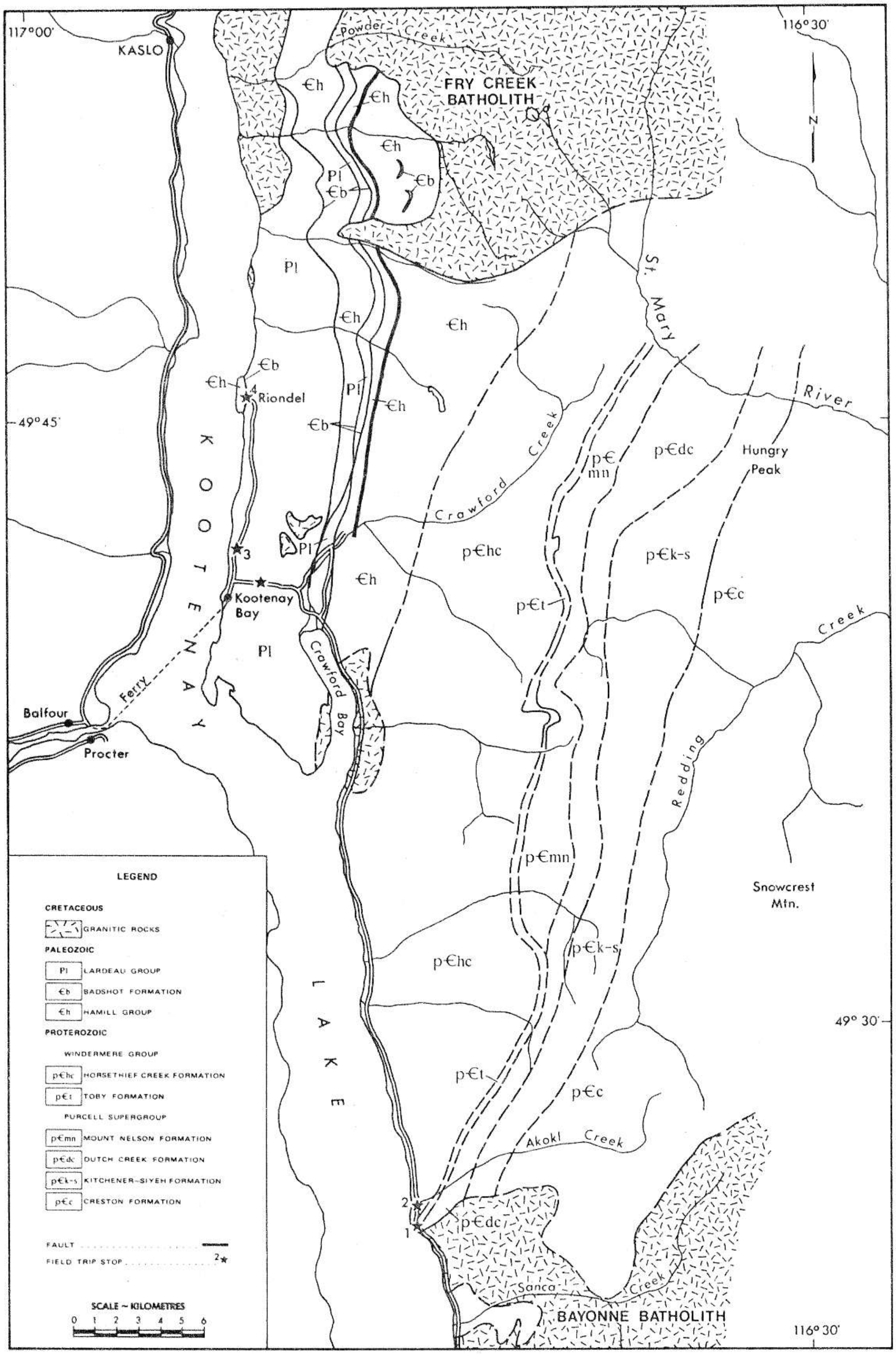

Figure 13. Geological map of the east shore of Kootenay Lake, central Kootenay Arc (after Rice, 1941; Höy, 1980).

and Hewlett, 1959; Fyles, 1970a) and the Duncan camp north of Kootenay Lake (Muraro, 1962; Fyles, 1964). The Bluebell deposit, described as a vein replacement desposit (Ransom, 1977b), is situated on the east shore of Kootenay Lake between the Duncan and Salmo camps.

Kootenay Arc deposits are hosted by intensely deformed Lower Cambrian marble or limestone. Dolomitization and associated brecciation of the limestone is common. The deposits consist of lenses, irregular bands, disseminated grains, or massive bodies in dolomite. They are irregular in outline and commonly elongated parallel to the regional structural grain. Contacts with country rock may be sharp or gradational.

The origin of Kootenay Arc deposits is enigmatic. Fyles and Hewlett (1959) describe the deposits as replacement deposits controlled by Phase 2 folds and locally, faults. They describe the close spatial association of mineralization to structures and the brecciated nature of some of the ore. Sangster (1970) and Addie (1970) describe the deposits as syngenetic, with sulphides accumulating in small basins in a deep-water carbonate platform. Muraro (1962), based on detailed studies of the Duncan deposit, suggests that Kootenay Arc deposits are hydrothermal replacement desposits, controlled by stratigraphy and formed before deformation and metamorphism.

DAY 2 (P.M.)

STRUCTURE AND STRATIGRAPHY OF THE KOOTENAY ARC FROM CRESTON TO RIONDEL

Trygve Höy

On the afternoon of Day 2, the excursion crosses obliquely the transition from western margin of the Purcell Anticlinorium to the Kootenay Arc at Riondel (Fig. 13). The transition is marked by an increase in the grade of regional metamorphism and by a noticeable change in structural style, from the broad, open north-plunging folds characteristic of the Purcell Anticlinorium to the overturned, isoclinal folds typical of this portion of the Arc. The metasedimentary units through which this portion of the field trip runs includes the Toby Formation, Horsethief Creek Group, Hamill Group, Badshot Formation and Lardeau Group (Fig. 2), as well as the discordant, post-tectonic Bayonne Batholith. The final stop of the day is at the Bluebell Mine at Riondel.

The structure of the central part of the Kootenay Arc along the east side of Kootenay Lake consists essentially of a homoclinal succession of generally west-dipping Horsethief Creek and lower Hamill strata overlain in the Riondel area (Höy, 1977, 1980) by intensely deformed upper Hamill, Mohican, Badshot and Index rocks. North-trending, generally open folds locally overturn the bedding in Horsethief Creek Group (Rice, 1941), and a pronounced foliation and lineation indicate substantial east-west flattening and an attendant north-south elongation of Horsethief Creek rocks (Lis, personal communication, 1974).

The most conspicuous structures in the Riondel area east of the Bluebell Mine are a set of very tight to isoclinal Phase 2 folds with west-dipping axial surfaces (Höy, 1977, 1980). West of a west-dipping listric reverse fault called the West Bernard fault, Phase 2 folds are imposed on a stratigraphic succession which is overturned and represents the underlimb of a large westward-closing Phase 1 nappe called the Riondel nappe (Fig. 14). The axial zone and upper limb of the Riondel nappe have been removed by erosion. The overturned limb extends under Kootenay Lake in the west and is bounded on the east by the West Bernard fault. Its western closure is deduced from regional considerations; in the Duncan Lake area parts of both the lower and upper limbs of comparable style folds are exposed and indicate western closures (Fyles, 1964).

Phase 2 folds plunge at shallow angles to the north (and locally to the south), and their axial planes dip west-ward becoming more upright to the east. A pronounced foliation and lineation are parallel to their axial planes and fold axes respectively. In the western part of the Riondel area they have developed in an inverted panel of rocks and older rocks from the cores of synforms and younger rocks the cores of antiforms. East of the West Ber-

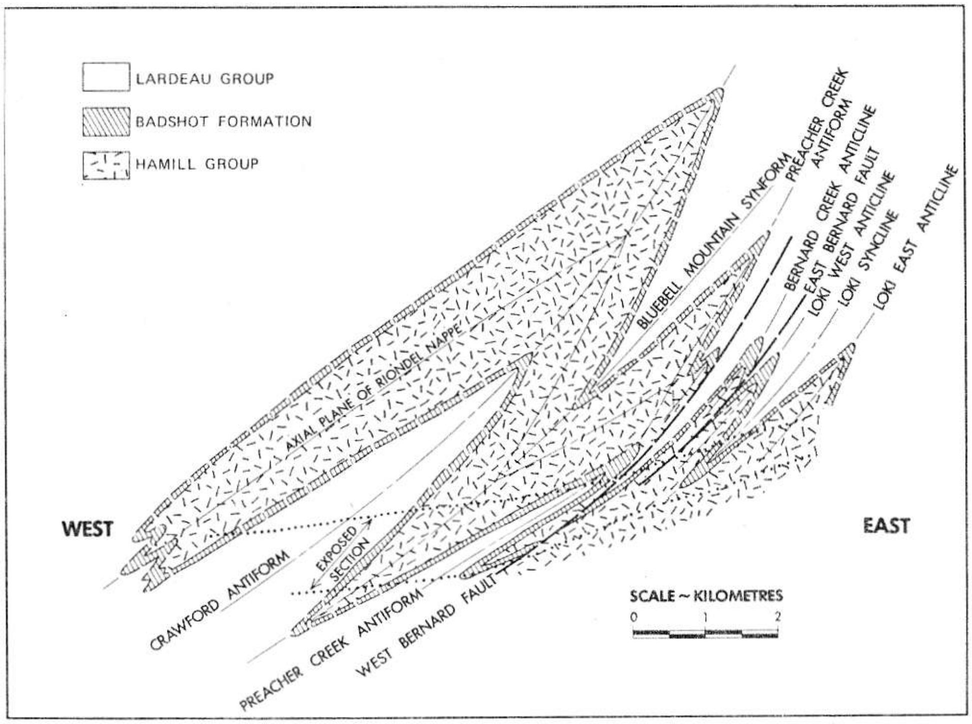

Figure 14. A schematic vertical composite section through the Riondel area (from Höy, 1980).

nard fault, the folds have developed in a right-way-up panel of rocks.

Road Log: Creston to Riondel (Highway 3)

On the afternoon of the second day the excursion continues generally up-section through the Upper Proterozoic succession on the west side of the Purcell Anticlinorium into highly deformed and metamorphosed lower Cambrian metasediments in the central Kootenay Arc. The road log begins at the town of Wynndel, 11 km north of Creston.

0.0 km Town of Wynndel.

6.3 km Contact at Bayonne Batholith with the Dutch Creek Formation. The route continues for approximately 22 km through the Bayonne Batholith. The Batholith is a discordant post-tectonic pluton that truncates several large folds in the Creston and Kitchener Formations. Its composition varies from a granite to a granodiorite, with the main bulk of the instruction being an equigranular to locally porphyritic granodiorite. It has been dated (K-Ar) at 100 m.y.

STOP 2-1, 33 km, Toby Formation, Columbia Point. The Toby Formation is exposed in a road-cut just north of the Bayonne Batholith. It consists of stretched pebbles and boulders of quartzite, dolomite, and argillite suspended in an originally argillaceous matrix. Aalto (1971) describes the Toby Formation as a tillite, deposited during widespread glaciation in early Hadrynian time. Lis and Price (1976) and Atkinson (1977) believe the Toby Formation, and similar coarse conglomerates in the overlying Horsethief Creek Group, may in part have been deposited as mudflows or as fluvial gravels in structural depressions north of prominent fault scarps.

STOP 2-2, 34.6 km, Horsethief Creek Formation. The basal part of the Horsethief Creek Formation is exposed in road-cuts just north of its contact with the underlying Toby Formation. The Horsethief Creek Group consists predominantly of slate, grit and phyllite with interbeds of grey crystalline limestone and occasional beds of quartzite and conglomerate or diamictite. The exposure, comprising grey sericite phyllite, grit and minor quartzite, is fairly typical of much of the Horsethief Creek Group.

44.0 Exposure of grey limestone, Horsethief Creek Group. This thick and prominent limestone unit near the central part of the Horsethief Creek Group is probably correlative with "the middle marble member" of the Horsethief Creek in the northern Selkirks, north of the Trans Canada Highway (Brown *et al.*, 1979). The grits, quartzite and pelite that we have just passed through are the equivalent of the "lower pelite member", and the overlying schist and quartzite, the "upper pelite member". The total thickness of Horsethief Creek sediments in the northern Selkirks is 4000 to 4500 m compared with greater than 8500 m in the Kootenay Lake area.

50.0 Exposure of conglomeratic mudstone in the upper part of the Horsethief Creek Group.

60.0 A small discordant granite pluton conceals the contact of the Horsethief Creek and the overlying Hamill Group.

The next stops are in metasedimentary rocks in the vicinity of the Bluebell deposit. The youngest exposed metasedimentary rocks on the east shore of the Lake, gneisses of the Index Formation are examined first, then the Bluebell deposit and host Badshot limestone is visited, and finally, underlying quartzite and schist of the Hamill Group.

68.5 Unit L4, the youngest member of the Lardeau Group exposed in the Riondel area (see Fig. 15), occurs within the core of an anitformal syncline called the Crawford antiform. It is a rusty-weathering paragneiss that commonly contains large boudins of amphibolite. Sillimanite and kyanite, largely altered to a white mica, are common in more pelitic layers.

70.0 Riondel road turn-off.

STOP 2-3, 70.8 km, Unit L3, Lardeau Group. The succession within this unit includes a basal quartzite interlayered with biotite schist, overlain by thinly laminated dark amphibolite, then thin rusty-weathering siliceous marble that grades upward into grey-green calc-silicate gneiss, the dominant part of the unit. The well-layered nature of the calc-silicate gneiss is evident in this exposure. A study of metamorphic reactions among calc-silicate minerals in this unit indicates that the regional metamorphic grade increases towards the west (Höy, 1976).

77.1 Unit L2. Dark grey to black hornblende gneiss and amphibolite comprise unit L2. Calc-silicate gneiss layers, quartzite layers and a number of thin calcite marble layers are also common within the unit.

79.0 Town of Riondel.

The next stop visits the Bluebell Mine (described in detail in the following paper). Exposed on Riondel peninsula are quartzites and interlayered pelitic schists and gneisses of the Hamill Group, structural footwall rocks to the Badshot marble and Bluebell mineralization.

GEOLOGY OF THE BLUEBELL DEPOSITS, RIONDEL B.C.

Trygve Höy and P. W. Ransom

Introduction

The Bluebell deposits are lead-zinc-silver replacement deposits of lower Cambrian Badshot Formation marble, located on a peninsula on the east shore of Kootenay Lake in southeastern British Columbia. Production from the property commenced in 1895 and continued intermittently until 1971 by Cominco and various other operators. Total production was 4,823,000 tonnes containing 5.2% Pb, 6.3% Zn and 60 g/tonne Ag:

	tonnes x 1000	Pb	Zn	g/tonne Ag
Pre Cominco Operators (1895-1927)	491	6.5%	8.2%	96
Cominco Ltd. (1952-1971)	4,333	5.1%	6.1%	55
Total	4,823	5.2%	6.3%	60
Nonrecoverable Sulphide	349	4.9%	5.6%	47
Total Sulphides	5,173	5.2%	6.3%	59

The general geology description in the following section is summarzied from Höy (1980). The geology of the Bluebell deposit has been described by Irvine (1957), Shannon (1970), Ohmoto and Rye (1970), Ransom (1977b) and Höy (1980). The descriptions of ore controls and sulphide mineralogy are based largely on these published reports, as well as personal communiction with E. W. Muraro of Cominco Ltd.

Regional Geology

The Riondel area is located within a local metamorphic culmination and near a regional structural culmination in the Kootenay Arc (Höy, 1976, 1977). North of the Riondel area, fold axes generally plunge to the north at low angles (Fyles and Eastwood, 1962; Fyles, 1964; Read, 1973) whereas to the south, axes most commonly plunge south (Fyles and Hewlett, 1959). Both syntectonic and post-tectonic quartz monzonite stocks are exposed in the area north of the Bluebell deposit.

The grade of regional metamorphism ranges from the greenschist facies south and east of the Bluebell deposit to the upper amphibolite facies in a narrow belt along the east shoreline of Kootenay Lake (Höy, 1976). Metamorphic isograds trend north-south approximately parallel to the dominant structural trends, although locally cut across them.

Group	Formation	Map Unit	Estimated Thickness *metres*	Description
Lardeau	Index	L4	top not exposed	micaceous schist and gneiss
		L3	400 – 450	calc-silicate gneiss, amphibolite, schist; impure marble; amphibolite layer and pure white quartzite layer near base
		L2	700	biotite-hornblende gneiss, amphibolite; minor calc-silicate gneiss, marble, and schist
		L1	150	micaceous schist
	Badshot	B	15 – 30	white crystalline calcite marble, dolomite
	Mohican	M	~50	interlayered quartzite, calcareous and micaceous schist, limestone, and dolomite
Hamill		H4	230	dark quartzite, dark fine-grained quartz-rich schist
		H3	60 – 200	massive, white quartzite
		H2	2 000	interbedded micaceous schist, quartzite, and siltstone; minor amphibolite
		H1	1 600	massive, white quartzite; gritty quartzite
Horse-thief Creek		HTC	base not exposed	fine-grained, light grey to green chlorite-muscovite schist and phyllite; rare white quartzite and marble near top

Figure 15. Table and descriptions of formations, Riondel area (from Höy, 1980).

Stratigraphy

The lithologic succession in the Riondel area is illustrated in Figure 15. In the vicinity of the Bluebell deposits on Riondel Peninsula, it comprises a north-trending and west-dipping succession of Lower Cambrian quartzite, pelitic schist, calcareous schist and marble. The succession is inverted; the older rocks of the Hamill Group outcrop along the western shoreline of Riondel Peninsula and overlie successively younger rocks of the Mohican Formation, the Badshot marble which hosts the sulphide mineralization, and schists of the Index Formation.

The upper part of the Hamill Group (unit H4, Fig. 15) consists of at least 200 metres of fine-grained, dark grey biotite-quartz schist and quartzite. Pelitic micaceous schist layers comprise at least half the Upper Hamill succession on Riondel Peninsula, and thin white quartzite layers are common. Calcareous schist, pelitic schist, quartzite, and marble characterize the stratigraphically overlying Mohican Formation. On Riondel Peninsula the Mohican is represented by a basal limestone (the Upper Limestone) and overlying micaceous schist. The Badshot Formation is a 30- to 50-metre-thick white calcite marble. Accessory minerals include tremolite, phlogopite and graphite. It is stratigraphically overlain by calcareous and pelitic schist and gneiss of the Index Formation (unit L1). These rocks are the youngest metasedimentary rocks on Riondel Peninsula, and comprise the footwall of the Badshot-hosted sulphide ore.

A variety of lithologies within the Badshot marble at Bluebell and in the Hamill and Lardeau Groups in the western part of the Riondel area are called pegmatite. They range from fine-grained chloritic rock to very coarse-grained rock consisting entirely of feldspar (albite?). It has been suggested that some of these rocks are metasediments. The most extensive in the mine area is the hanging wall pegmatite, a laterally continuous sheet 1 to 3 metres thick, 1 to 10 metres below the hanging wall of the marble (Fig. 16).

East-dipping early lamprophyre dykes about 1 to 2 metres thick have been broken into long segments 5 to 20 metres high by bedding parallel faulting within the marble (Fig. 16). Later dykes, referred to as diabase, greenstone and lamprophyre, are grouped with Cenozoic age lamprophyres mapped regionally. These dykes are close to vertical and generally strike west to northwest. They range in thickness from a few centimetres to 3 metres and cut both metasediments and pegmatite (Irvine, 1957). The dykes are prophyritic containing phenocrysts of plagioclase (labradorite to andesine), olivine, pyroxine, hornblende and biotite which have been largely replaced by calcite, epidote, chlorite and magnetite (Ohmoto and Rye, 1970).

Structure

The structure of the Riondel area is dominated by a large, recumbent Phase 1 anticlinal fold, the Riondel nappe (Fig. 14). It closes west of the Riondel area, probably beneath Kootenay Lake. The rocks in the western part of the Riondel area are the inverted stratigraphic panel of the lower limb of the nappe. North-trending isoclinal Phase 2 folds are superposed on this panel. The Bluebell deposit is on the western limb of a Phase 2 fold, the Crawford antiform, that has in its core the youngest rocks within the map-area. Hence the mine section is inverted and Hamill Group structurally overlies Badshot and Lardeau. The intense regional metamorphism culminated during the Phase 2 deformation and many of the coarse-grained granitic pegmatite bodies are in-situ metasomatic "sweats". Others were mobilized and in-

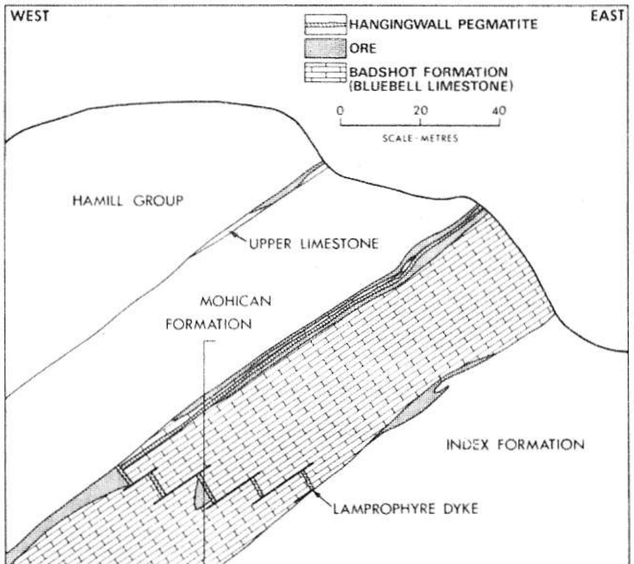

Figure 16. Idealized sections through the Bluebell Mine (after Ransom, 1977).

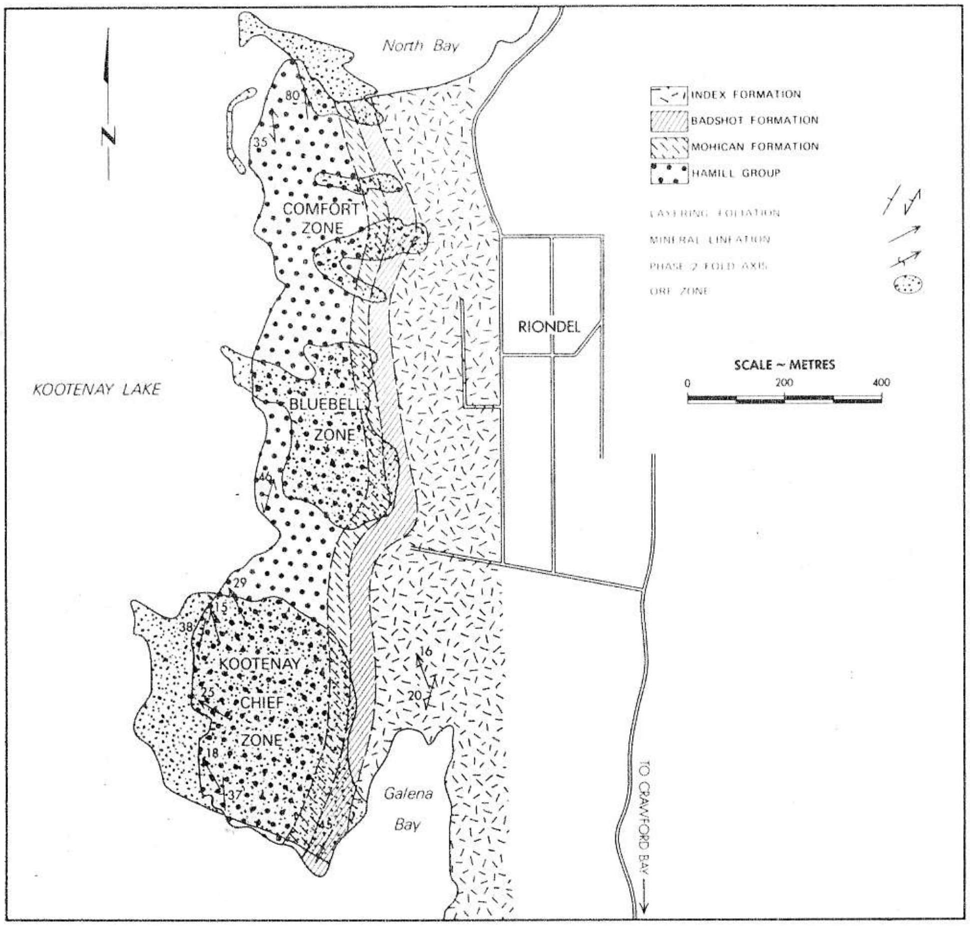

Figure 17. Sketch map of Riondel Peninsula showing distribution of formations and Bluebell ore zones, projected to surface; modified from Shannon (1970).

truded the metasediments during various stages of development of Phase 2 folds.

Phase 3 folds warp the limbs of earlier folds. They are not conspicuous on a regional scale, but are common as fairly open to tight S-shaped folds in the vicinity of the Bluebell deposits.

Orebodies

The Bluebell ore deposits consist of three main zones spaced approximately 500 metres apart along the strike of the Badshot marble: the Comfort zone at the north end of Riondel Peninsula, the Bluebell zone in the centre, and the Kootenay Chief at the south end (Fig. 17). The zones are localized along steep cross-fractures that trend west/northwesterly (north 62 degrees west to north 75 degrees west) and dip 80 to 90 degrees north (Irvine, 1957). Idealized sections through the Bluebell mine are illustrated in Figure 16. Within the zones are tabular ore shoots that are transverse to the bedding and plunge westward following the intersection of the fractures with the marble. The ore shoots: "ranged in size from irregular pods of a few thousand tons to continuous masses of up to one million tons that extended down-dip as much as 500 m. In cross-section, an average ore shoot was mushroom-shaped, the stem representing cross-cutting keels 1 to 30 m wide and the cap representing a bedding-conformable horizon up to 6 m thick that extended laterally as much as 50 m from the keel zone. The keel zones extended below the conformable ore some 10 to 20 m, narrowing and grading into a series of steep mineralized fractures that become uneconomical to mine. Some of the fracture zones and keels of the larger ore shoots extended to the footwall. A few ore shoots also developed along the footwall in a style complementary to hanging wall ore shoots. Depressions along the footwall and arches along the hanging wall were particularly favourable areas for ore accumulations. Ore shoots more than doubled in thickness and spread out laterally as 'runs' along strike on the down-dip side of displace lamprophyre dyke segments. Very little ore occurred on the up-dip side of these segments and not until 30 m further up-dip did ore attain normal thicknesses" (Ransom, 1977b, p. 41).

The main ore hanging wall is the Mohican schist; however, the 'hanging wall pegmatite' capped ore as well. Within ore shoots the interval between this pegmatite and the Mohican schists was entirely ore in the keel zones but only the top portion contained ore away from the keels (Fig. 16).

The late lamprophyre or diabase dykes appear to have controlled the location of some ore. Where such dykes intersected mineralized fractures or keel zones ore tended to be localized on the underside of these dykes (Fig. 16). Mine geologists considered this relationship and the accumulation of ore on the west sides of the early faulted lamprophyre dyke segments as compelling evidence for a hydrothermal replacement origin of the deposit.

The central Bluebell zone is separated from the Comfort zone to the north and the Kootenay Chief zone to the south by approximately 300 metres of barren marble. Five ore shoots within the zone occur as "stubby keels" that follow cross-fractures in the Badshot marble for a small stratigraphic distance beneath hanging wall Mohican quartzite (Irvine, 1957, p. 100). The shoots occupy a strike length of approximately 220 metres and were mined 600 metres down-dip from the surface.

Ore shoots in the Comfort and Kootenay Chief zones generally extended further into the Badshot marble and occupied a greater stratigraphic length. In the Comfort zone the five orebodies were each separated by barren marble (Irvine, 1957).

Ore Mineralogy. The mineralogy of the most southern of the ore zones, the Kootenay Chief zone, is described in detail by Westervelt (1960). Two types of mineralization are identified, referred to as the knebelite (an Fe-Mn olivine) zone and the siliceous zone.

Sphalerite, occurring either in coarse-grained masses or in veins, is the most abundant sulphide in the knebelite zone. Galena, and less commonly arsenopyrite and chalcopyrite, occur as scattered grains enclosed within the sphalerite or in other sulphides, and pyrrhotite is common as disseminated flakes and blebs within a siliceous gangue. The gangue is a highly altered rock containing abundant chlorite, and knebelite largely altered to chlorite, serpentine, and carbonate. Minor quartz, rare magnetite, and varying amounts of carbonate are also present in the knebelite zone gangue.

The siliceous zone is characterized by an abundance of quartz and pyrrhotite, and the development of spectacular crystal growths in vugs and cavities. Pyrrhotite occurs as large well-developed crystals and as masses within other sulphides. Dark sphalerite and coarsely crystalline galena are abundant. Chalcopyrite, arsenopyrite, pyrite, and marcasite occur in small euhedral crystals or intergrown with each other or other sulphides. Coarsely crystalline quartz and carbonate are the common gangue minerals in the siliceous zone.

Oxidation and Thermal Waters. Oxidation of the Bluebell ore is common, and occurred to depths in excess of 300 metres below the surface (Shannon, 1970). The sequence of oxidation appears to be:

(1) alteration of pyrrohotite to lacy or spongy pyrite.
(2) alteration of pyrite to hematite or limonite.
(3) oxidation of arsenopyrite, sphalerite, and then knebelite.

An unusual geological feature of the Bluebell mine was CO_2 charged thermal water which flowed from fissures encountered at depth and commonly produced various forms of $CaCO_3$ deposits in the mine workings.

Origin of the Bluebell Deposit. The alignment of ore shoots with steep tensional cross-fractures, the crosscutting nature of the ore shoots, ore-dyke relationships, and the occurrence of sulphides in both Badshot marble and structurally overlying marble in the Mohican Formation argue strongly that the deposits formed as fracture-controlled replacement bodies. The common occurrence of coarsely crystalline sulphide minerals, associated with well-formed quartz crystal clusters in numerous vugs and cavities in the siliceous zones, is evidence of late deposition, post-regional metamorphism and deformation.

A detailed study of fluid inclusions in ore and gangue minerals of the Bluebell deposit by Ohmoto and Rye (1970) showed that the late development of crystals in vugs was associated with the saline brines of probably meteoric rather than magmatic origin, and that temperatures of 320 to 450 degrees centigrade were indicated. The temperature and salinity of fluids associated with earlier deposition of massive sulphide-quartz-carbonate ores were both probably higher. They also concluded that the depth of the Bluebell marble at the time of ore deposition was probably 6 ± 2 kilometres.

Lead isotope data (LeCouteur and Sinclair, in preparation; Reynolds and Sinclair, 1971) indicate that the lodes of the Bluebell mine have a common origin with the vein deposits in the Ainsworth camp.

In summary, the bulk of the deposits were emplaced as late fracture-controlled replacement bodies within the Badshot and, to a lesser extent, Mohican marbles after the intense Phase 2 deformation of Middle Jurassic age. However, Muraro (personal communication, 1979) described stratiform lead-zinc mineralization in the Bluebell deposit suggesting that the obvious late fracture-controlled mineralization may in large part represent in situ remobilization of an older stratabound deposit.

DAY 3

GEOLOGY OF THE DUNCAN DEPOSIT AND THE DUNCAN LAKE AREA, KOOTENAY ARC

T.W. Muraro and T. Höy

Road Log: Nelson to Duncan Lake (Highway 31)

On the morning of Day 3 the excursion follows the west shoreline of Kootenay Lake northward to the Duncan deposit at Duncan Lake. A number of stops in the Duncan area will examine the detailed stratigraphic succession in the vicinity of the deposit, exposures of lead-zinc mineralization, and core from the deposit. In the afternoon we will retrace our route southward to Kaslo, a small town within the Ainsworth silver camp, then cross the Slocan Ranges to New Denver on Slocan Lake, passing through the historical Slocan mining camp. The route continues northward along the western margin of the Kootenay Arc and the eastern margin of the Shuswap Complex to an overnight stay at Revelstoke.

STOP 3-1, Highway 31, overlooking north end of Kootenay Lake and south end of Duncan lake area. The stop allows a view across Kootenay Lake to the intensely deformed Lower Paleozoic succession at the north end of Kootenay Lake, the Fry Creek batholith to the south, and the Sal Prospect on Salsbury and Bulme Creeks.

Mineralization on the Sal property occurs in the upper and lower dolomite members of the Badshot Formation, on the east limb of the Duncan anticline (Fyles, 1964). Three zones were discovered. Sal A zone consists of fine-grained pyrite with minor sphalerite and galena in narrow bands of white crystalline limestone in grey dolomite. One zone, 5-6 metres thick and about 100 metres in length, occurs along the contact between dolomite and siliceous dolomite. Sal B zone consists of bands of fine-grained disseminated pyrite with minor sphalerite and galena in white crystalline limestone separated by grey dolomite. The bands are up to a metre thick and extend across a 10 to 12 metre stratigraphic interval. The Sal C zone is similar to the B zone, occurring also in the lower dolomite.

STOP 3-2, Exposure of Mohican section, Glacier Creek. A significant portion of the Mohican Formation is exposed in Glacier Creek. It represents a facies somewhat distinct from Mohican facies to the west and east (see Fig. 18). Particularly noteworthy are the number of quartzite and dolomite layers within this section, and within other Mohican sections along the east limb of the Duncan anticline. The complete Mohican section is summarized below (from Muraro, 1962), and many of the individual units are represented in the Glacier Creek exposure.

Badshot Formation
unit 19: Limestone — white, medium grained recrystallized limestone

Mohican Formation
unit 18: Greenschist — muscovite quartz schist, minor limestone, some garnet, chloritoid and pyrrhotite in places
unit 17: Dolomite — light grey to white, vague grey banding
unit 16: Quartzite — white near top; light brown, argillaceous, calcareous, laminated
unit 15: Dolomite — light grey, massive, fine grained, some vague irregular banding
unit 14: Quartzite — brown, micaceous, calcareous; laminated

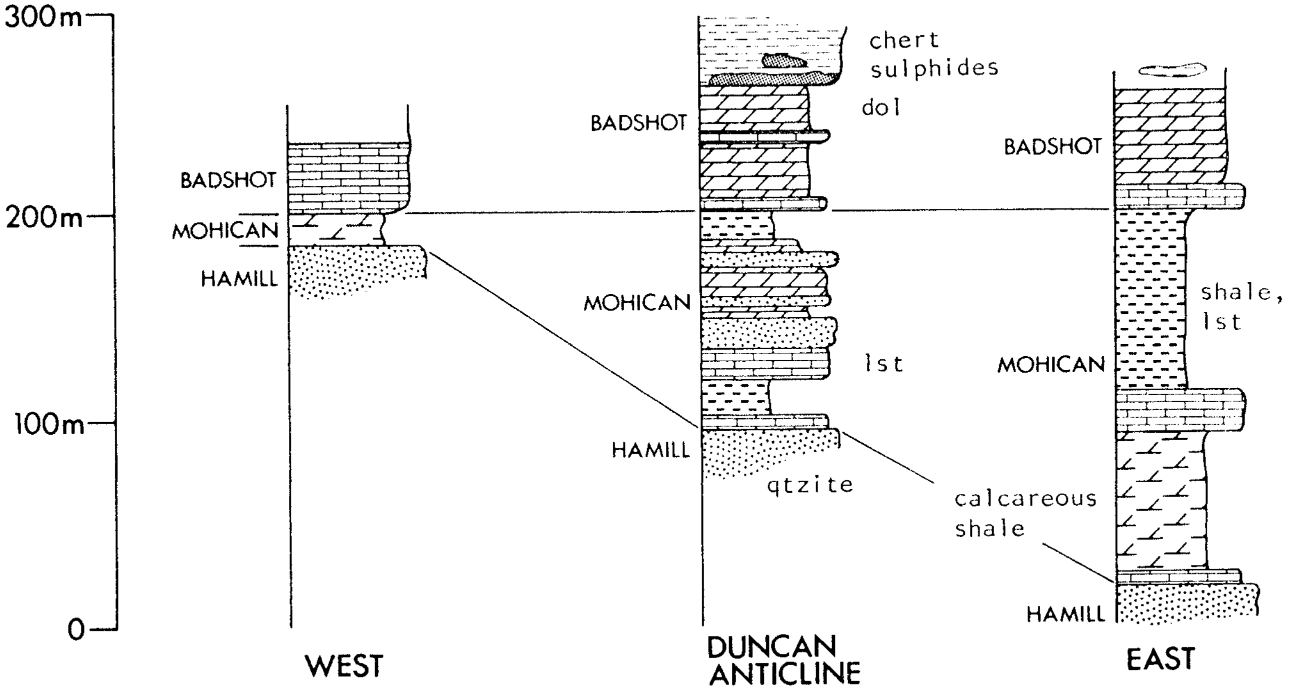

Figure 18. Badshot and Mohican Formations facies changes in the Duncan Lake area (adapted from Muraro, 1962 and Fyles, 1964).

unit 13: Dolomite — light grey to white; massive fine grained
unit 12: Quartzite — brown, micaceous, calcareous; laminated
unit 11: Dolomite — light grey to white; blue grey with vague irregular banding; fine grained
unit 10: Quartzite — brown, micaceous, calcareous; phyllitic near base
unit 9: Phyllite — dark brown, dense, calcareous
unit 8: Dolomite — grey and white streaked, fine grained
unit 7: Limestone — grey and white; medium grained
unit 6: Schist — dark green, finely crenulated biotitic quartz mica schist
unit 5: Schist and Limestone — light brown and red weathering quartz mica schist (muscovite); biotite porphyroblasts common
unit 4: Limestone — impure, thin bedded, micaceous, grey to cream, recrystallized

STOP 3-3, Duncan property, Duncan Lake. The stop at the Duncan property is at "zone No. 5" where mineralized outcrop and core can be examined.

GEOLOGY OF THE DUNCAN DEPOSIT

Properties in the Duncan area were located in 1925 and have been drilled extensively since then. The Consolidated Mining and Smelting Company of Canada, Limited (Cominco) optioned the Duncan property in 1957 and subsequent geological mapping, diamond drilling and underground exploration have outlined a small, low-grade deposit (approximately 9 million tonnes containing 2.7% Pb and 2.9% Zn). The property and surrounding geology have been described in detail by Muraro (1962) and Fyles (1964) and the following descriptions are summarized from these reports.

Stratigraphy

The stratigraphic succession in the vicinity of the deposit is illustrated in Figure 12. Mineralization is confined to a number of zones within the Badshot Formation in the complexly deformed and faulted hinge zone of the Duncan anticline.

The Mohican Formation, a heterogeneous succession of dominantly calcareous rocks, underlies the Badshot Formation. In the vicinity of the deposit it comprises a well-layered and variable sequence of massive to laminated limestone and dolomite, micaceous and calcareous quartzite, dark brown calcareous phyllite, and light brown or dark green micaceous schist. Individual layers can be traced along strike to the north and south, and the mine sequence can be readily matched with the Glacier Creek sequence examined earlier. However, facies changes within the Mohican to the east and west are apparent; eastern facies are thicker and more argillaceous and western facies are also more argillaceous, but considerably thinner (Fig. 18) (Muraro, 1962; Fyles, 1964).

The Badshot Formation at the Duncan property consists of a lower and upper dolomite separated by a thin layer of crystalline limestone (Fig. 18). The following description is taken from Fyles (1964, p. 24):

"The uppermost (part of the Badshot), as much as 200 feet thick, is a dark-grey to black very fine-grained siliceous dolomite. Commonly it has a strong lineation plunging northward parallel to the axes of minor folds; elsewhere it has an irregular black and white banding. The siliceous dolomite is underlain by about 50 feet of dark-grey and white flecked and banded dolomite, and the siliceous and underlying dolomite together are referred to as the upper dolomite. Beneath this dolomite is 5 to 20 feet of light-grey medium- to fine-grained crystalline limestone that commonly has a thin bed of grey micaceous phyllite at the base. Another thin member of dark-grey dolomite, called the lower dolomite, underlies the limestone, which in turn is underlain by a few feet of grey to white crystalline limestone at the base of the formation."

Mineralization in the Duncan deposit, and in most other deposits within the Kootenay Arc, is hosted by a dolomitic phase of the Badshot Formation. In the Duncan camp, mineralization and associated dolomitization is largely confined to a north-south belt coincident with the Duncan anticline. Two types of dolomite are known: a textured dolomite and a massive dolomite (Fyles, 1964). Textured dolomite contains black or dark-grey flecks, streaks, or bands outlined by concentrations of carbon which may give the rock a well-developed foliation or lineation.

"Massive dolomite is most commonly light grey and weathers cream coloured, buff, or brown. . . . In the Duncan anticline much of the Badshot dolomite and some of the Mohican dolomite is textured. Away from the Duncan anticline most of the dolomite is massive, though some lenses in both the Badshot and Mohican are textured. Dolomite associated with the mineralized zones in general is textured." (Fyles, 1964, p. 68).

Siliceous dolomite that grades to a relatively pure, fine-grained dark quartzite, that is interpreted to be chert (Muraro, 1962), is common in the upper part of the Badshot Formation. Its areal distribution in the Badshot is similar to that of dolomite, largely confined to a north-south belt coincident with the Duncan anticline.

The Badshot Formation is overlain by a thick succession of fine-grained dark grey and green schists named the Index Formation (Fyles, 1964). On the Duncan anticline and to the east a few metres of dark coloured carbonaceous argillite immediately overlies the Badshot limestone or dolomite. In general, the Lower Index comprises grey, dark green or black carbonaceous and siliceous argillites, calcareous mica schists, slate, dark grey limestone, and minor quartzite. The Upper Index consists dominantly of fine-grained green chloristic schist.

Structure

A complex pattern of superposed folds, charcteristic of the Kootenay Arc, dominates the structure of the Duncan area (Fyles, 1964). Isoclinal Phase 1 folds, including the Duncan anticline, plunge to the north at low angles. More open, north to northwest plunging Phase 2 folds deform the limbs and axial surface of Phase 1 folds. They are visible in many outcrops, in contrast to Phase 1 folds which generally cannot be seen but must be reconstructed from studies of the distribution of rock sequences and the mapping of formations (Fyles, 1964). The Duncan anticline plunges to the north "and Duncan Lake is thought to cover the place where the hinge zone in the Badshot and Lower Index Formation plunges beneath the surface" (Fyles, 1964, p. 58). Mineralization at the Duncan property is on the east limb of the anticline, just east of its hinge. Relatively tight asymmetric or overturned Phase 2 folds, including the anticlinal hinge zone shown in Figure 19, are the most common structures observed on the property.

Mineralization

The Duncan deposits include a number of sulphide zones located in the complexly deformed and faulted hinge zone or east limb of the Duncan anticline.

"The deposits occur as sheets which parallel the bedding. Galena-sphalerite-pyrite mineralization in the Upper dolomite commonly occurs along the dolomite-chert contact. Sphalerite-pyrite mineralization in the Lower dolomite is commonly associated with a relatively dark grey stratigraphic zone. . . . Dolomite or dolomite and quartz (chert) — the host rocks — constitute most of the non-sulphide portion of the deposits. . . . gangue minerals are calcite and very minor amounts of dolomite and quartz. The mineralogy is simple. Pyrite, sphalerite and galena, in order of abundance, are the principal sulphides. Minor sulphides are pyrrhotite, chalcopyrite (minute amounts widely distributed) and individual minute occurrences of marcasite (with the pyrrhotite), ruby silver, and menaghenite. Silver values are negligible. Sulphide veins do not occur in the deposits. The ore minerals are characteristically fine grained and disseminated in calcite and dolomite. The uniform granular nature of the pyrite is due to shearing and deformation which accompanied the shear folding in the rocks. Local, relatively undisturbed parts of the deposits commonly exhibit a distinct persistent relation which is characterized by a layer of sphalerite and pyrite sandwiched between unaltered, fine grained, sheared dolomite (host rock) on one side and medium to coarse grained, white, foliated calcite (gangue) on the other side. Trace elements in pyrite indicate that it was deposited under low intensity conditions and that it is possibly of hydrothermal origin." (Muraro, 1962, pp. 154-155).

Nos. 1 and 2 zones outcrop just north of Glacier Creek. No. 3 zone outcrops just south of the Lower Arm of Duncan Lake and No. 4 zone on the south end of the Duncan Peninsula. Nos. 5, 6, 7 and 8 zones are in the vicinity of the Duncan mine, and the latter three are projected onto the schematic vertical section of Figure 19.

The following descriptions of the zones are taken largely from the descriptions given by Muraro (1962). They are quoted from Fyles (1964, p. 74).

"No. 7 zone is a steeply dipping tabular body averaging 15 to 20 feet (5 - 6.5 m) thick along the western contact of the siliceous dolomite. The zone as indicated by drilling plunges about 7 degrees to the north and is about 400 feet (120 m) high. It has been followed for 3,000 feet (900 m) in the drift and found in drilling beyond. The zone is layered, with a western layer in which dolomite, pyrite, and sphalerite are found in fairly well-marked bands; a central layer with lenticular masses of pyrite, galena, and

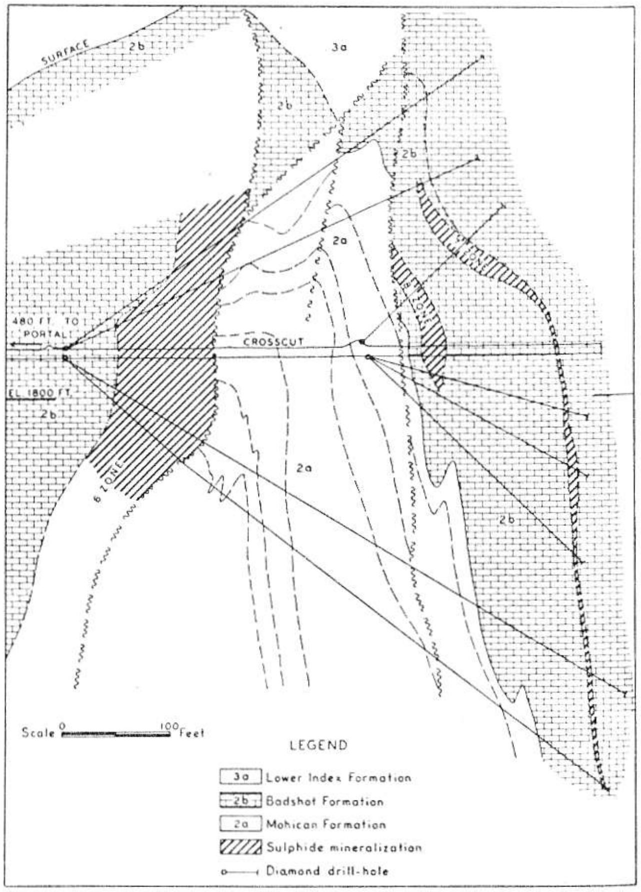

Figure 19. A schematic vertical section through the Duncan deposit (from Fyles, 1964).

sphalerite in carbonate layers associated with fine-grained quartz; and an eastern siliceous layer in which pyrite and sphalerite are the dominant sulphides. Some bands of sulphides within the layers follow small discontinuous, nearly isoclinal folds which plunge to the north at low angles. Bands of sulphides are a fraction of an inch to a few inches thick, and the grains of sulphides within them are generally less than 1 millimetre across.

No. 5 zone is below and to the south of No. 7 zone along the same western contact of the siliceous dolomite. It has the same plunge as No. 7 zone and is separated from it by a zone along the contact about 200 feet (60 m) high in which there is only scattered sulphide mineralization.

No. 8 zone is a relatively small lens in the upper dolomite about 100 feet (30 m) west of No. 7 zone. It dips at moderate angles to the east and, although not fully outlined, is 300 to 400 feet (90-120 m) high parallel to the dip. It plunges to the north and appears to be offset on a steeply dipping strike fault above the main crosscut. Pyrite and sphalerite are the main sulphides, and galena has been found only in polished sections.

No. 6 zone is 300 to 400 feet west of No. 7 zone and is the most westerly and the largest zone found in the mine. The dominant sulphide is pyrite, with minor amounts of sphalerite and galena. Pyrrhotite is present locally in bands an inch to a few inches wide. The zone is lenticular in cross-section, approximately 300 feet high and 20 to 100 feet thick. The zone has been found in drilling for 3,000 feet along the plunge which is at low angles to the north, parallel to that of the other zones. The zone in the main crosscut is bounded on the east and probably offset by a westerly dipping fault (see Fig. 20). Most of the mineralization is uniformly fine-grained pyrite with varying small amounts of galena and sphalerite disseminated in closely spaced thin lenses or bands in siliceous dolomite. The siliceous dolomite appears to form a tight syncline. Pyrite near the fault on the eastern side locally forms rounded clusters resembling a sheared breccia. In the trough of the syncline it occurs in massive layers associated with limestone and siliceous dolomite."

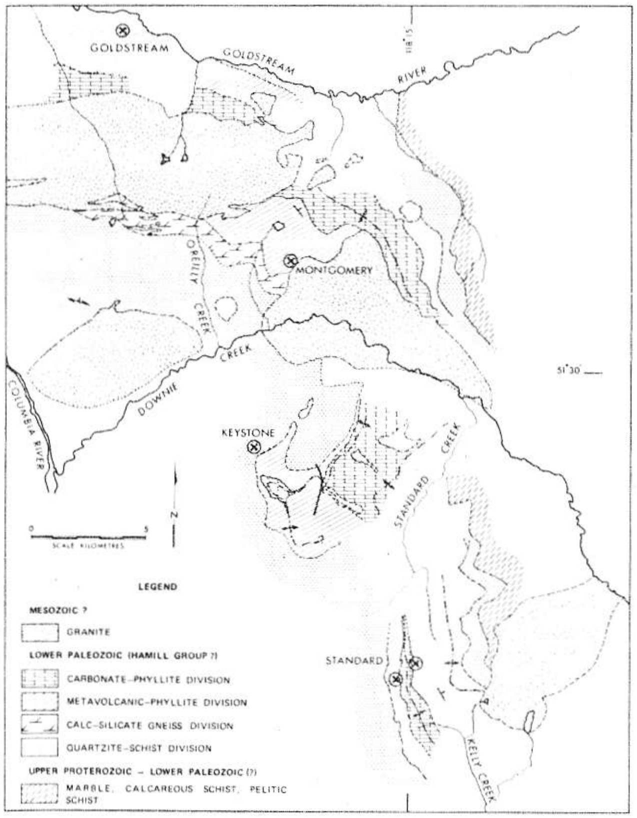

Figure 20. Regional geology of the Goldstream area (from Höy, 1979).

Muraro (1962) concludes that the deposits in the Duncan area were originally emplaced in relatively flat-lying strata with the dominant control stratigraphic and not structural. Lead-zinc zoning in the deposits is consistent with stratigraphic tops regardless of the structural position; zinc is concentrated in the stratigraphically lower dolomite, and lead and zinc in the upper dolomite. The mineralization appears to have been metamorphosed and folded along with the country rocks (Muraro, 1962, 1966). Faults associated with folding cut the deposits, and microscopic studies of pyrite show evidence of granulation and deformation.

Ainsworth and Slocan Camps

The geology of the Ainsworth-Kaslo area and the Ainsworth silver-lead-zinc Camp is described by Fyles (1967) and lead isotope data on the Ainsworth and Slocan Camps by Reynolds and Sinclair (1971) and LeCouteur and Sinclair (in preparation). The Ainsworth area is underlain by intensely deformed Lower Cambrian to upper Triassic rocks of the Lardeau, Milford, Kaslo and Slocan Groups. More than 50 properties have produced in excess of 700,000 tonnes of ore since production initially began in 1889. The deposits are small; the largest stopped areas are about 200 metres along strike, 250 metres parallel to the dip, and 1 - 2 metres thick. They are mainly simple quartz-carbonate veins containing shoots and lenses of galena, sphalerite, pyrite, and locally pyrrhotite. Fyles (1967, pp. 50-51) concluded that:

1) there are three dominant vein attitudes striking generally north, northwest, and west/northwest and dipping variably toward the west or southwest;

2) "the vein fractures have been the locus of small and repeated movements"; and

3) the fracture system and mineralization post-date the regional deformation, metamorphism, and granitic intrusion.

The geology of the Slocan Camp is described by Hedley (1952). Deposits in the camp are in veins associated with faults that crosscut deformed argillites, quartzites, and limestones of the Triassic Slocan Group.

DAY 4
GOLDSTREAM DEPOSIT

Trygve Höy

On the morning of Day 4 the excursion follows Highway 23 northward to Goldstream River then east along the south side of Goldstream River to the Goldstream deposit. Just north of Revelstoke we detour above the B.C. Hydro Revelstoke dam which is presently under construction. The Columbia River on the west side of Highway 23 follows closely the trace of the Columbia River fault. West of the fault are metamorphic rocks of the Shuswap Complex and on the east side are the deformed lower Paleozoic rocks of the Selkirk terrain.

On the afternoon of Day 4 the excursion returns to Revelstoke, then follows the Trans Canada eastward to Calgary. This will provide a cross-section through the Rocky Mountain Thrust Belt, and a brief view stop of the Monarch-Kicking Horse deposits, lead-zinc deposits in Middle Cambrian platformal carbonate rocks at Field, B.C.

Introduction

The Goldstream deposit is a stratabound massive sulphide layer in metasedimentary rocks of probable late Proterozoic to Cambrian age (Höy, 1979b). It is located at an elevation of 700 metres just south of Goldstream River in the Selkirk Mountains in southeastern British Columbia. The deposit is accessible by road from Revelstoke, 90 km to the south.

The Selkirk Mountains in the vicinity of the deposit are rugged and exploration and geological mapping are difficult. Valleys are filled with till and covered with thick underbrush, and rock exposures are rare. Above treeline, at 1800 and 1950 metres elevation, exposures are abundant although precipitous cliffs, névé, and glaciers hamper exploration and geological mapping.

Mineral exploration in the Goldstream area dates back to 1865 with the discovery of placer gold on Carnes and French Creeks. In 1866 the town of Kirbyville was founded near the mouth of the Goldstream River and by the end of that season it is estimated that there were between 8,000 and 10,000 people in the region (Gunning, 1928). Interest in lode mining increased in the late 1800's resulting in renewed exploration in the Goldstream area and discovery in 1895 of Montgomery, Standard, and Keystone, copper-zinc deposits in metasedimentary and basic volcanic rocks south of Goldstream.

The Goldstream deposit was located in 1973 by Gordon and Bruce Bried and Frank E. King. Noranda Exploration Company, Limited optioned the property in December 1974 and in 1975 drilled 50 holes outling a deposit with reserves of 3.175 million tonnes grading 4.49 per cent copper, 3.123 per cent zinc, and 20 grams per tonne silver. In 1976 an adit was driven south to the mineralized zone, and an east-west drift developed along the zone. Continued feasibility studies through to 1979 has resulted in a decision by Noranda to bring the property into production.

Regional Geology

The Goldstream area is within the Big Bend map sheet of Wheeler (1965). The geology of the area has been described by Lane (1977); Gibson, Hughes and Bradish (1977); and Brown, Höy and Lane (1977). The information in this paper is largely summarized from Höy (1979b).

The Goldstream area is underlain by dominantly miogeoclinal rocks that were deposited along the eastern margin of cratonic North America in late Proterozoic to early Paleozoic time (Wheeler, *et al.*, 1972). These rocks were deformed in Jurassic time, and intruded by granitic rocks of probable Cretaceous age.

The regional structure of the Selkirk Mountains is dominated by a northwest trending anticlinal fan structure called the Selkirk fan structure (Wheeler, 1963; 1965). Northeast of the axis of the fan structure, the regional foliation and axial surfaces of overturned folds dip southwest, whereas southwest of the axis, structures dip to the northeast (Wheeler, 1963; 1965). For a discussion on the nature and evolution of the Selkirk fan structure refer to Brown and Tippett (1978a, 1978b) and Price (1978).

The Goldstream area, southwest of the Selkirk fan axis, is dominated by tight to isoclinal north-trending folds with nearly horizontal to steeply east-dipping axial surfaces (Höy, 1979b). The grade of regional metamorphism ranges from greenschist facies in the south to lower amphibolite facies in the region of the Goldstream pluton in the north.

Stratigraphy

Metasedimentary rocks in the Goldstream area have been subdivided into five major lithologic packages (Höy, 1979b). These include dominantly pelitic and calcareous schists and marble exposed east of the Goldstream deposit (Fig. 20) that are tentatively correlated with the late Proterozoic Horsethief Creek Group (Brown, Höy and Lane, 1977; Brown, Tippett and Lane, 1978). They are overlain by a succession of rocks that has been subdivided into four main divisions (Fig. 20):

(1) dominantly pelitic phyllite and quartzite of the lower 'quartzite-schist' division;

(2) dominantly calcareous rocks of the 'calc-silicate gneiss' division;

(3) greenstone, amphibolite, dark calcareous phyllite, and carbonate of the 'metavolcanic-phyllite' division;

(4) limestone, dolomite, marble, calcareous phyllite, and micaceous phyllite of the upper 'carbonate-phyllite' division.

Based on a number of stratigraphic tops recognized in the Goldstream area, the sequence of rock units outlined above is believed to represent an original stratigraphic succession, with rocks correlated with the Horsethief Creek Group being the oldest, and the dominantly calcareous rocks of the 'carbonate-phyllite' division, the youngest.

Metasedimentary rocks of the 'quartzite-schist' division south of the Goldstream pluton (Fig. 20) comprise in excess of several thousand metres of pelitic schist interlayered with micaceous quartzite, massive thick-bedded pure to micaceous quartzite, and thin-bedded interlayered pelitic schist, rusty-weathering hornblende gneiss, and calcareous phyllite.

Thin-bedded, rusty-weathering calcareous phyllite and quartzite, pure to siliceous marble, and biotite gneiss of the 'calc-silicate gneiss' division overlie the dominantly siliceous rocks of the 'quartzite schist' division. These are exposed along the southern margin of the Goldstream pluton.

The 'metavolcanic-phyllite' division consists of massive greenstone units, cholorite phyllite, ultramafic pods, and dark calcareous to pelitic schist. The more prominent mineral occurrences, including the Goldstream deposit, occur within metasedimentary and metavolcanic rocks of this division.

The most prominent metavolcanic unit is a massive, fine- to medium-grained greenstone that is composed primarily of chlorite, actinolite, epidote, plagioclase, and minor carbonate. Analyzed samples from within the Goldstream area and to the east are within or along the edge of the "basalt" field of Church's (1975) triaxial oxide plot (Lane, 1977; Höy, 1979b). The greenstone is intercalated with chlorite phyllite, dark calcareous to pelitic schist, and, north of Keystone Peak, with greenstone that has well-developed though deformed pillow structures. The massive greenstone is generally not at a discrete stratigraphic horizon; rather it is a series of lenses that thin and thicken along strike and commonly grade laterally and vertically to chlorite phyllite that may originally have been pyroclastic or volcaniclastic rocks.

In the Standard area (Fig. 20) coarse-grained ultramafic talc-chlorite-serpentine dolomite pods overlie grey limestone and are overlain by a coarse-grained intrusive diorite and massive greenstone.

Dark green to black calcareous phyllite is common in the 'metavolcanic-phyllite' division. It has a well-defined foliation outlined by micaceous minerals, dark carbonaceous material (graphite?) and the alignment of clear quartz eyes. Grey limestone and discontinuous thin chlorite phyllite layers are common in the calcareous phyllite unit.

Dolomite, limestone and calcareous phyllite of the 'carbonate-phyllite' division overlie rocks of 'metavolcanic-phyllite' division. The carbonates are typically grey to buff coloured, thin-bedded limestone or dolomite interlayered with less pure rusty weathering calcareous schist, biotite schist, and less commonly chlorite schist. Both Keystone and Downie Peaks (Fig. 20) are composed of massive limestone, considerably thickened in the fold cores.

Regional Correlations. Based on lithologic similarities, rocks in the area south of Goldstream River have been correlated by Wheeler (1965) with the Lower Paleozoic Lardeau Group. Those along the eastern edge of the map-area have been correlated with a more calcareous western facies of the upper part of the Proterozoic Horsethief Creek Group of Hadrynian age (Brown, Höy and Lane, 1977), and the overlying package of psammites, grits, pelites, and metavolcanic rocks, with the Eocambrian Hamill Group (Höy, 1979b). Rocks of the 'metavolcanic-phyllite' division and the bulk of the overlying 'carbonate-phyllite' succession may correlate with the upper part of the Hamill Group or the Mohican Formation, and the thick carbonate that is exposed in the core of the Downie and Keystone antiforms may be the Badshot Formation (Höy, 1979b).

Structure

Tight to isoclinal east-dipping to recumbent folds dominate the structure of the Goldstream area. They have well-developed axial plane schistosities and variable, but generally northeast trending fold axes. A number of stratigraphic top determinations in graded grit beds indicate that younger metasedimentary rocks are ex-

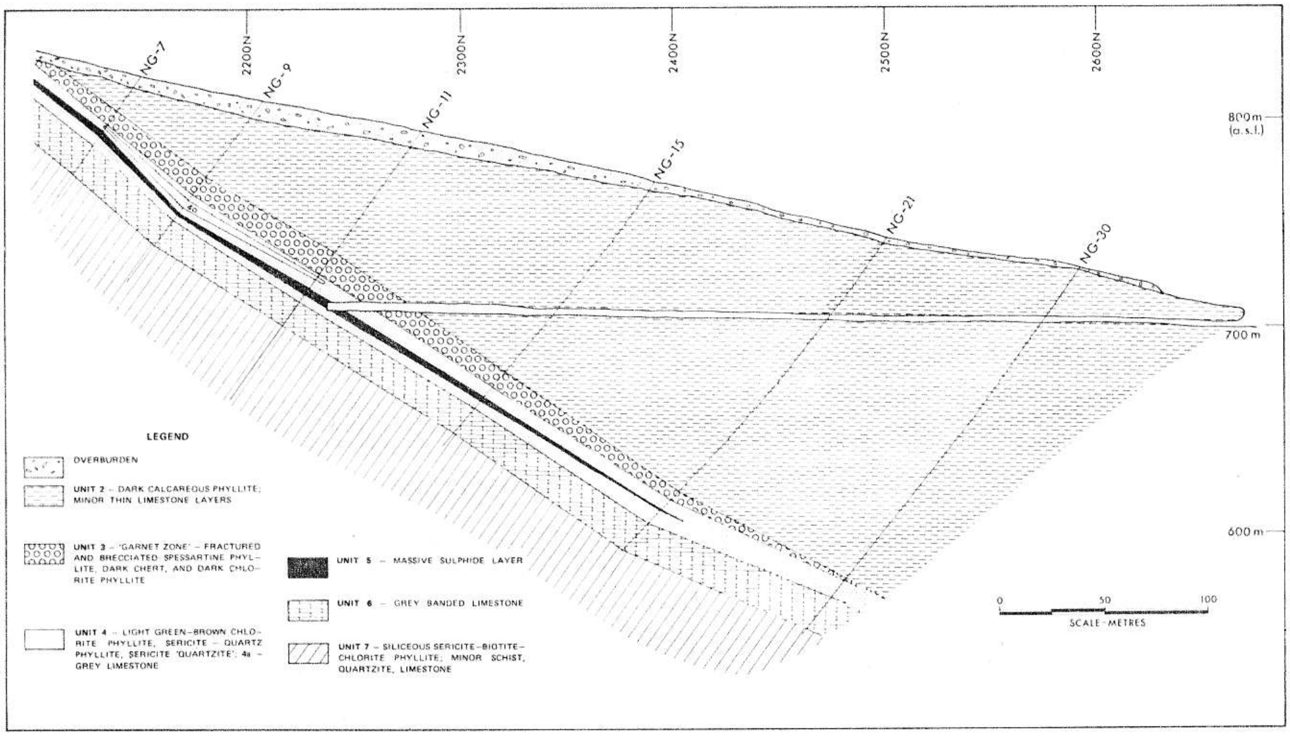

Figure 21. A north-south vertical section through the Goldstream deposit (from Höy, 1979).

posed in the cores of large antiformal structures in the Downie Peak area (just northwest of the Montgomery occurrence, Fig. 20), and in the Keystone and Standard areas. This suggest (Lane, 1977; Brown, Höy and Lane, 1977; Höy, 1979b) that these folds developed in an inverted stratigraphic panel, believed to be the underlimb of an earlier (Phase 1) recumbent nappe. Minor structures that can be related to Phase 1 deformation are not obvious; it was not possible to assign many rootless, isoclinal fold hinges to Phase 1 or Phase 2.

Small scale chevron folds and kink folds, crenulation cleavage, and small open folds in more competent units are superposed on earlier structures. They are common in the adit of the Goldstream deposit.

The Goldstream Deposit

The Goldstream deposit is a sheet of massive sulphides 2 to 4 metres thick, 500 metres wide and at least 1200 metres long in calcareous and chloritic phyllite. The only exposures are restricted to a number of weathered pits where the south end of the deposit subcrops. A pronounced mineral foliation, crenulation cleavage, small-scale late folds, and brecciation, quartz-carbonate veining, and fault gouge in the hanging wall phyllite, are the obvious structures observed underground.

Rock Units. The Goldstream deposit is within the 'metavolcanic-phyllite' division described above. Figure 21, a north-south vertical section through the deposit, illustrates the sequence of metasedimentary rocks above and below the deposit. Unit 1 (structurally above the section of Fig. 21) includes siliceous chlorite-biotite phyllite, phyllitic quartzite, calcareous and graphitic phyllite and a few impure limestone layers.

Unit 2 includes approximately 220 metres of dark carbonaceous phyllite interlayered with thin grey limestone layers. Calcite and biotite are common within the unit, and pyrrhotite is ubiquitous. The alignment of sericite, chlorite, and graphite (?) grains produce a well-defined foliation, and augens of quartz and carbonate and the abundant limy partings give this rock a distinctive layered appearance.

The 'garnet zone', unit 3, coincides with a pronounced fault zone. The rock is generally medium to dark green or grey in colour and contains abundant spessartine garnets. In part, it consists of dark, banded 'chert' layers, medium green chlorite-phyllite layers, and dark grey to black greasy lustered chlorite-graphite-calcite-quartz layers. Pyrrhotite may be very abundant, concentrated in layers or in discontinuous streaks, and grunerite occurs in some dark siliceous layers. Chemical analyses of unit 3 show an abnormally high manganese content and only trace copper and zinc (Appendix A in Höy, 1979b).

The garnet zone is sheared and broken, and cut by numerous quartz-carbonate layers. The garnets pre-date this deformation and probably an earlier deformation which produced the prominent mineral foliation in the metasedimentary rocks. This early foliation is bent and warped around the garnet porphyroblasts.

The garnet-rich layer is interpreted to be a metamorphosed manganiferous iron-rich chert horizon. It is areally restricted and terminates to the west away from the massive sulphide layer.

The massive sulphide layer is enclosed within light green to brown, very siliceous chlorite and sericite phyllite (unit 4) that grades to fine-grained sericite-chlorite quartzite. A grey limestone layer, 1 to 2 metres thick, occurs within unit 4 above the sulphide layer. Pyrrhotite, chalcopyrite and minor sphalerite, generally uncommon within the unit, increase substantially just above the sulphide layer. Here they occur as fine disseminations, discontinuous blebs, and as layer-parallel streaks. Sulphides are less common below the massive sulphide layer, occurring primarily as discontinuous layers in a dark, layered siliceous rocks.

A light grey banded limestone (unit 6), averaging 10 to 20 metres in thickness, occurs below the phyllites of unit 4. The limestone is underlain by siliceous sericite-biotite-chlorite phyllite, schist, minor quartzite, and limestone of unit 7.

Greenstone was encountered in three drill holes west of the deposit. Höy (1979b) believes that the greenstone lies structurally below the mineralized horizon as a grey limestone lying above it is tentatively correlated with the footwall limestone (unit 6). However, Noranda geologists correlate the grey limestone with unit 4 implying that greenstone is approximately stratigraphically equivalent to the massive sulphide layer (G. Gibson, personal

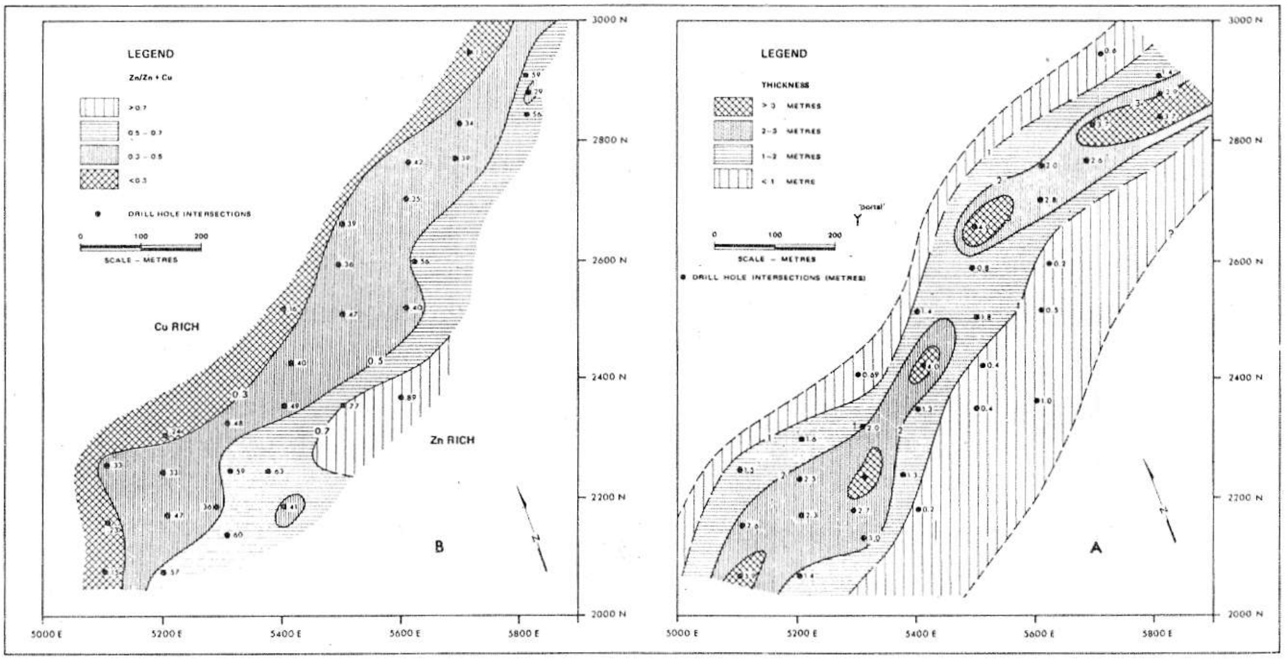

Figure 22a. Isopach map of massive sulphide (>40% sulphides) layer (from Höy, 1979).

Figure 22b. Weighted Zn/Zn + Cu ratios, massive sulphide layer, Goldstream deposit (from Höy, 1979).

communication, 1981). The greenstone varies from fine-grained, massive varieties to chloritic phyllite. In thin section, it is composed dominantly of actinolite, chlorite, epidote, and albite.

Mineralization. The massive sulphide-layer (unit 5) averages from 1 to 3 metres in thickness, has a strike length of at least 400 metres, and a plunge length of at least 1200 metres (Fig. 22a). Underground drilling by Noranda indicates that the massive sulphide layer splits into two layers in the western part of the deposit (Fig. 23). Only its western and truncated southern boundaries are defined. Its northern boundary is open, although at 3000N it is approximately 350 metres below Goldstream River. Its eastern boundary is restricted by a barren hole (at 25 + 62N, 59 + 00E) approximately 300 metres east of the last known sulphide mineralization.

The sulphide layer consists largely of pyrrhotite, chalcopyrite, sphalerite and trace galena. Pyrite is rare, occurring primarily in fractures or shears that cut the massive sulphides. The sulphides are generally medium to coarse grained and intimately mixed. However, finely recrystallized (<0.004 mm) admixtures of pyrrhotite and sphalerite are noted in thin section (Gibson, 1980). Sulphides generally have a granular texture, although streaking and shearing is fairly common, particularly toward the boundaries of the sulphide layer and there the finer grain size is more common. Layering, defined by alternation of the various sulphides, is not present (or at best, very rare).

Numerous rounded clear quartz fragments, darker 'chert' fragments, and dark siliceous chlorite-phyllite fragments are scattered through the massive sulphides. These may contain hairline fractures filled with chalcopyrite, inclusions of chalcopyrite, sphalerite, or pyrrhotite, or may be free of sulphides. They resemble the mineralized and non-mineralized siliceous metasediments in the country rock.

In general, the lower (footwall) contact of the massive sulphides is fairly sharp, whereas the upper contact may be more gradational with the mineralization in the hanging wall.

Sulphides in the country rock above the sulphide layer (the hanging wall) are in the form of fine disseminations, discontinuous blebs, and layer-parallel streaks in quartzites and siliceous phyllites. They also occur in bull quartz veins and in a complex network of thin, interconnected, generally layer-parallel frac-

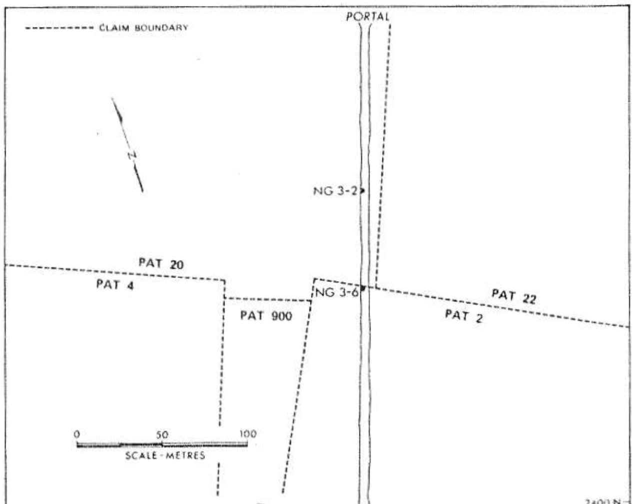

Figure 23. Plan view of underground workings, Goldstream deposit (from Noranda).

tures. Dark grey to black, 'greasy' lustered chlorite alteration may be associated with hanging wall sulphides.

Sulphides are less common and restricted to a considerably lesser stratigraphic interval below the massive sulphide layer. They occur primarily as discontinuous layers in a dark, layered siliceous rock.

Metal Zoning. Zn/Zn + Cu ratios in the massive sulphide layer generally increase toward the east (Fig. 22b). This tendency for increasing relative abundance of copper to the west is not apparent in either the hanging wall or footwall where higher ratios occur in central zones that parallel the northeast trend of the deposit. There is not a consistent vertical zonation in the deposit. The Zn/Zn + Cu ratios in the hanging wall and massive sulphides are very similar, and in the footwall, slightly less (Höy, 1979b).

Conclusions. Goldstream is interpretated to be in the north limb of a major Phase 2 antiform. Rock units in the limbs of Phase 2 structures are very attenuated as indicated by the pronounced Phase 2 foliation and boudinaging of more competent units. Hence, the form of Goldstream deposit must be substantially modified by this deformation. Its pronounced northeast trend is probably due to structural elongation in the direction of plunge of Phase 2 structures and its thin-layered form due to flattening in the plane of the foliation.

A number of features of the hanging wall may be contrasted with the footwall to suggest that the Goldstream deposit is inverted. The more gradational contact with the massive sulphides, greater abundance of sulphides, greater thickness of mineralization, and nature of mineralization and alteration in the structural hanging wall are generally more typical of stratigraphic footwall characteristics (the 'stringer ore') of Precambrian massive sulphide deposits (Sangster, 1972) than of stratigraphic hanging wall characteristics.

Consideration of regional structures in the Goldstream area also suggests that the sequence of rocks in the immediate vicinity of the deposit is inverted. At Downie Peak, 10 kilometres to the southeast, graded grit beds indicate that rocks young toward the core of the Downie antiform. The axial trace and younger core rocks of this antiform swing east/west just northwest of Downie Peak and are located south of the deposit in the Goldstream area.

An attempt to recognize an alteration 'pipe' may be futile due to the intense regional deformation in the area. An alteration pipe, if it existed in the deposit, may be so attenuated as to be no longer recognizable.

A number of features of the deposit indicates that it has been intensely deformed and metamorphosed along with the country rocks (Höy, 1979b; Gibson, 1980). These include the common gneissic textures of sulphides, locally contorted and brecciated sulphide wall-rock contacts, rounded gangue inclusions ("Durchbewequng" fabric described by Vokes, 1969) and overall geometry of the deposit. Regional and contact metamorphism has partially recrystallized some of the massive sulphides, resulting in medium-grained, granoblastic textures and secondary growth of euhedral pyrites in gneissic pyrrhotite. Remobilization of sulphides, particularly chalcopyrite, into pressure shadows and fractures in silicate inclusions is very common.

The Goldstream deposit is one of a number of stratabound massive copper-zinc deposits in the area. They are hosted by either basic volcanic rocks (Standard) or metasedimentary rocks spatially associated with basic metavolcanic rocks. They compare closely with the 'bedded cupriferous iron sulphide' or 'Besshi' type deposits in Japan (Kanehira, *et al.*, 1970). These are both bed-like or lenticular in form, are composed primarily of massive compact pyrite (pyrrhotite at Goldstream) chalcopyrite ore, and occur in geosynclinal crystalline schists associated with submarine basic volcanism. In contrast, some of the typical features of Kuroko deposits are absent: the association with acid volcanism, the obvious metal and ore-type zoning, the alteration pipe, and the association with sulphates (barite, gypsum, and anhydrite).

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