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**LEAD-ZINC DEPOSITS IN THE KOOTENAY ARC,
NORTHEASTERN WASHINGTON AND ADJACENT BRITISH COLUMBIA**

**SOCIETY OF ECONOMIC GEOLOGISTS
1970 NORTHWEST FIELD CONFERENCE GUIDEBOOK**

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I

INTRODUCTION^{①②}

By

A. E. WEISSENBORN, FRANK C. ARMSTRONG, and JAMES T. FYLES

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Mines and Petroleum Resources

INTRODUCTION

By A. E. WEISSENBORN,^① FRANK C. ARMSTRONG,^① and JAMES T. FYLES^②

Southeastern British Columbia and northeastern Washington, together with the panhandle of Idaho and a part of northwestern Montana, compose one of the great lead-zinc provinces of the world. Within this area are found deposits of lead and zinc of diverse types. The area includes the unique Sullivan ore body at Kimberley, British Columbia, and the deposits of the famed Coeur d'Alene district in Idaho, as well as the lesser known but nonetheless important deposits that occur in carbonate rocks in the Kootenay Arc.

The Society of Economic Geologists has selected the lead-zinc deposits in the carbonate rocks of the Kootenay Arc as the scene of its 1970 Field Conference. This seems a happy choice, both because of the economic importance of these deposits and the geologic relations that can be observed in them.

As far as we know, no overall appraisal of the potential productivity of the stratiform deposits in the carbonate rocks of the Arc has ever been made, but it obviously is large. Weissenborn (1966, p. 139) has summarized the potential of the Metaline district in the following terms:

Only that part of the Metaline district in the vicinity of the producing mines has been explored except by widely scattered drill holes. This amounts to only a small part of the area within the Metaline graben that is potentially favorable for ore. Data on which to base ore reserve estimates are lacking, but the potential of the district is obviously large. In testimony given in 1959 before the Federal Power Commission in a hearing on Boundary Dam, the writer made the following statement regarding potential ore reserves of the Metaline district: "It is likely that the Josephine horizon will produce 36 million tons of ore, and it is not impossible that the ultimate production will exceed 146 million tons. The grade will average between 3 and 4 percent combined lead and zinc, and the amount of zinc in the ore will be approximately twice the lead." Estimates of ore reserves given at the same hearing by other witnesses were greatly in excess of these tonnage figures. With adequate price levels for lead and zinc, it seems likely that the Metaline district will continue to be an important producer of these metals for many years.

Lead-zinc deposits in the Kootenay Arc in British Columbia contain known reserves of several million tons grading 3 to 6 percent combined lead and zinc. The potential productivity of these deposits and of those in Stevens County, Washington, is high, but known individual deposits in general appear to be smaller than deposits in the Metaline district and others away from the Arc in southeastern British Columbia.

Deposits of the Kootenay Arc may be classed with the Mississippi Valley type, but they differ from them in several respects. Most of the largest lead-zinc deposits in the Arc are stratiform, and although they have a num-

ber of similar characteristics, diverse types can be recognized. This is in contrast to the Mississippi Valley type deposits, which in most instances are remarkably similar over large areas (Ohle, this volume). Another contrast is that the deposits of the Kootenay Arc occur in several distinct stratigraphic units in Middle and Lower Cambrian formations, and not far from the Arc similar deposits occur in Precambrian formations. A third contrast is that the Kootenay Arc deposits are in a highly deformed terrain and none is more than a few miles from a granitic mass that intrudes the host rocks.

In deposits that have been termed the "Metaline type" (Fyles, this volume) the ore occurs as replacements in dolomitized limestone. The host rock has been only slightly metamorphosed and is characterized by broad, open folding. Many of the ore deposits are found on the crests and flanks of anticlinal structures, but this relation may be more apparent than real. The ores are commonly associated with breccias. Open space filling is common, and the deposits have many similarities to those of the Mississippi Valley type (Ohle, this volume).

Deposits that are termed the "Salmo type" by Fyles (this volume) or the "Remac type" by Sangster (1970) occur in dolomitized limestone that has been highly deformed by isoclinal folding. The deposits are closely associated with minor folds on the limbs of larger anticlinal structures. The ores are finer grained than those of the Metaline type and may be well banded. Pyrite is a common gangue mineral; pyrrhotite, at least in part, is a product of thermal metamorphism. Textures of the sulphides indicate that they have been deformed (Muraro, 1966; Sangster, 1970).

A third type of deposit has been termed by Fyles (this volume) the "Bluebell" type. This type consists of deposits of massive sulphides formed by replacement of certain limestones near fractures. Simple structural features control the formation of ore bodies.

Other deposits in Stevens County, Washington, differ in some respects from these types. Except for the Van Stone mine, none are in production now, and consequently they are not described in this Guidebook and they will not be visited during the Conference. They have been described in various publications of the Washington State Division of Mines and Geology.

Muraro (1966) states that the deposits of the Electric Point area in Stevens County are comparable to those of the Salmo type in gross structures, but that the internal deformation of the carbonate rocks is not normally as

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intense. He points out that where regional metamorphism is medium to intense, pyrrhotite is the dominant iron sulphide and a major constituent of the ore. He postulates that the least metamorphosed deposits of the Electric Point area may offer the nearest approach to the original nature of Salmo-type deposits.

Evidence concerning the origin of the deposits is conflicting. At first thought it would seem reasonable to suppose that all of the lead-zinc deposits in the carbonate rocks of the Kootenay Arc have had a common or similar history. This is not necessarily true. The geologic history of the deposits is complex, and the deposits may not even be of one age. Early workers, and some more recent ones (Park and Cannon, 1943; Dings and Whitebread, 1965) attributed the origin to hydrothermal solutions derived from the igneous intrusives. With respect to the Metaline deposits, McConnel and Anderson (1968) and Addie (this volume) present evidence that the deposits are closely associated with reeflike structures or occur at a facies change in the Metaline Limestone, and argue that at least in part the deposits may be of syngenetic origin. However, they recognize that there is also a structural control. In the Salmo area the lead-zinc deposits are confined to one stratigraphic unit, but they are also closely related to tight folds in the Reeves Limestone. Sangster (1970) believes that the ore sulphides were precipitated simultaneously with the host rock carbonate and were remobilized into economic deposits during the folding. Muraro (1966) also argues that the deposits were emplaced prior to the folding and the metamorphism. Fyles (this volume) believes that the deposits were emplaced after the folding but before the granitic rocks were intruded. Clearly, if the deposits were originally formed syngenetically, a considerable amount of remobilization has taken place since deposition.

On the basis of isotopic studies of leads from the Kootenay Arc, Sinclair (1966) concluded that the leads are of Precambrian ages and have had a complex multi-stage history. His conclusions are not in accord with a simple syngenetic origin.

These inconclusive and in part conflicting lines of evidence make the deposits of the Kootenay Arc of more than usual interest to geologists, particularly exploration geologists. We hope that discussions during this Conference will help reconcile some of the conflicting evidence and will give rise to suggestions for further work that

may help resolve the problems. Also, some additional light may be shed on the origin of Mississippi Valley type deposits in general.

On behalf of the Society of Economic Geologists we acknowledge with sincere thanks that the organization of this Field Conference has depended on the cooperation of many companies and individuals. Particular thanks are extended to the managements of Pend Oreille Mines & Metals Co., Reeves MacDonald Mines Limited, Canadian Exploration Limited, and the American Smelting and Refining Company for permission to visit their mines. The large amount of work by the authors of papers for this Guidebook and the help of geologists and other company personnel in acting as guides is very much appreciated. Thanks are extended to Mr. P. E. Olson, Inspector of Mines in Nelson, B. C., who arranged accommodations there. Finally, the job of printing the Guidebook for the Society was undertaken by the State of Washington Division of Mines and Geology through its Supervisor, Marshall T. Huntting, to whom we extend our thanks.

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II

MISSISSIPPI VALLEY TYPE ORE DEPOSITS—A GENERAL REVIEW

By

ERNEST L. OHLE

MISSISSIPPI VALLEY TYPE ORE DEPOSITS—A GENERAL REVIEW

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MISSISSIPPI VALLEY TYPE ORE DEPOSITS—A GENERAL REVIEW

By ERNEST L. OHLE^①

INTRODUCTION

For many decades, lead and zinc deposits of the Mississippi Valley (or Alpine) type have provided the largest share of these metals produced in the United States. Even on a worldwide basis, they have contributed annually at least 20 percent of the newly mined lead and zinc. In addition to lead and zinc, these distinctive deposits yield large quantities of fluor spar and barite and small amounts of copper, cobalt, nickel, and silver. The zinc concentrates are an important source of cadmium, gallium, indium, and germanium.

Extensive field studies conducted during and since World War II have provided much new knowledge about the geology of these deposits. Much effort has been expended by Government and academic personnel in many areas, and mining company geologists have made elaborate studies in several districts. Exploration programs have been well rewarded, with new discoveries of lead in southeast Missouri and zinc in central Tennessee being particularly noteworthy. Likewise, the first mining development of the large and rich deposits at Pine Point, N.W.T., has revealed details of this important occurrence.

In parallel with the above field studies, there have been advances in laboratory knowledge of the physico-chemical systems which must have been involved in the formation of these important mineral concentrations. Reports on carbonate rock replacement reactions, temperature data from fluid inclusion studies, and lead isotope determinations have been especially helpful. All of these efforts—both in the field and in the laboratory—have led to numerous scientific papers and to several symposia. The volume published after the SEG-UNESCO sponsored conference in New York in 1966 as Monograph 3, *Economic Geology*, contains a wealth of data. It includes bibliographies listing virtually all of the significant literature to that date.^② In 1969, the Society of Economic Geologists sponsored a field conference devoted to the East Tennessee zinc area.

Concentrated geologic studies by company exploration groups have contributed significant data of much benefit to scientific understanding of the regions where Mississippi Valley type deposits occur. Among these data might be mentioned: the minute details of stratigraphy of the Knox dolomite and the recognition of a great paleo-

aquifer system in east Tennessee; details of a fascinating sedimentary pattern in the Upper Cambrian of the Ozark Region; and three-dimensional knowledge unequalled elsewhere in the world of Paleozoic reefs in Missouri and at Pine Point. The present level of understanding of these features could never have been reached without the drilling of thousands of drill holes and without access to mine openings. The science of ore deposits owes a debt of gratitude to those mining companies which have generously shared their hard-earned information.

DISTRIBUTION

The distribution of known ore deposits which are classified by most geologists as belonging to the Mississippi Valley type is shown in Figure II-1. While it is possible that there are unreported deposits of this type in the great expanse of central Asia or under the polar ice caps, this map shows a marked clustering of major districts in the interior of North America and in western Europe-north Africa. Deposits of the type have not been reported in the southern hemisphere.

Although the mid-continent area of North America has been tectonically relatively quiescent since Precambrian times, a stable environment is not an essential feature of the habitat of this ore type. At least, the ore-bearing area does not need to have been passive throughout its history. The northeast Washington-British Columbia district, where this field conference is being held, is an example of a Mississippi Valley type district that has seen mountain building. Likewise, the East Tennessee district was involved in the great overthrusting of the Valley and Ridge province and the Algerian-Moroccan deposits have been cut by large block faults. Numerous deposits were involved in the Alpine orogeny. While it is not always possible to determine precisely where the mineralization fits into the tectonic history of all the areas, there are a few places, such as East Tennessee, where it is quite positive that deposition of the ore preceded tilting and thrusting. The ore deposits, thus, were simply caught up in the tectonic movement.

Nonetheless, there is a statistical basis for regarding interior areas that are relatively devoid of mountain building or major igneous intrusions and that contain a rather thin stratigraphic section of Paleozoic or Mesozoic carbonate rocks as the most likely hunting ground for more ore of the Mississippi Valley type.

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^②Rather than list all of the large number of papers which describe deposits of the Mississippi Valley type, an exercise which would unduly prolong the present summary paper, the author refers the interested reader to *Economic Geology* Monograph 3.

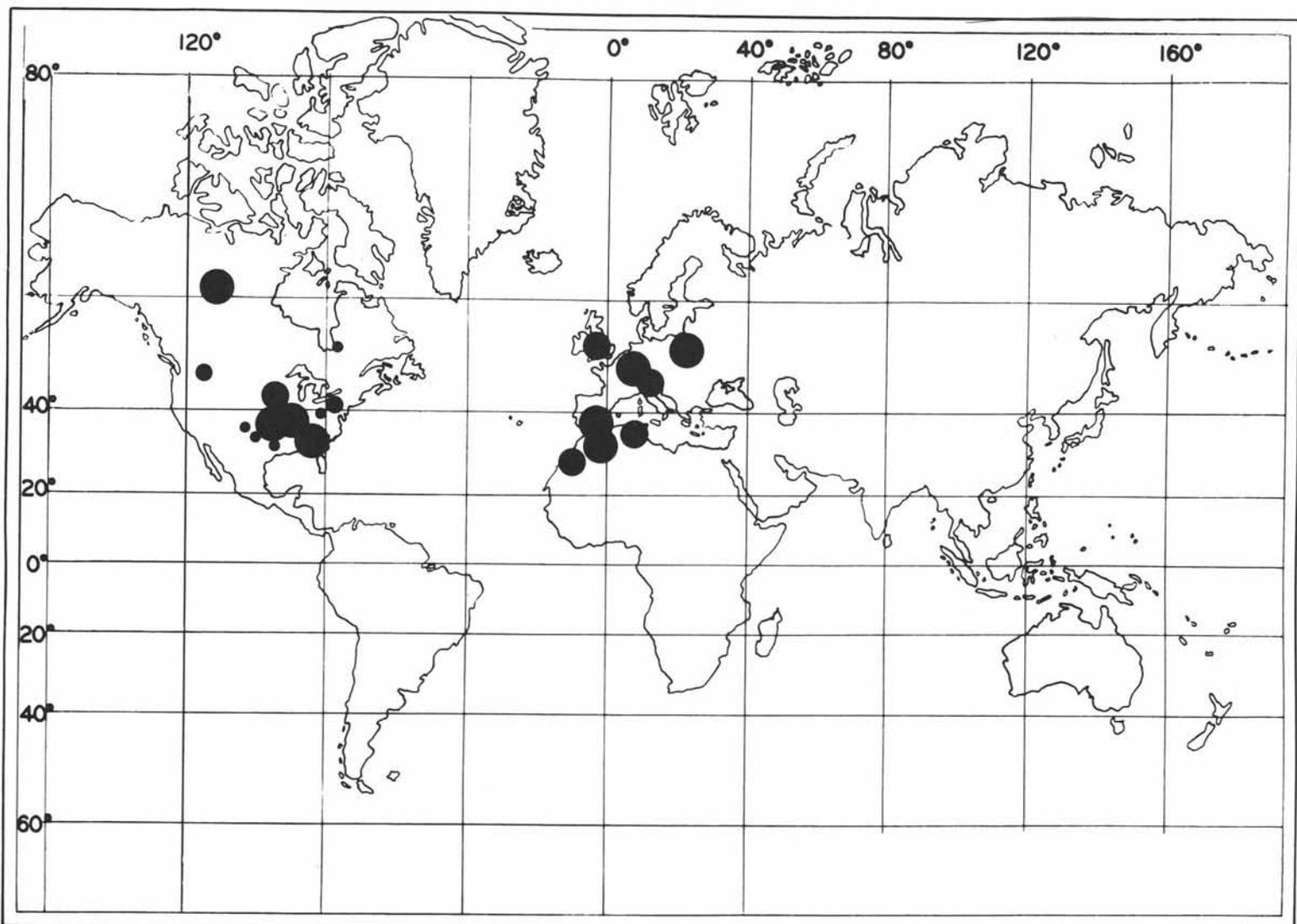


FIGURE II-1.—Worldwide distribution of ore deposits of the Mississippi Valley type. The size of the circles indicates the relative productivity of the different areas. (Courtesy of Economic Geology.)

GENERAL CHARACTERISTICS

SHAPE AND SIZE

The shape and size of ore bodies of the Mississippi Valley type vary widely, reflecting the dominant stratigraphic, lithologic, and structural control which influenced their localization. Most of them are tabular and most are in trends or "runs" that are from 25 feet to several hundred feet in width. The ore bodies may extend as much as a mile or more in length. Vertical dimensions range from a few inches to 200 feet or more.

Some "runs" are relatively straight, but more tend to be sinuous. In several districts "circle runs" which loop back on themselves result from a relationship to buried Precambrian knobs (southeast Missouri) or solution slumps (Joplin and Wisconsin-Illinois). Crosscutting ore bodies may be in simple veins such as those at Rosiclare, Illinois, and in central Kentucky; steplike veins called "pitches" in Wisconsin-Illinois; or near-vertical pipelike breccia masses such as are found in parts of the Joplin

and East Tennessee districts. At the other extreme are the "sheet ground" ore bodies near Webb City, Missouri, which replace a stratigraphic interval less than 10 feet thick and spread over many acres.

The tonnages in individual ore bodies range from a few thousand to several million tons, and a few of the major districts have produced over 300 million tons of crude ore. Ore grade has varied from approximately 2 percent upward. Most of the districts being mined today average between 3 and 10 percent of the valuable metals, but exceptionally rich parts of Pine Point have produced several hundred thousand tons exceeding 40 percent combined lead and zinc, and important parts of the bedded ore near Cave-in-Rock, Illinois, contained more than 80 percent fluorite, galena, and sphalerite.

The areal dimensions of the major districts are very impressive. Both the Tri-State and Southeast Missouri districts cover over 2,000 square miles, and in the Appalachians, very similar ore is found at numerous places between Sweetwater, Tennessee, and Friedensville, Pennsylvania, a distance of 600 miles.



FIGURE II-2.—An ore-bearing face of a stope in the reef zone of the Bonneterre Formation in southeast Missouri. The vertical, planar feature with the upward-arching parting planes is algal reef, and the bedded material on either side is reef detritus and calcilutite. The favored place for galena to be found is in disseminations and tiny veinlets along the vertical contacts on either margin of the algal mass. (Courtesy of the Geological Society of America Bulletin.)

STRATIGRAPHIC RELATIONS

These ore deposits tend to occur in restricted stratigraphic intervals. This is true even in districts where much of the ore is veinlike or in breccia rather than bedded so that the restriction is not just a matter of selective replacement. Certain carbonate intervals preferentially develop permeable repositories for ore, often through solution and collapse as in the Joplin, Southern Illinois, Wisconsin-Illinois, and East Tennessee districts.

Although lateral variations in lithology do not correlate with ore bodies in most of these areas, Southeast Missouri presents many striking examples of a close relation between ore and facies changes. In this district, much ore is found where contrasting rock types are in juxtaposition as on the margins of reefs or where coarse dolomite grades over short distances into fine-grained dolomite or limestone (Fig. II-2).

It is perplexing that despite much effort to isolate the factors which make certain rock units favorable for ore there still is no clear-cut explanation of this widely observed phenomenon.

STRUCTURAL RELATIONS

Mississippi Valley type ore deposits are associated with structures of many kinds. On a regional scale, several important districts are peripheral to the Ozark and Wisconsin Domes. On a smaller scale, the tendency for individual ore bodies to flank Precambrian highs in Southeast Missouri has already been mentioned. Other very important ore "runs" in this district are related to anti-form sedimentary depositional features (Ohle and Brown, 1954). In Joplin and East Tennessee, solution and cave collapse resulted in formation of favorable breccia bodies. In Wisconsin-Illinois, the "pitch" ore bodies and the "circle runs" likewise are due to fracturing which resulted from solution and subsidence. Whereas faulting is of no major consequence in controlling the main ore distribution in Southeast Missouri, the Miami Graben definitely guided some ore in the Tri-State (Fowler and Lyden, 1934). Also, at Cave-in-Rock, the parallelism of the Peters Creek fault and the bedded fluor spar replacements in Mississippian limestone seems too close to be coincidental (Grogan and Bradbury, 1967). Many of the Illinois-Ken-

tucky veins are in minor faults, as are some of the slightly productive veins in Missouri such as those at Palmer and Valles mines.

From the above it is apparent that no one particular kind of structure is found in every district but that each district does exhibit individual structural controls which are important to ore localization.

CARBONATE HOST ROCKS PREFERRED

Most of the Mississippi Valley type ores occur in carbonate rocks and, in most cases, the rock is dolomite rather than limestone. However, significant ore bodies have been mined in sandstone in Southeast Missouri, at Maubacher-Bleiberg in West Germany, and in the Caledonides of Scandinavia (Grip, 1967). Likewise, an unusual body of granite boulders was host for the Hayden Creek ore body in Missouri (Ohle, 1952). Thus a carbonate host rock is not essential.

HOST ROCK ALTERATION

Dolomitization, chertification, and carbonate recrystallization constitute the main changes that occur in the rocks surrounding the ore deposits. In some districts there is apparently so little alteration that the introduction of the ore minerals themselves seems to constitute almost the complete effect of the ore-forming activity. In other areas the alteration effects are widespread and difficult to distinguish from regional sedimentary variations.

SIMPLE MINERALOGY

Although the total list of minerals reported in Mississippi Valley type districts is substantial, each individual district contains only a few, and, most interestingly, the ore in each district is so distinctive in appearance that there rarely is a problem in identifying the source of a given specimen. Thus, in the Ozark Region, lead-zinc mineralization occurs in no less than seven stratigraphic positions in a total rock column of only a few thousand feet. In each stratigraphic position, the ore is distinguishable by a distinctive mineralogy, texture, and alteration, as well as a special structural environment. This constitutes one of the more puzzling features noted in Mississippi Valley ores.

A low precious metal content compared to most other types of lead-zinc deposits also is characteristic. Some of the zinc-rich sections in southeast Missouri may contain 2 to 3 ounces of silver per crude ton, but galenas everywhere contain little silver, and no gold has been reported. No silver mineral has been reported, and, since the silver invariably shows up in the zinc concentrate, it is concluded that that element is caught within the sphalerite lattice.

Galena from several districts is notable for its unusually large content of radiogenic lead, and in southeast Missouri the spread of isotopic ratios is sufficiently large to have promoted efforts to show a distribution pattern of the heavier isotopes (Brown, 1967). These efforts have indicated that generally galenas in the lower part of the Bonnetterre Formation are more radiogenic than those

found at higher levels. The patterns are not as clear cut as would be desired to show flow channelways, but might be sharpened if the number of samples were increased severalfold.

Isotopically anomalous lead in the Mississippi Valley is termed J-lead because it was first noted in Joplin material. In all cases the excess of radioactive isotopes renders it impossible to make age determinations by the standard techniques.

REPLACEMENT VERSUS OPEN-SPACE FILLING

In Southeast Missouri open-space filling is rare. In contrast, Joplin and Southern Illinois contain many vugs and caves in which large and beautiful museum specimens of galena, sphalerite, fluorite, dolomite, and calcite are found. Other districts fall between these extremes. The East Tennessee district is especially interesting in that there the great solution breccias (Fig. II-3), which at one

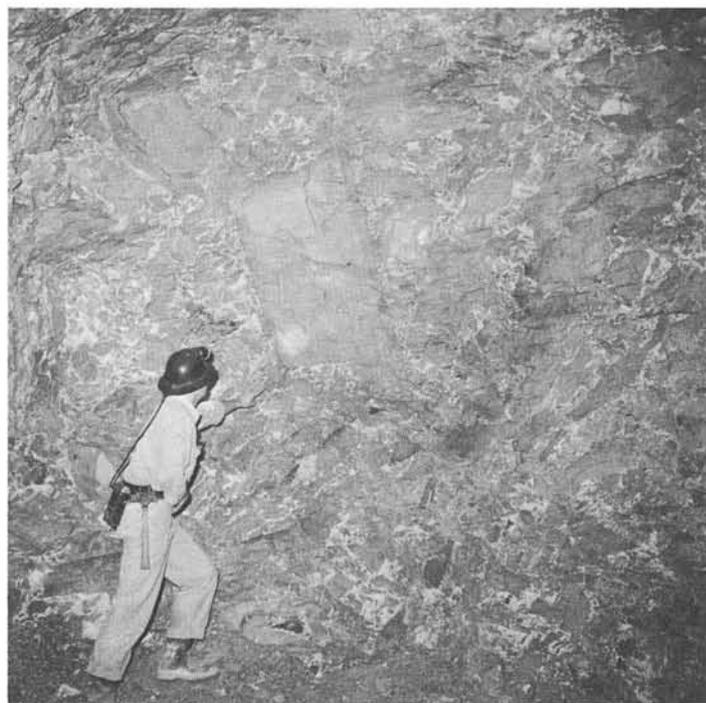


FIGURE II-3.—A spectacular collapse breccia in the post-R bed horizon at the Jefferson City mine of the New Jersey Zinc Co., Jefferson City, Tennessee. (Photo by D. L. Kendall.)

time contained large amounts of open space, today contain very little, as the filling by ore and gangue minerals is virtually 100 percent complete.

LACK OF GENETICALLY RELATED IGNEOUS ROCKS

Intrusive rocks to which the ores might be genetically related are lacking in these districts. Most areas contain a few igneous dikes and, in Missouri, there are diatremes of mid-Devonian or later age not far removed from lead deposits in Cambrian rocks. However, in no Mississippi Valley type area are there deposits with a zonal pattern around an igneous mass suggesting it as a source for the

ore solutions. In recent years, as more has been learned about the wider and wider occurrence of ore in various districts such as East Tennessee (now spreading into Central Tennessee) and Southeast Missouri, it has become apparent that a simple hydrothermal plumbing system with a localized source simply is not applicable here.

EVIDENCE OF SOLUTION ACTIVITY

A number of the Mississippi Valley deposits show evidence of carbonate solution preceding deposition of the ore minerals. In some instances, solution, generally accompanied by collapse, seems to have occurred simultaneously with ore fluid circulation and must have been accomplished by it. Thus the ore fluid not only brought in the metallic elements but actually contributed to the ground preparation. Southern Illinois, East Tennessee, Joplin, and Wisconsin-Illinois districts are especially notable in this regard.

Much has been learned about the plumbing system associated with the breccia and ore in East Tennessee in recent years. In 1969, the Society of Economic Geologists sponsored a field conference at Knoxville entitled "A Paleoquifer and its Relation to Economic Mineral Deposits." Agreement is now quite widespread that the large solution cavities in the Knox dolomite, now filled with breccia, are related to the unconformity at the top of the Lower Ordovician as first advocated by Laurence and his coworkers in the U. S. Geological Survey, and more recently illustrated by Callahan (1964) and Hoagland, Hill, and Fulweiler (1965). Sphalerite-bearing detritus in which fine laminations are parallel to the host rock bedding occurs in some of these former open caves. As the beds were tilted by the Appalachian orogeny and now dip steeply, these observers have concluded that the ore is pre-Appalachian Revolution and probably is contemporaneous with the unconformity. Contemporaneity with the unconformity is not imperative, however, and is not agreed to by all workers in the district.

SHALLOW DEPTH OF FORMATION

For many years it was postulated that ore of the Mississippi Valley type would be found only near the surface because of limitations in the modes of origin which were favored at that time. It is true that virtually all of the production to date has come from depths of less than 1,500 feet. This, however, could quite possibly be due to the fact that ore bodies near the surface are easiest to find. Furthermore, because of the gentle dips of the host rocks in most areas, the mine workings do not extend to depth very rapidly.

New discoveries in the past 20 years have challenged the old axiom. Ore is now known to be present below 1,500 feet in Missouri, and some of the new finds in Tennessee are over 2,000 feet below the surface. If the zinc noted in oil wells cutting the Arbuckle Formation in central Kansas represents ore of the Mississippi Valley type, as is not unlikely, then the depth range exceeds 5,000 feet (Evans, 1948). It seems probable that the depth at which ore exists today is governed principally by the depth at which the favorable formation exists. If, as in

the case of East Tennessee, the ore was formed early in the sedimentary history of the region, the depth at which the ore is now found bears little relation to its location at the time it was deposited. However, in localities where the total stratigraphic column is only a few thousand feet thick, it is obvious that the ore bodies were formed close to the surface.

HYPOTHESES OF ORIGIN

Lack of conclusive evidence supporting any of several possible modes of origin has made the ore deposits of the Mississippi Valley type the subject of a prolonged debate. Opinions have vacillated from one extreme to the other, partly reflecting the amount of evidence available at a given time but mostly keeping in step with the changing ideas among geologists as to the origin of ore deposits in general. For the most part the ideas presented in the past have been based on evidence gathered in individual districts. This was a shortcoming because there is insufficient evidence in any one district to guide a tightly drawn hypothesis. In recent years there has been more general acceptance of the probability that all of the deposits had a similar origin so that guiding and limiting evidence could be gathered from all deposits to formulate a hypothesis. This has helped, but no general agreement has yet resulted.

If it is admitted that Mississippi Valley type deposits have so many features in common that a common mode of origin is likely, it is obvious that this mode of origin must be a flexible one, capable of fitting all the observed patterns and of explaining all the variations from district to district. It must give a logical explanation for the presence of large lead ore bodies on one side of the Ozarks and equally large zinc ore bodies on the other side; it must account for the veins and the bedded ores, for replacement and open-space filling, and for extensions to several thousand feet below the present surface. And further, it must account for the great lateral extent of the Mississippi Valley type deposits.

There are six modes of origin for which substantial arguments have been advanced in one or more districts:

1. Original syngenetic deposition.
2. Original dispersed and low-grade syngenetic deposition with later concentration by regional metamorphism.
3. Original dispersed syngenetic deposition with later concentration by ground water moving upward in artesian flow.
4. Original dispersed syngenetic deposition with later concentration by downward-moving ground water.
5. Deposition from fluids of igneous derivation with hydrothermal or gaseous transport either with volatile aid or simply as metallic vapor.
6. Deposition from connate basinal water which is caused to move structurally up dip by compaction or other loading.

These modes of origin will be discussed in the order listed.

1. Although the concept of syngenetic sulphide deposition as the result of an extraordinary sedimentary process is becoming more and more widely accepted for certain copper and zinc deposits in shales and siltstones, such as those at White Pine, Michigan; in Zambia and the Congo Republic; and at McArthur River, Northern Territory, Australia, the process simply is not generally applicable as an explanation for many Mississippi Valley type districts. The unusual richness of certain areas such as Pine Point which would require deposition of a sediment exceeding 75 percent in sulphide content, makes it unlikely that simple sedimentary processes are the mechanisms which formed the ore deposits. Also, evidence for replacement is abundant; and the widespread occurrence of ore in breccia and crosscutting veins indicates that the rocks were well lithified prior to metal introduction. On this evidence, syngeneses can be dismissed from serious consideration as the mode of origin for Mississippi Valley type ore deposits.

2. The second idea, that of original deposition modified by later metamorphism, is the one advanced by Schneiderhoehn (1952). As he visualizes the mechanism, the distribution of atoms in nature is largely determined by environmental stability, and a sufficiently intense change in the environment of a mineralized area may set the metal atoms in motion toward a new location where they will again be stable. The process might be repeated several times during successive orogenies. The idea was first proposed for certain Alpine deposits and later extended to explain the Tunisian-Algerian-Moroccan lead-zinc field. For the African deposits, it is postulated that lead-zinc concentrations in the Paleozoic basement rocks, very possibly of subcommercial grade, were set in motion by the Alpine orogeny and came to rest as ore grade replacements of the Lias dolomite.

As John S. Brown states in his introduction to his privately circulated translation of Schneiderhoehn's paper (Schneiderhoehn, 1952), the hypothesis may be supported by the geological situation in the Alps or North Africa where there are great thicknesses of sediments underlying the ore bodies and where tectonic activity has been intense, but in the Mississippi Valley area much of the ore is low in the Paleozoic section and there has not been intense metamorphism in post-Precambrian time. Also, there are no known concentrations of lead or zinc in the many square miles of exposed Precambrian rock in the area which might have served as the "source bed," even if there had been sufficient environmental change to regenerate these concentrations. Thus the hypothesis breaks down if an attempt is made to apply it to Mississippi Valley type deposits generally.

3-4. The third and fourth genetic alternatives involve ground water, with variants that invoke artesian circulation on the one hand and downward percolating and laterally moving water on the other. As proposed in this hypothesis, the original deposition of the ore elements was during more or less normal sedimentation with the lead, zinc, and other constituents being highly dispersed. Later the widely scattered metal atoms were leached from the country rock and concentrated by redeposition in favored sites.

These hypotheses are not discussed much today, but they were widely held for several decades at the start of the present century. The reason for listing them separately is that each idea had very outspoken advocates and much was written on behalf of both of them. Much was made of the presence or absence of impervious shales above or below the ore zones, and exploration programs were guided into certain areas or avoided certain areas because of the distribution of the presumed essential circulation systems.

The artesian hypothesis was proposed especially for the Tri-State and Wisconsin-Illinois districts (Siebenthal, 1915). In both places the deposits are on the flanks of major domal structures and artesian circulation exists there today. Other Mississippi Valley type areas also would fit the idea. As mentioned above, the East Tennessee district is associated with a large paleo-aquifer system. While no artesian flow exists today in the "Old Lead Belt" of Southeast Missouri, because major faults have reversed the original dip away from the Ozarks, an artesian system probably was present in the past. However, for Pine Point, Southern Illinois, and certain others of the Mississippi Valley type districts, structural conditions and the known geological history suggest that artesian circulation probably never has been present.

Downward-percolating ground water, of course, is present everywhere, and, indeed, certain of the districts such as those in the Appalachian Valley have experienced enormous flows underground. Thus there is no shortage of the potential "ore fluid" if it can be shown to be capable of doing the job of ore transportation.

The objections to both the artesian and downward-moving ground water hypotheses are based mainly on 1) the questionable efficacy of ground water to leach metals out of a block of rock and to get them into solution, and 2) the difficulty in explaining why ground water moving through very similar rock sections should become charged mainly with lead in Southeast Missouri, mainly with zinc in Southwest Missouri, and mainly with barium or fluorine in other areas.

In the mines, leaching has penetrated for a short distance into the walls of solution channels, as indicated by a pronounced color change, but much rock shows no evidence of leaching. If large bodies of rock cannot be penetrated sufficiently by solutions to remove the metal, is not the supply of available metal too limited to give rise to an ore body, even if one grants complete efficiency in leaching and in later deposition?

5. The fifth of the major hypotheses that have been advanced is the "hypogene hydrothermal theory," advocating deposition from primary solutions emanating from igneous magma. This hypothesis has an advantage in its ready flexibility, so that there is no problem in explaining ores of different metals in close proximity to each other, or the occurrence of ore minerals in "tight" rocks. However, in the Mississippi Valley areas the hypothesis does have the disquieting problem that no adequate igneous sources are exposed. Likewise, advocacy of a simple zonal association around some unexposed intrusive body is not satisfying because the districts are so large. Either the

igneous source would have to be very large, or circulation over distances of tens, even hundreds, of miles would be required. Such igneous rocks as are known in the regions are almost all basic without acid counterparts and there is no evidence that magmatic differentiation has occurred. Is it possible that the ore fluids stem from a different kind of igneous source, more widespread in extent, which occurs at great depth and has no tendency to move upward to form major intrusions? The Joplin and Southeast Missouri districts cover at least 2,000 square miles in which, given the favorable formations, certain structures, and access to the mineralizing solutions, ore of a specific type will be found—not just similar ore but ore identical in mineralogy, texture, and general appearance.

It is this departure from the normal hydrothermal picture, as advocated for many deposits around the world, that raises a question as to whether the Mississippi Valley type ores, if they are hydrothermal, have not resulted from some unique aspects of the process of ore formation. The low silver content and the anomalous lead isotope ratios in the galenas also may be evidence that these deposits are products of some variant from the simple hydrothermal family tree.

6. In recent years, as more has been learned about the composition of deep basin brines and of fluid inclusions within ore minerals, the idea has gained some consideration that these brines are the ore solutions that give rise to the ore deposits. Helgeson (1967) has been a prominent proponent. The concept is based on the fact that interstitial connate water in buried arkosic sediments may contain as much as one part per million of lead when it reaches equilibrium. The amount of pore solution, source rock, and time required to derive and transport sufficient lead to form a major ore deposit are described as "geologically realistic." Movement of the solution from the basins to points of deposition is promoted by compaction.

The fact that basin sediments contained much water which moved up dip due to burial as sedimentation proceeded seems most likely. Also the occurrence of important districts like Southeast Missouri, Tri-State, and Wisconsin-Illinois on the flanks of domes adjacent to deep basins makes it quite likely that some of the basin water moved through the ore-bearing areas. However, there are still many questions to be answered, such as: why only a certain small percentage of the basin margin has ore when the solutions presumably moved in all directions; whether there are indeed arkosic source beds containing lead in these basins; and how, if you admit the lead could have come from such a source, do you account for sphalerite, barite, fluorite, and the other constituents in the ore. Much work remains to be done to develop this idea further and to test in the field whether the necessary source rocks are present. Many oil wells have been drilled in the Illinois and Forest City basins flanking the Ozark and Wisconsin domes and it should be relatively easy to determine whether the field facts support the physico-chemical possibility. If there is confirmation, the hypothesis is strengthened; without confirmation, the idea cannot be accepted.

SUMMARY

Ore deposits of the Mississippi Valley type are important sources of lead, zinc, barite, and fluor spar, as well as significant sources of several other metals. Each area has individual characteristics in mineralogy and structural environment, but all of them have certain similarities which suggest that they have a common origin. Each of the various hypotheses of origin, however, has certain objections, and it may be some time before diagnostic evidence is found which supports a hypothesis that gains universal acceptance. Nonetheless, much has been learned both in the field and in the laboratory over the past 20 years and there is hope. Meanwhile, exploration continues to find important new ore areas and it is evident that these deposits will continue to be valuable sources of lead, zinc, barite, fluor spar, and the other commodities for many years to come.

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III

**GEOLOGIC BACKGROUND OF THE METALINE AND NORTHPORT
MINING DISTRICTS, WASHINGTON^①**

By

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^①Publication authorized by the Director, U.S. Geological Survey

GEOLOGIC BACKGROUND OF THE METALINE AND NORTHPORT
MINING DISTRICTS, WASHINGTON

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By ROBERT G. YATES

INTRODUCTION

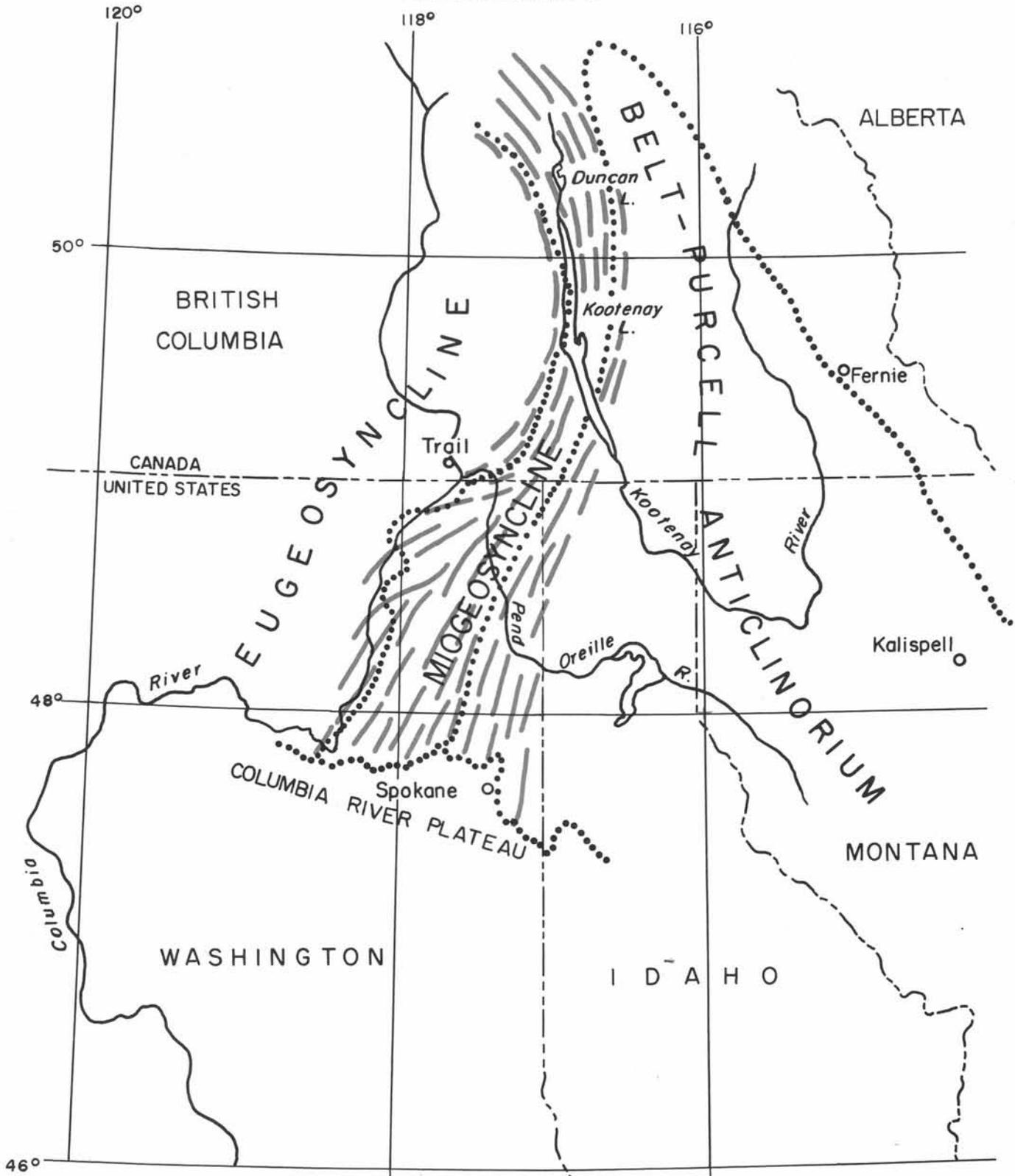


FIGURE III-1.—Location and trend of Kootenay Arc. Structural trends shown in red.

The lead-zinc mines of northeastern Washington fall within the domain of the Kootenay Arc, as do those farther north in the Duncan Lake area of British Columbia. The Kootenay Arc (Hedley, 1955) is a narrow arcuate belt of folded and faulted rocks of anomalous northeasterly trend that dominates the regional structure of Stevens and Pend Oreille Counties, Washington, and adjacent British Columbia (Fig. III-1). Beyond Kootenay Lake, 100 miles north of the International Boundary, the arc begins with the regional northwesterly trend of the Canadian Cordillera; 60 miles north of the Boundary, the trend is north-south; and from there southward to and 100 miles beyond the Boundary; the trend is northeasterly. The arc disappears under the late Tertiary basalt flows of the Columbia Plateau. Broadly, the arc separates the late Paleozoic and Mesozoic eugeosyncline of British Columbia and Washington from the Precambrian rocks of the Belt-Purcell anticlinorium, which extends across southeastern British Columbia, northwestern Montana, and the panhandle of Idaho. Although the Kootenay Arc is primarily a tectonic feature of highly folded and faulted rocks, it has a special regional stratigraphic significance, because it preserves and exposes the westernmost miogeosynclinal lower Paleozoic sedimentary rocks in north-

western United States. Among the miogeosynclinal rocks are limestones and dolomites of Cambrian age, the hosts for almost all of the lead-zinc deposits in northeastern Washington.

Accompanying the Cambrian carbonate rocks and falling within the tectonic limits of the arc are clastic rocks of Belt age, Windermere (late Proterozoic) conglomerates and tholeiitic basalts, Cambrian quartzites and phyllites, Ordovician, Silurian, and Devonian black slates, upper Paleozoic argillites, greenstones, clastics, and limestone, Mesozoic marine volcanic rocks, and Cretaceous and Eocene continental sedimentary and volcanic rocks as well as plutonic and hypabyssal intrusive igneous rocks.

To assign the process of lead-zinc mineralization to its proper place in the complex of rock forming and deforming processes that produced the above conglomeration, it is necessary to dissect the Kootenay Arc; this means that one must explore that part of its history that extends back into the Precambrian, restore the facies distribution of Paleozoic and Mesozoic layered rocks, decipher the tectonic evolution that produced the structural pattern, and identify the thermal events that produced the batholiths and plutons and metamorphosed the intruded rocks. It is obvious that all this cannot be done in a definitive

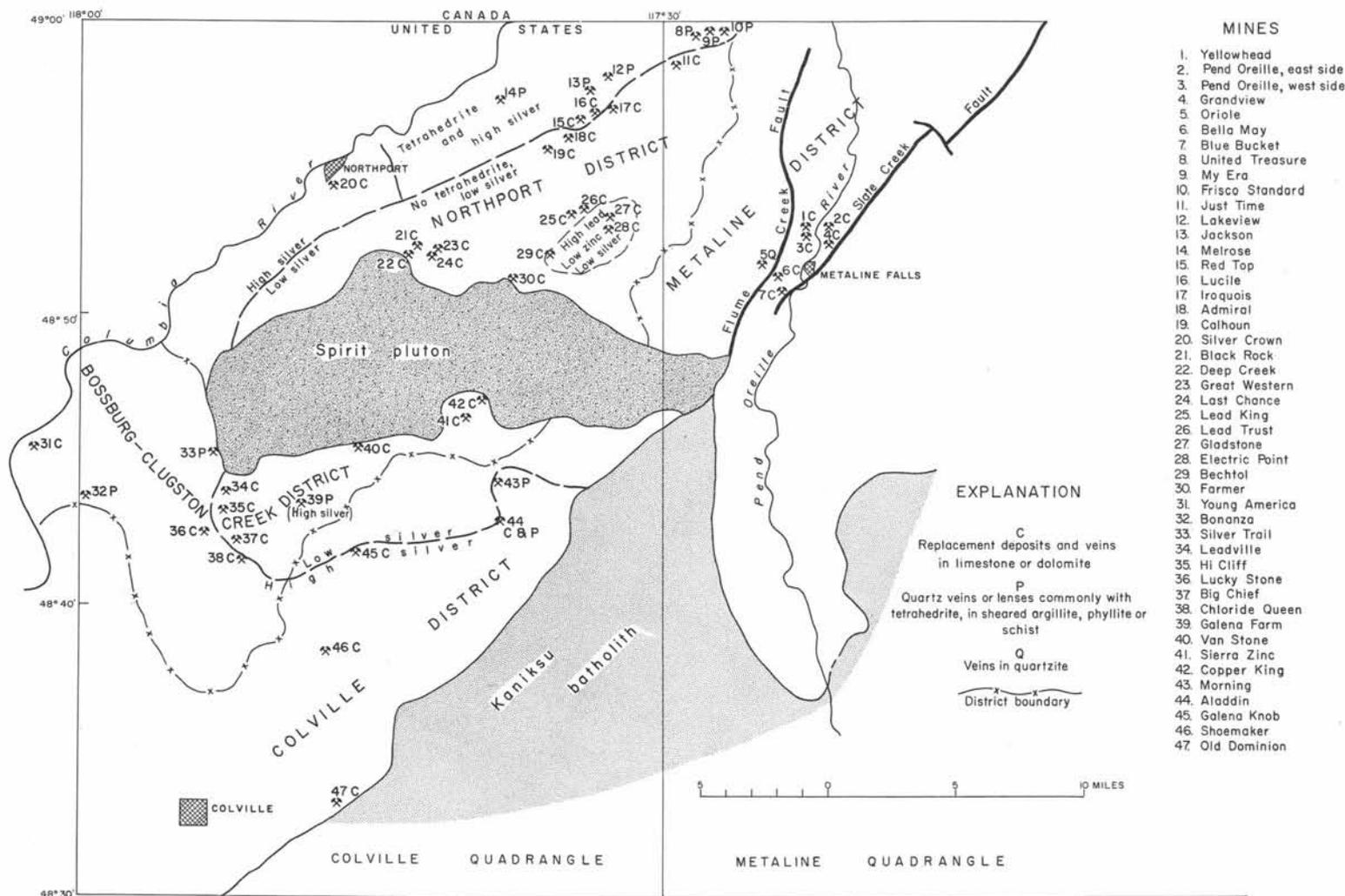


FIGURE III-2.—Index map showing location of mines and mining districts.

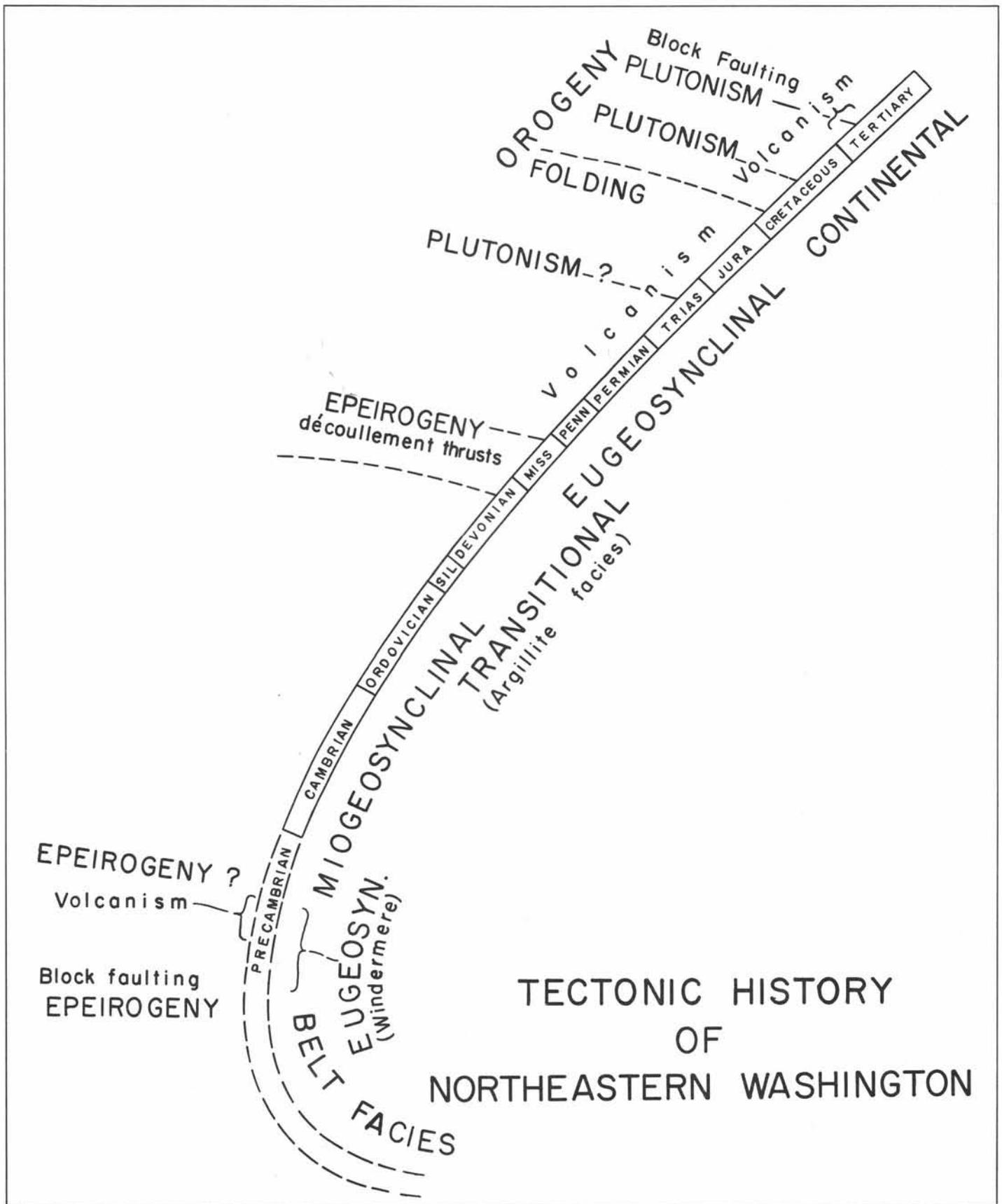


FIGURE III-3.—Diagram showing tectonic history of northeastern Washington.

manner in a paper as brief as this must be, even if all the facts were at hand; only salient features can be described and these but briefly. Only that part of the Kootenay Arc will be described that includes the Metaline and Northport districts (Fig. III-2), which fall within the 30-minute Metaline and Colville quadrangles. The area described extends southward from the International Boundary for 35 miles and westward from the Idaho boundary to the Columbia River.

DEPOSITIONAL HISTORY

Almost all lead-zinc deposits in northeastern Washington are replacements in carbonate rocks deposited during the Cambrian; therefore, the history of mineralization began no earlier than that time. However, to establish the character of the crustal environment that the Cambrian rocks inherited, it is necessary to touch on certain tectonic events that occurred during late Precambrian, Windermere, time (Fig. III-3). It is also desirable to review, if only briefly, the pre-Windermere time occupied in depositing the Belt sediments.

This review can be made by referring to Campbell's description (Yates and others, 1966, p. 47-48) of the Belt terrane that occupies much of northern Idaho and northwestern Montana:

Perhaps the most impressive aspect of Belt Series rocks is the monotony of gross lithology through about 50,000 feet of stratigraphic section and over thousands of square miles. Their characteristics include fineness of grain, drab colors, and uniform mineralogy. Such gross similarities have resulted in significant problems in long-range correlation. The monotonous character of the Belt Series is interpreted as reflecting a long, uniform depositional history. Much is still unknown concerning erosion, transport, and sedimentation in the Precambrian environment, and different workers have greatly differing views on sources of sediments and environment of deposition. Characteristics of the Belt rocks in the Coeur d'Alene subprovince suggest remarkable stability of both source area and depositional basin. These sediments were transported in streams of low gradient and were deposited in a marine or brackish environment that for long periods of time consisted of seas so shallow that extensive mudflats were common. Changes in lithofacies are subtle over long distances.

The rocks Campbell is speaking of are a trilogy of fine-grained quartzite, still finer grained siltites, and argillites. Also present, but in lesser quantity, is limestone.

In the Metaline mining district, the Belt sedimentary rocks are believed to be proxied by an assemblage of rocks called the "Priest River Group" by Park and Cannon (1943, p. 6). The Priest River Group is predominantly phyllite but includes substantial amounts of limestone, dolomite, and quartzite. Although "volcanics" are mentioned as present, and units of greenstone were separated in the mapping, Park and Cannon specifically mentioned (p. 9) that no volcanic detritus was seen in the conglomerate that unconformably overlies the Priest River Group. Inasmuch as this conglomerate is composed of clasts identified with the Priest River rocks, it seems probable that the "greenstones" are intrusive rocks contemporaneous with the volcanic rocks that overlie the conglomerate.

Since the reconnaissance mapping of Park and Cannon (1943), no stratigraphic study of rocks of the Priest River Group has been made; consequently, one can only speculate on whether they correlate more closely with the Belt Supergroup or rocks of the Deer Trail Group that occur in the Chewelah quadrangle 30 miles to the southwest. In the eastern part of the Chewelah quadrangle are rocks that can be identified as positively belonging to the Belt Supergroup, as well as similar rocks that can be identified with the Deer Trail Group. According to Clark and Miller (1968), all but the uppermost units of the Belt can be recognized. The highest identified unit, the Wallace Formation, is overlain conformably and sequentially by three rock units listed as "Precambrian rocks of questionable stratigraphic position," which are a carbonate unit, a siltite-argillite unit, and a "maroon argillite, siltstone, and quartzite" unit. According to Miller (oral communication, 1970) the carbonate unit may be equivalent to the Edna Dolomite and the siltite-argillite unit to the McHale Slate, both of the Deer Trail Group, which are found immediately to the west across the Jumpoff Joe fault. This makes the Wallace Formation equivalent to the Togo Formation (the lowermost formation in the Deer Trail Group) and leaves the Deer Trail's two upper units, the Stensgar Dolomite and Buffalo Hump Formation, without demonstrable Belt counterparts.

The fact that Belt stratigraphy is decipherable without too great difficulty over the thousands of square miles of the anticlinorium, but is exceedingly difficult to integrate with obviously related rocks along the western margin of the anticlinorium (the eastern margin of the Kootenay Arc), indicates that the part of the Belt depositional basin along the eastern margin of the Kootenay Arc was much less stable than that which produced the rocks of the anticlinorium. This contrast between the interior of the Belt terrane and its western margin did not end when the last of the Belt sediments were deposited. The marginal parts of the anticlinorium are strongly deformed, but its interior is characterized by broad open folds and normal and strike-slip faults, and it is strikingly free of any well-defined foldbelt. Although some folds and some faults were formed, or at least originated, in the Precambrian, most are believed—albeit without compelling evidence—to be Laramide, or Late Cretaceous in age. For long periods of time after the deposition of the Belt Supergroup and before the Laramide orogeny, the anticlinorium may have undergone several epeirogenies, but certainly it was never deformed into foldbelt mountains. For practical purposes, the Belt terrane was from Cambrian time an integral part of the craton.

The area that lies west of the anticlinorium in the Kootenay Arc had a more complex history. The arc lies along a zone of crustal weakness that existed from at least Windermere time and probably from Belt time up until well into the Mesozoic. Within this zone of crustal weakness accumulated an anomalously great thickness of volcanic and coarse clastic rocks, rocks that are almost the only record of the epeirogenic movements to the east.

The first uplift, or epeirogeny, of the Belt terrane that can be recognized in the Metaline district is what White (1959, p. 62-63) refers to as the "East Kootenay orogeny,"

an event recorded in the hiatus between the Purcell Series (Fig. III-4) and the overlying Windermere Group. Although the unconformity that records this event has a stratigraphic relief of many thousands of feet, the angular relations are commonly less than 10°, not the intensity of folding commonly associated with an orogeny.

The term Windermere System, or Windermere Series, is used by Canadian geologists as the time-stratigraphic term for an assemblage of the youngest Precambrian

rocks occurring in southern British Columbia, rocks that unconformably overlie the Purcell Series. The Windermere Series was originally introduced by Walker (1926, p. 13-20) in the Windermere map-area of British Columbia for the assemblage of rocks unconformably overlying the Purcell Series. Originally included were a basal conglomerate, now called the Toby Formation, and the overlying Horsethief Creek Formation, a unit of fine to coarse clastic rocks; but Walker later extended the Windermere

METALINE AND NORTHPORT DISTRICTS WASHINGTON				SALMO DISTRICT BRITISH COLUMBIA			
MISSISSIPPIAN?	Argillites of Russian Creek, Grass Mountain sequence and Flagstaff Mountain sequence						Active Formation
	SILURIAN DEVONIAN	Argillite and reef limestone					
ORDOVICIAN		Ledbetter Slate					
	CAMBRIAN	Metaline Formation Metaline section	Josephine unit	Leadpoint section	Transitional unit	Laib Formation	Upper Laib
Upper limestone			Dolomite breccia		Emerald Member		
Middle dolomite			Middle dolomite		Reeves Member		
Lower limestone			Lower limestone		Truman Member		
Maitlen Phyllite			Reeves Limestone Member		Reno Formation		
PRECAMBRIAN	Gypsy Quartzite						Quartzite Range Formation
	Monk Formation						Three Sisters Formation
	Leola Volcanics						Monk Formation
	Shedroof Conglomerate						Irene Volcanics
	Priest River Group (Belt equivalent)						Toby Formation
						Purcell System	

FIGURE III-4.—Correlation table of stratigraphic units.

to include rocks as young as the Badshot Formation, which is now regarded as Lower Cambrian. A more recent and geographically nearer usage is that of Little (1960, p. 14), who, in describing rocks in the West Nelson map-area of British Columbia, just north of the Metaline quadrangle, restricted the term Windermere System to rocks of late Precambrian age, that is, rocks no younger than the Horsethief Creek Series. In the Metaline quadrangle, Little's (1960) upper limit for the Windermere, the Precambrian-Cambrian boundary, would fall within the Gypsy Quartzite, the oldest formation containing Lower Cambrian fossils. In this report, for reasons of simplicity, I arbitrarily place the upper limit of the Windermere at the base of the Gypsy Quartzite and the top of the Monk Formation.

In the Metaline district, formations of Windermere age are a basal conglomerate, the Shedroof conglomerate, and the succeeding Leola Volcanics and Monk Formation, which grades upwards into the Cambrian Gypsy Quartzite. Twenty miles to the south in the Loon Lake quadrangle, the Gypsy equivalent, the Addy Quartzite, rests directly on Belt-equivalent rocks. The missing 12,000 feet or more of section may be interpreted as lost by erosion from a local positive area during late Windermere time; or the area with the missing section may be interpreted as an allochthonous tectonic block brought in, in post-Cambrian time, from an area where Windermere rocks were never deposited.

The rocks of early Windermere age, the conglomerate and greenstone, represent a change from the Belt environment of shallow continental seas and extensive mud flats to the eugeosynclinal environment of deep water and submarine volcanism; whereas the rocks of late Windermere age, the Monk Formation, being depositionally gradational to Cambrian lithology, represent a change to the miogeosynclinal environment of clean sands and mud-free quartzites. The contact between the Leola Volcanics and the Monk Formation is everywhere sharp and is locally marked by conglomerate; it is probably a disconformity.

The conglomerates of the Windermere are distinctive for their bimodal character, having a fine phyllitic matrix associated with coarse angular clasts, and their great thickness, a minimum of 1 mile and possibly 2 miles. Clasts, many measured in feet, are derived largely from the quartzite, dolomite, and limestone of underlying Priest River Group rocks. Some believe their characteristics favor an origin by glacial transport, others, transport by turbidity currents; but regardless of the agency of transport, their site of accumulation requires a nearby northeasterly elongate positive source area of some relief and a parallel, rapidly subsiding, perhaps troughlike, place of accumulation. All of this suggests related and contemporaneous faulting. The close spatial association between the Leola Volcanics and intrusive greenstones is compatible with the hypothesis that fault conduits controlled extrusion and that the Leola Volcanics are related to a Precambrian fault zone. The lavas locally have pillow structure and contain amygdules and are, according to Fred K. Miller, tholeiitic basalt in composition.

The Monk Formation as described by Park and Cannon (1943, p. 11-13) is a "wastebasket" formation. Park and Cannon say:

The Monk Formation is recognized with confidence only where it can be viewed as a whole or in its normal relation to either the Leola Volcanics or the Gypsy quartzite. The details of lithology shown by nearly any hand specimens or within a limited area of exposure find their analogs in nearby younger and older rocks, particularly in the rocks of the Priest River Group.

Nevertheless, it is the polymerous character that sets it apart from the other formations. Although the formation is predominantly phyllite, it contains numerous intercalations of gritty limestone, structureless quartzite, pebble conglomerates, and basal conglomerate that thins southwestward.

The Monk grades upward into the Gypsy Quartzite; Park and Cannon (1943, p. 11) made the separation where quartzite predominates over phyllite, but locally they were able to approximate the same horizon at the top of a sandy limestone, the host rock of the Oriole mine. There is thus some suggestion that the Gypsy "came in with a limestone and went out with a limestone," because the upper transition from quartzite to phyllite is also terminated with a limestone unit (the Reeves Limestone Member), which separates the Gypsy from the quartzite-free phyllites of the overlying Maitlen Phyllite. The quartzite, which ranges in thickness from one to almost 2 miles, is remarkably pure; only at the bottom and at the top of the formation are phyllite interlayers common, and even here the quartzite layers seldom contain more than a few percent of nonsilica minerals. Most of the rock is coarse grained, and pebbly conglomerate is common in the lower part of the formation. Crossbedding suggests a shallow-water deposition.

The limestone that lies above the Gypsy Quartzite, the Reeves Limestone Member of the Maitlen Phyllite, is important both economically and geologically. It is the important host rock for the lead-zinc deposits of the Kootenay Arc in British Columbia; but although it is widely distributed in northeastern Washington, it is mineralized only in the Red Top Mountain area of the Northport district. As a geologic marker, it is of equal importance: it can be traced for at least 200 miles north of the International Boundary; it is known in the Salmo district as the Reeves Member of the Laib Formation and in the Duncan Lake area as the Badshot Formation. As a stratigraphic marker, it is of double importance because it is fossiliferous: on the Sheep Creek anticline of British Columbia and Washington, it contains a rich assemblage of archaeocyathids, judged to be Early Cambrian by Okulitch (1948, p. 340-344).

Although the Reeves Limestone is the best, if not the only, regional stratigraphic marker in the Kootenay Arc, it is not without area variation. In the Metaline quadrangle, this limestone unit, though recognized by Park and Cannon (1943, p. 15) and by Dings and Whitebread (1965, p. 7), was never formalized with a name nor indicated on the geologic maps. Park and Cannon describe it as a "gray-white limestone about 200 feet thick." The band recognized in the Metaline quadrangle continues

westward into the Colville quadrangle, where it was mapped in the Deep Creek area by Yates (1964) as a limestone unit 400 feet thick. Yates correlated it with the Reeves Member of the Laib Formation (Canadian equivalent of the Maitlen Phyllite) (Fyles and Hewlett, 1959, p. 25-26) and gave it the amended name, Reeves Limestone Member of the Maitlen Phyllite (see Fig. III-4). Here in the eastern part of the Deep Creek area, it is overlain by several thousand feet of phyllite, which comprises the remainder of the formation. West of the Leadpoint fault, however, the limestone unit that immediately overlies the Gypsy Quartzite is not itself overlain by a thick phyllite sequence but by an alternation of thin phyllite and limestone units, none of which, except for the uppermost limestone, is more than 200 feet thick. Unfortunately, the entire limestone-phyllite sequence is shown on Yates' (1964) map of the Deep Creek area as belonging to the Reeves Limestone Member. In the light

of later mapping to the west in the Northport quadrangle (NW ¼ of Colville 30' quadrangle), Yates (in press) found that this same limestone-phyllite sequence was overlain by a thick carbonate section that is correlated with the Metaline Formation, the unit that overlies the Maitlen Phyllite in the Metaline quadrangle. The limestone-phyllite sequence therefore represents the whole of the Maitlen; to be consistent, the name Reeves should be restricted to the lowermost limestone unit.

The Maitlen Phyllite is gradationally succeeded by the Metaline Formation in both the Metaline and Colville quadrangles. The Metaline Formation, the great ore-bearing formation of northeastern Washington, being of economic importance, is described in greater detail than all other rocks. In the Metaline district, it is divisible into three stratigraphic units (Dings and Whitebread, 1965, p. 10): a lower thin-bedded limestone-shale sequence, an intermediate light-gray bedded dolomite, and

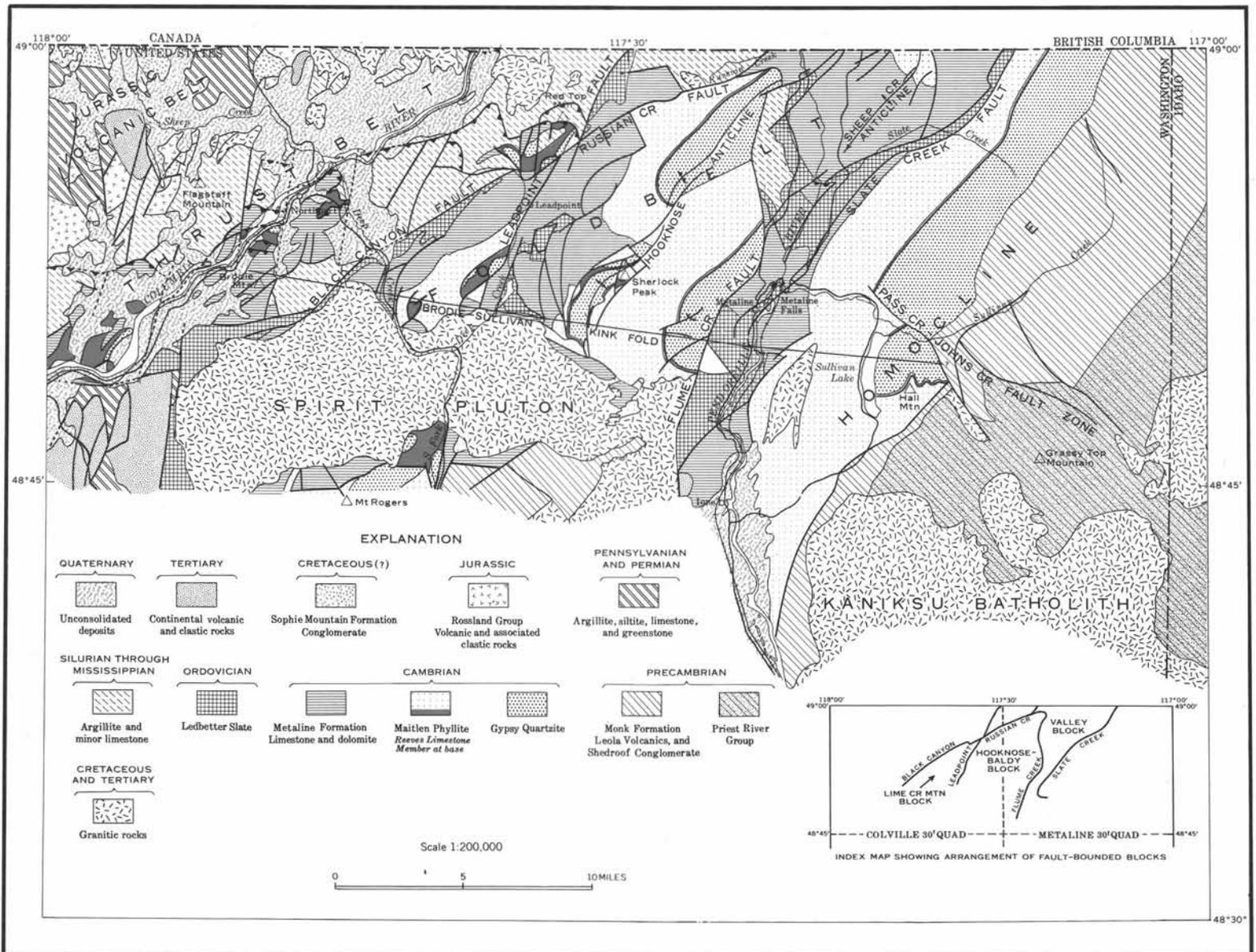


FIGURE III-5.—Geologic map of north half of Colville and Metaline 30-minute quadrangles.

an upper gray massive limestone. The general section of the Metaline mining district is also present in the Deep Creek area, but only on the west side of the Leadpoint fault (Fig. III-5); this section is hereafter referred to as the "Metaline section." East of the Leadpoint fault, the Metaline Formation has the same lower and intermediate units, but the upper unit is an intraformational dolomite breccia instead of a limestone. This section will be called the "Leadpoint section." The spatial distribution of these two sections cannot be interpreted logically unless pre-fold decollement thrust faults are postulated.

Descriptions of the Metaline Formation that follow are largely those of Dings and Whitebread (1965, p. 9-21) for the Metaline quadrangle. Minor differences in thicknesses and lithologies between the rocks in the Metaline district and those in the Deep Creek area are largely disregarded.

The lower, dolomitic limestone is the only unit in the Metaline section that is transitional to another lithology; contacts between the middle dolomite and limestones of the upper and lower units are sharp, as is the contact of the upper unit with the Ledbetter Slate. The interbedding of shale (phyllite) and limestone occurs not only in the lower part of the lower unit but also near the top, where shale may total as much as 10 percent. Although most of the 1,000 feet of the lower unit is composed of a dark-gray thin-bedded limestone and limy shale interbeds, there is a subunit of wavy-bedded limestone in which reddish-brown shaly limestone separates elongate eyes of gray limestone. To the west in the Deep Creek area, this lithology dominates the lower unit in both the Metaline and Leadpoint sections.

The lower unit has never yielded any lead-zinc ore and is not considered favorable for prospecting. It is, however, the source of both limestone and shale for the plant of the Lehigh Portland Cement Company at Metaline Falls. The Lehigh quarries have yielded, in addition, a Middle Cambrian trilobite fauna.

Dings and Whitebread (1965, p. 13) describe the middle, bedded dolomite unit as, "largely composed of a creamy-white to light-gray, fine- to medium-grained dolomite that shows no recognizable stratigraphic variations over wide areas." Lithologic varieties include a black and white banded dolomite and a sublithographic banded dolomite. Chemically, the rock approaches very closely the composition of the mineral dolomite. Grains of clastic quartz are rare, but vug fillings and irregular masses of introduced quartz are common. Bedding is not everywhere obvious. In the Metaline district, only a few ore bodies occur in the middle dolomite, whereas in the Northport district, it is the most productive stratigraphic unit.

The upper unit of the Metaline Formation, the gray limestone unit, although only a minor producer of lead and zinc in the Northport district, is by far the most productive stratigraphic unit in the Metaline district. This upper unit, from 1,300 to 1,500 feet thick, is basically a pure limestone, probably averaging 99 percent calcium carbonate; but in many places, particularly in mine workings, this purity is obscured by intense dolomitization, and, in and near the ore bodies, by silicification. In addi-

tion to the silica introduced during mineralization, the uppermost 200 feet of the unit contains nodular quartz, probably recrystallized chert. Dings and Whitebread (1965, p. 18) say, "rock typical of the gray limestone unit is a light- to medium-gray irregularly mottled very fine grained soft massive limestone." Rarely can individual beds be recognized, but within the upper 25 feet of the formation, bedding is indicated by shaly partings and thin shale beds, an indication of the abrupt change from deposition of carbonate rocks to deposition of black slates of Ordovician age that overlie the upper unit.

From the upper limestone unit of Dings and Whitebread (1965), McConnell and Anderson (1968, p. 1467-1468) have split off a subunit they named the Josephine Unit after the Josephine ore horizon that it includes. This subunit makes up the uppermost part of the Metaline Formation and is "an irregularly brecciated stratum" ranging in thickness from a few feet to more than 200 feet. McConnell and Anderson describe the Josephine Unit as "a breccia of gray limestone, coarse light-gray dolomite, and zebra rock . . . embedded in a matrix of black dolomite or black jasperoid or some mixture of these two." They relate the dolomite, jasper, breccia, and sulfides to the sedimentary and diagenetic processes in a reef environment. In contrast, Dings and Whitebread believe all these features to be products of hydrothermal processes (1965).

In the Metaline district an undeformed contact between the Metaline Formation and the overlying Ledbetter Slate is rarely seen, but considering the deformation the district has undergone, movement is expectable at the junction of such contrasting rocks. Regionally, there is no angular discordance between the two formations; if a hiatus exists, it is a disconformity. If there is no hiatus, the upper part of the Metaline Formation extends into the Ordovician, because Middle Ordovician graptolites occur only a few feet above the limestone-slate contact.

The Leadpoint section of the Metaline Formation consists of the same two lower units as the Metaline section, a lower limestone and the middle dolomite, and two upper units endemic to the Leadpoint section—an intraformational dolomite breccia and an overlying upper unit of dolomite breccia, limestone, and dark argillite that is transitional to the Ordovician Ledbetter Slate. The lower limestone is largely wavy bedded limestone containing several thin subunits of black dolomite. The middle dolomite, differing only in detail from that of the Metaline district, grades into the intraformational breccia unit, a monomineralic rock of light-gray to white dolomite fragments set in a matrix of darker dolomite. Layers of breccia alternate with layers of unbrecciated dolomite similar to that of the middle dolomite unit. Mud cracks support the interpretation of a shallow-water origin. This unit extends into the northwest corner of the Metaline quadrangle, where Park and Cannon (1943, p. 22) mapped it as "Devonian?"

The uppermost transitional unit, which has a thickness greater than 2,000 feet and contains lithologies of all lower units, as well as slate similar to that of the overlying Ordovician Ledbetter, is stratigraphically a puzzling

unit. Nowhere can it be seen in contact with the probable underlying breccia unit. Its contact with the overlying Ledbetter is conformable and gradational through a short distance. In the Leadpoint section, the change from carbonate deposition to black mud deposition was not the sudden irrevocable change of the Metaline section but a more gradual adjustment to changing conditions that began before carbonate deposition had ended.

The Ordovician Ledbetter Slate is a black fine-grained clastic rock that locally has faint tonal bands to indicate bedding and commonly has one or more cleavages or fractures. Most of the slate is noncalcareous, but in places there are dark fine-grained limestone beds. Graptolites are found near the base of the section as well as at indeterminate distances above the base. In the eastern part of the Metaline district, a band of Ledbetter Slate extends southwestward from the International Boundary down Slate Creek to the Pend Oreille River and thence down the river to the Pend Oreille mine. This slate band contains lenses, beds, and pods of fine-grained, black to light-gray quartzite. The quartzite does not appear in the slate west of the river.

The Ledbetter Slate west of the Pend Oreille River occurs in close association with rocks that contain both Silurian and Devonian fossils. Near Beatty Lake, graptolites of Middle Ordovician age occur in slates a few score feet above the top of the Metaline Formation; 2,200 feet stratigraphically above these fossils, in bleached and weathered silty slate, the graptolites are Silurian (Dings and Whitebread, 1965, p. 29-30). Exposures between the two fossil horizons are poor, but float indicates the rock to be dominantly black slate with some light-colored silty porous beds in the upper half. A short distance above the Silurian beds is a bed of black chert, and not far above are lenses of bioclastic limestone, rich in fossil trash. Farther north and to the south occur other richly fossiliferous limestone pods and lenses that have all the characteristics of a reef of biohermal origin. Where determinable, the fossils are Middle Devonian in age.

These bioherms are surrounded by black slate: slates both underlie and overlie them. Within the black slates above the limestone are thin beds and lenses of pebble conglomerate composed largely of chert and quartzite clasts in a siliceous matrix. The paradoxical occurrence of conglomerate and limestone in the black slate environment suggests two disparate provenances. I suggest that black muds were being deposited continuously from the Ordovician through the Devonian in a relatively deep trough and that the muds were transported by longshore currents from an unknown provenance, probably to the north. The coral-bearing limestones, which clearly accumulated in shallow waters, were formed east of the trough on a shelf flanking a positive area that was just beginning to rise in the panhandle of Idaho. Continuous uplift of this positive area and contemporaneous sinking of the trough resulted in unstable conditions along the margin, causing blocks of limestone to slide off into the muds of the trough. As the uplift of the Idaho high continued into the Late Devonian, hypothetical sandstone and chert of the Ordovician and Silurian were exposed to erosion. Gravels derived from these rocks were swept

across the platform on which the limestones had accumulated into the deeper water mud environment of the trough. Turbidity currents may have helped move the gravels downslope into the trough.

The fossil record preserved in the Metaline district ends with the Devonian, although it is quite conceivable that the black argillite west of the Pend Oreille River in the drainage of Russian Creek—argillite mapped as Ledbetter Slate by Dings and Whitebread (1965, p. 26)—includes beds as young as Mississippian. If it does, the Russian Creek lithology contrasts sharply with the Mississippian 65 miles to the south in the Loon Lake quadrangle (Miller, 1969, p. 3-4), where a shallow-water carbonate facies is represented. In this same area, fossils of Pennsylvanian age (Enbysk, 1954, p. 15) have been found, as well as fossils of Devonian age (Miller, 1969, p. 4), both in carbonate rocks. It is probable that during the late middle Paleozoic, the depositional facies of the Metaline district was black mud, whereas that to the south was carbonate.

Rocks similar to the argillites of the Russian Creek area are found in the Deep Creek area, where they were mapped under the name Grass Mountain sequence (Yates, 1964), and in the Northport quadrangle, where they were mapped as the Flagstaff Mountain sequence (Yates, in press).^② There is no fossil evidence for the age of these rocks; they are regarded as upper Paleozoic because they fit most closely into this part of the section. The reasoning behind this assignment is as follows: In both Metaline and Deep Creek areas it is established by fossils that the deposition of black fine-grained clastic continued from the Ordovician into the Devonian. In this part of the Kootenay Arc, there is a "hiatus" in the fossil record; the next youngest fossils are probably Pennsylvanian, found in the Rossland Map area (Little, 1960, p. 45-51) in British Columbia across the International Boundary from the Northport quadrangle. The Mount Roberts Formation, in which these fossils occur, consists of an assemblage of predominantly fine clastic rocks intercalated with pyroclastic debris and mafic flow rocks as well as the limestone pods that contain the fossils. These are the oldest Phanerozoic eugeosynclinal rocks in this part of the Kootenay Arc. Nearby Permian, Triassic, and Jurassic rocks also belong to this facies. In general, the younger the rocks, the greater the amount of volcanic material. Volcanic material, though rare, is nevertheless present in the argillite sequence, which appears to be transitional between the middle Paleozoic black slates and the upper Paleozoic volcanic assemblages.

This interpretation considers the transition between the argillite sequence and the eugeosynclinal rocks as changes through time, but lateral changes in lithologies, foreshortened by thrusting, may also be significant.

We know nothing of the former eastern limits of the Pennsylvanian, Permian, Triassic, and Jurassic rocks. Rocks of these ages do not occur in the Metaline quadrangle nor have they been found in the downfaulted wedges in the Belt anticlinorium to the east; consequently,

^②These argillites, those in the Russian Creek area of the Metaline quadrangle and those in the Deep Creek area, are mineralized by quartz veins that carry pockets of sphalerite and galena and silver-bearing tetrahedrite. The veins, which occur in low-angle shear zones in the argillite, are cut by lamprophyre dikes dated as middle Eocene (Yates and Engels, 1968).

there is no basis for assuming their former presence across the Metaline quadrangle. However, it seems highly unlikely that the Metaline quadrangle and the anticlinorium were above sea level and being eroded during the 300 million years since the Mississippian Period. On the other hand, the lack of any mafic intrusive bodies that might have acted as feeders for the volcanic rocks of the eugeosynclinal facies supports the belief that this facies never accumulated over the Cambrian miogeosynclinal terrane of the Metaline quadrangle; any post-Mississippian rocks that might have been eroded from this area would have to be some facies other than eugeosynclinal. The fact that outliers or windows of either facies have not been found in the terrane of the opposing facies argues against the presence of great regional thrusts similar to those described from the Great Basin.

Post-Mississippian sedimentary rocks are not of direct concern in establishing the geologic setting of the lead-zinc deposits, however, as such rocks either were never deposited or, if deposited, were eroded from the area of the Metaline and Northport mining districts. The process of volcanism and cognate tectonism that accompanies the accumulation of rocks of this age to the west of the Northport district is of interest, however, to those who would demonstrate or postulate hydrothermal mineralization as a process related to volcanism. The operation of volcanism in a neighboring area permits one to entertain the hypothesis that processes related to volcanism, such as mineralization, may have operated in the adjacent areas. Of greater importance is the need to go outside the lead-zinc district to areas of younger rocks to establish the age of the tectonic events that folded and faulted the rocks.

The nearest fossiliferous Permian occurs southwest of the Northport quadrangle in the Kettle Falls areas as the Mission Argillite, which is composed of argillite, andesitic volcanic rocks, graywacke, chert-pebble conglomerate, and limestone bioherms containing fusilines of early Permian age (Mills and Davis, 1962, p. 41, 44). On Kelley Hill, a short distance from the Kettle Falls locality, Early Triassic bioherms occur in puzzling close association with Permian bioherms (Kuenzi, 1963, p. 88-89). Closer association of volcanism and the Triassic is preserved to the north in British Columbia in the Kaslo Group in the northeast corner of the Nelson Map area, West Half (Little, 1960, p. 52-53). Rocks of the Kaslo Group are almost entirely of volcanic origin and include flows and pyroclastic rocks of andesitic and dacitic composition. The Jurassic, even more than the Triassic, is characterized by volcanic rocks; although widespread in the Northport quadrangle, the Jurassic sequence is well dated only in British Columbia, where ammonites are found in sedimentary units intercalated with volcanic units. Frebald and Little (1962) have divided the Jurassic rocks of southern British Columbia into four main units, from bottom to top: (1) Archibald Formation, composed predominantly of sedimentary rocks and minor tuffs and flows; (2) Elise Formation, predominantly andesitic to basaltic lava flows, flow breccias, agglomerates and tuffs; (3) Hall Formation, predominantly sedimentary rocks with local accumulations of volcanic material; and (4)

an upper unit called the Upper Rossland Group (the Elise and Hall Formations are the Lower Rossland Group), predominantly volcanic and free of diagnostic fossils. Time represented extends from the lowermost Jurassic stage, the Hettangian (?), to fossil-confirmed ages younger than middle Bajocian and possibly to the lowermost Upper Jurassic stage, the Callovian.

From the Late Jurassic to the Eocene, the geologic record in northeastern Washington is, except for middle Cretaceous plutonism, rather hazy. The only rocks believed to have been deposited during this period are the boulder conglomerates found along the northern edge of the Northport quadrangle and adjacent British Columbia. These conglomerates, the Sophie Mountain Formation, contain fossil leaves on which W. A. Bell made the following comment, "Although a Tertiary age is not ruled out, an Upper Cretaceous age seems more probable." (Little, 1960, p. 74). The Late Cretaceous age seems even more probable now that we know the conglomerates are cut by dikes of Coryell syenite dated as 50 million years old (Yates and Engels, 1968, v. 1, p. D-245).

The Sophie Mountain Formation is commonly a coarse clastic unit that contains boulders as much as 4 feet in diameter; it locally contains thin sand and silt lenses. Roundstones are quartzite, chert, and fine-grained igneous rocks, solidly cemented with quartz. Clasts of coarse-grained granitic rocks of nearby batholiths are absent. Exposures of quartzite and igneous rocks that could be a source for the roundstones are unknown. These conglomerates are the oldest nonmarine rocks deposited in the areas. They clearly indicate that by Late Cretaceous time the orogenic cycle was well advanced.

Although Eocene volcanic rocks were not seen resting on the upper surface of the Sophie Mountain Formation, it is nevertheless certain that the two rock units are separated in time by an erosional interval. At many places the volcanic rocks are floored by moderately consolidated conglomerates composed of clasts of local rocks that are distinctly different from the quartzite-like conglomerates of exotic clasts of the Cretaceous.

The section of Eocene volcanic rocks in the Northport quadrangle consists of the above-mentioned conglomerate overlain successively by a water-laid feldspathic crystal tuff, hornblende-biotite rhyodacite flows with intercalated conglomerate, and biotite-bearing basaltic flows. The accumulation of these rocks was accompanied by high-angle normal faulting.

Similar volcanic rocks of probably Eocene age occur 35 miles south of Metaline Falls in the Newport quadrangle. Schroeder (1952, p. 24-25) has mapped these rocks and named them the Pend Oreille Andesite. The only volcanic rocks that occur within the Metaline quadrangle (Dings and Whitebread, 1965, p. 32-33) are erosional remnants of a flow of olivine trachybasalt that occurs a mile northwest of Ledbetter Lake. No feeder for this flow has been found, but it is quite likely that it is genetically related to the lamprophyre and mafic dikes that occur sparsely but widespread throughout the district.

Overlying the Pend Oreille Andesite in the Newport quadrangle and resting on pre-Tertiary rocks in the valley of the Pend Oreille River in the southern part of the

Metaline quadrangle are poorly sorted, semiconsolidated conglomerates composed of angular clasts derived largely from underlying rocks. Fossil leaves indicate a nonspecific Tertiary age.

All the rocks described above, from Precambrian to Tertiary, are difficult to study because of the Pleistocene glacial deposits of unconsolidated gravel, sand, and silt that mantle the slopes and choke the valleys. Outcrops are covered by two principal kinds of glacial deposits: (1) glaciofluvial deposits and (2) lake or terrace deposits. The glaciofluvial deposits are mainly gravels that cover much of the slope areas as thin mantles a foot or less thick. The lake or terrace deposits cover the valleys of the Pend Oreille and Columbia Rivers.

TECTONIC EVOLUTION

Throughout the preceding account of the sedimentary and volcanic history, I have alluded to tectonic events that influenced, controlled, or determined the facies and distribution of the rocks. To separate the tectonic events from the details of sedimentation, I recapitulate, with the aid of diagramming in Figure III-3, the events that produced the structures of the present terrane. Included schematically among the tectonic events to the left of the time scale are the volcanisms of Precambrian, late Paleozoic, Mesozoic, and Tertiary, for surely these also are tectonic in cause and effect. Perhaps the same defense can be made for the inclusion of the two, possibly three, plutonic events.

The first Precambrian tectonic event is recorded in the unconformity between the Priest River Group and rocks of Windermere age, in the earliest Windermere conglomerate, and by faults that cut Windermere and older rocks but not younger rocks. The unconformity in the Metaline quadrangle and to the southwest in the Chewelah quadrangle is only slightly angular; therefore Precambrian folds are interpreted as gentle warps. The Precambrian faults were subject to the intense folding of the Mesozoic orogeny, and all but those normal to Mesozoic fold axes are consequently difficult, if not impossible, to recognize. Fortunately, the orientation of some faults was such that they escaped folding during the Mesozoic. A good example is the southeastern part of the Pass Creek-Johns Creek fault zone mapped by Park and Cannon (1943, plate 1) (see Fig. III-5). Here the boundary between the Priest River Group and the Shedroof Conglomerate has an apparent right-lateral offset aggregating about 4 miles; however, the boundary between the Shedroof and overlying Leola Volcanics projects across the fault zone without an offset. The reduction of the outcrop width of the Shedroof Conglomerate from 5 miles to less than 1,000 feet indicates a post-faulting, pre-Leola erosion. The fact that the northwestern end of the fault zone, the John Creek fault, cuts Cambrian rocks is of no significance, because the offset on this fault is opposite in direction and younger than the northeast folds of the Kootenay Arc. Miller's map of the Loon Lake quadrangle (1969) shows several faults that cut Precambrian rocks but not the Cambrian Addy Quartzite.

The rocks of Windermere age, coarse conglomerates and mafic volcanic rocks, are indicative of tectonism. The unconformities within the sequence also clearly indicate that this period was one of unrest and that a positive provenance existed not far to the east. Stability, however, was attained before the end of the Windermere, because the Monk Formation represents the beginning of the stable miogeosynclinal environment that continued through the Cambrian. The tectonic harmony of continued uninterrupted subsidence of the miogeosynclinal trough continued until well into the Paleozoic.

The first suggestion of a late Paleozoic diastrophism is the appearance of pebble conglomerate of possible Early Mississippian age in the black shale sequence northwest of Ledbetter Lake. If, as suggested, these conglomerates represent the development of a positive area in the Idaho panhandle, they could also identify the time of the juxtaposition of the two different sections of the Metaline Formation, which I interpret to be the result of a westward-moving decollement thrust that originated in the eastern part of the Metaline quadrangle and adjacent Idaho. By this device the Leadpoint section of the Metaline Formation is believed to have overridden the Metaline section. However, there is no compelling evidence that decollement thrusting occurred at this time; it could have occurred at any time after the Devonian and before the Mesozoic, probably Late Jurassic, folding took place. The two requirements for the thrusting are a positive area in the Idaho panhandle and essentially unfolded rocks.

The suggestion of a Triassic disturbance in the Kootenay Arc along the International Boundary comes from the relations reported by Little (1960, p. 66-67), who described basal flow breccias of the Rossland Formation (Lower Jurassic) resting on different members of the Pennsylvanian(?) Mount Roberts Formation. Although this identifies the hiatus only as pre-Jurassic, a Triassic age is regionally more acceptable than an older age. Consistent with this suggested Triassic tectonism are the 200-million-year radiometric dates obtained by Miller and Engels (personal communication, 1970) for the Flowery Trail pluton in the Chewelah Mountain quadrangle, southwest of the Metaline quadrangle. It seems quite possible that there are rocks of this age in the sparsely dated Kaniksu batholith.

At best, any tectonic disturbance in the Triassic is a very minor disturbance indeed, compared with the great Mesozoic orogeny that involves the Jurassic and all older rocks. This great folding in the very late Jurassic or Early Cretaceous was certainly over by the time the granodiorite of the Spirit pluton was emplaced, 100 million years ago. The folds produced are northeasterly in trend and, where asymmetric, are steeper or overturned on their northwest limbs (Fig. III-5A). Their axes are horizontal over long distances, although locally they plunge to the southwest. Although the pattern of these folds has been modified by crossfolding and interrupted by the block-faulting of the Cretaceous and Tertiary, it is basically the pattern of the Kootenay Arc.

Up to this point, I have presented the depositional and tectonic history of the area as a sequence of events

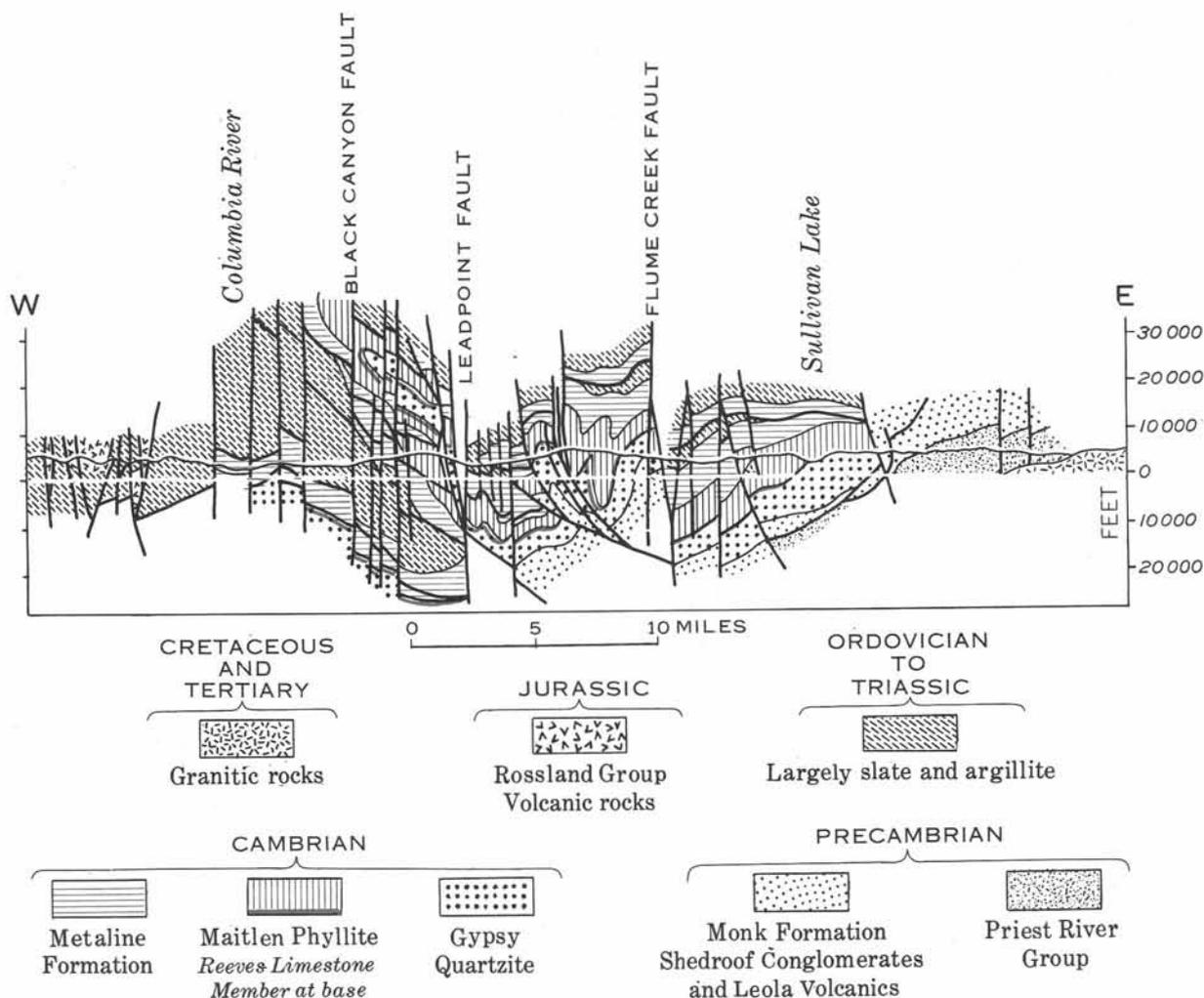


FIGURE III-5A.—Geologic structure section across Colville and Metaline quadrangles.

and have made no attempt to show on a geologic map the interrelated consequences of these events. In the recounting of the geologic history, we have now arrived at a stage where we are ready to relate event to outcrop and describe the geology in terms of features on the map (Fig. III-5).

South of the International Boundary, the Kootenay Arc is divisible into four tectonic units, from east to west: (1) homoclinal belt, (2) fold belt, (3) thrust belt, and (4) a Jurassic volcanic province. The first three belts represent different intensities of compressional deformation in the Mesozoic. The Jurassic volcanic province, named from the youngest of the eugeosynclinal assemblages that were deposited in this area, is a tectonic division only in the sense that it represents the distinctive volcano-depositional tectonism of the eugeosyncline. The homoclinal belt³ consists of the northeast-striking, northwest-dipping Precambrian and Cambrian rocks of the Metaline quadrangle southeast of the Slate Creek fault, a steep reverse strike fault. The fold belt consists of lower and

middle Paleozoic rocks and is the southwest extension of the group of folds concentrated about the Sheep Creek anticline of British Columbia. The thrust belt, in a strict sense, is the part of the fold belt that has been deformed by a later compression, which buckled and thrust-faulted the earlier northeastward-trending folds along approximately east-west axes. This thrusting is separate and distinct from the pre-fold thrusting. The boundary between fold and thrust belts is arbitrary and tenuous. Both the homoclinal belt and the Jurassic volcanic province are outside the lead-zinc area; consequently, they are described only as they relate to the structural development of the area.

The fold belt was subdivided during the Late Cretaceous and Eocene into blocks by high-angle faults trending N. 10° - 20° E. The easternmost block, named the valley block by Park and Cannon (1943, p. 28), is separated from the central, Hooknose-Baldy block by the Flume Creek fault. The Hooknose-Baldy block, in turn, is separated from the Lime Creek Mountain block by the Leadpoint fault. These faults terminate to the north against a N. 60° E.-trending fault that crosses the International

³The homocline is probably the western limb of the westernmost anticline of that concentration of folds that compose the Belt-Purcell anticlinorium that extends eastward into Idaho and Montana; but as the eastern limit of this fold has not been identified, homocline seems the proper term.

Boundary just west of the Pend Oreille River to join with the Black Bluff fault of British Columbia.

The valley block is the broad southwestward-plunging nose of the Sheep Creek anticline (the Boundary anticline of Dings and Whitebread, 1965, p. 34), the structure that hosts the Metaline lead-zinc district. A local reverse plunge on this fold is the site of the Pend Oreille and Grandview ore bodies. This structure is broken by innumerable faults, some related to the Mesozoic folding, others to the normal faulting of the early Tertiary. The faults that bound the block, the northeast-trending Slate Creek fault and the N. 10° E. Flume Creek fault, were interpreted by Park and Cannon (1943, p. 28-33) as normal faults bounding a triangular-shaped graben. Dings and Whitebread (1965, p. 41), and more recently McConnell and Anderson (1968, p. 1470), interpret the Slate Creek fault as a southeast-dipping reverse fault.

This interpretation is more compatible with the regional structural pattern; it permits the Slate Creek fault to result from the same compressional forces that produced the northeast folds of the Kootenay Arc. On the other side of the "graben," the bounding Flume Creek fault appears to be a true "graben fault" along which the valley block has moved downward a distance estimated to be as little as 12,000 feet and as much as 20,000 feet. The Hooknose-Baldy block is dominated by an upfaulted segment of the Sheep Creek anticline, here called the Hooknose anticline. This part of the fold is tight and asymmetric, with an axial plane that dips to the southeast and an axis that plunges southwestward. It is flanked by a minor fold on the southeast and has a complexly folded and faulted western flank. The western part of the Hooknose-Baldy block contains the eastern half of the Northport mining district; most of the western half of the district is in the Lime Creek Mountain block. The Hooknose-Baldy block is terminated to the west by another high-angle fault, the Leadpoint fault, on which relative movement is down on the east and up on the west. Displacement on this fault is comparable to that on the Flume Creek fault, of the order of 20,000 feet.

The Lime Creek Mountain block is the southeastward-dipping overturned limb of a fold composed largely of Cambrian rocks. The Lime Creek Mountain block is separated from the thrust belt by the Black Canyon fault, a high-angle fault of uncertain character. The block terminates to the south against the Spirit pluton, a granodiorite mass about 100 million years old which clearly was intruded after the folding.

The thrust belt combines the extensions of what Fyles and Hewlett (1959, Fig. 2) called, in the Salmo lead-zinc area of British Columbia, the "Mine Belt" and the "Black Argillite Belt." All lead-zinc mines in the Salmo district are in the mine belt; a few on Red Top Mountain in the Northport district are in the thrust belt.

The thrust faults in the Washington part of the thrust belt are faults that are later than, and apparently unrelated to, the folds of the fold belt. The strikes of the faults and the axes of related buckles and kinks are within 10° to 15° east-west; dips are commonly to the south, a few are to the north. The strikes approximate the trends of the axes of the folds in the complex of Monashee Gneiss

in the Vernon area (Jones, 1959) of the Shuswap Terrane, as well as the trend of the belt of the Permian Cache Creek Group that extends across the terrane as the western extension of the Slocan fold, an overturned synclorium north of the Nelson batholith. The north and south boundaries of the Nelson batholith also follow this trend. All these compressional features along east-west axes are as old as, or are older than, the Nelson and Kaniksu batholiths. If the Waneta fault, which lies largely in British Columbia and cuts Lower Jurassic rocks, is a fair representative of the deformation, the maximum age of this east-west compressional event is Late Jurassic, the minimum, Early Cretaceous.

The east-west features are not restricted to the thrust belt, but are regional. One in particular that is very useful in establishing the ages of the late tectonic events extends from the thrust belt across the fold belt and into the homoclinal belt. This feature, here called the Brodie-Sullivan kink fold, extends from Brodie Mountain, 3 miles south of Northport, S. 80° E. past Sullivan Lake and Hall Mountain for a traceable distance of 30 miles. This line of deformation extends west of Brodie Mountain as a south-dipping thrust fault. More detailed geologic mapping east of Hall Mountain is necessary to determine the eastern extent of the zone beyond Hall Mountain.

This kink fold is a short-limbed fold superimposed on the northeast fold system and is a giant representative of the kink bands common in the phyllitic rocks in the thrust zone along the Columbia River south of Northport. The axial plane of the fold strikes N. 80° W. and appears to have a steep southerly dip. The kink bands and the kink fold are believed to have resulted from the same stress that produced the thrust faults. The kink bands, chevron-shaped, contrast with the kink fold, which lacks the angularity of the bands and is a smoothly curved fold formed by the rotation of the northeast-striking beds, fold axes, and cleavage counterclockwise 90° about a vertical axis or near-vertical axis. Minor folds within the kink fold are vertical or plunge steeply.

The Brodie-Sullivan kink fold obviously is younger than the northeast fold system it deforms; is older than, or contemporaneous with, the high-angle north-trending faults, which are not bent by the kink fold, as, for example, the Flume Creek fault; and is older than the 100-million-year-old Spirit pluton, which has engulfed the southern part of the fold. The kink fold, therefore, enables us to establish the following post-Late Jurassic sequence of events: (1) folding along northeastward-trending axes, (2) thrusting and kinking along east-west axes, and (3) high-angle, "graben faulting." It is not so easy to time the emplacement of the Spirit pluton and Kaniksu batholith within this sequence. The plutonism is clearly younger than the thrusts, but the question of whether it is older or younger than the high-angle faulting has never been settled. Park and Cannon expressed this uncertainty as early as 1943 (Park and Cannon, 1943, p. 35): "Along the Flume Creek fault west of Lost Valley . . . the batholith may either be faulted or intruded along pre-existing or contemporaneous faults." Although Dings and Whitebread did not extend their mapping this far south, they

believed the evidence favored pre-batholith movement on the fault. They say (p. 39), "South of Ione the fault (Flume Creek) is probably cut off by the younger Kaniksu batholith." The writer tends to agree with the judgment of Dings and Whitebread and believes that the Flume Creek fault, and the Leadpoint fault as well, are older than the Cretaceous plutonic rocks, but this opinion is based on tenuous evidence.

The difficulty in establishing the relations between faulting and plutonism is largely due to the glacial deposits that cover critical areas. The Flume Creek fault northwest of Ione is shown on Park and Cannon's geologic map of the Metaline quadrangle (1943, plate 1) as projected under a cover of glaciofluvial deposits between outcrops of the Kaniksu batholith[Ⓞ] and the Ledbetter Slate; nowhere is the granite rock seen in contact with the slate. The thick cover of unconsolidated deposits in the valley of the South Fork of Deep Creek effectively conceals any possible southern projection of the Leadpoint fault that might be made, but the presence of a strong fault south of the pluton coinciding with such a projection furthers the credibility of this interpretation. However, the lack of offset of either the north or south boundaries of the pluton is indeed remarkable, considering that more than 20,000 feet of displacement is believed to have taken place on this fault. On the other hand, some faults having the same trend as the Flume Creek and Leadpoint faults, notably those in the Williams Lake area of the Colville 30-minute quadrangle, do cut rocks as young as Eocene. Nevertheless, even some of these faults can be demonstrated to be older than the youngest rocks they cut. The best example is the N. 10° E. fault that bounds the volcanic area on the east in the valley of Phelan Lake. Just west of this fault, the Eocene O'Brien Creek Formation rests on, and is intercalated with, a megabreccia composed in places of great blocks, some several scores of feet long, of granodiorite derived from the Spirit pluton one-fourth mile to the east. The megabreccia is a talus-like deposit that accumulated at the foot of the scarp formed by this Phelan Lake fault, which apparently was old enough to allow deroofting of the pluton before the basal breccia was deposited.

Although the Flume Creek and Leadpoint faults are regional, other faults of similar strike are strictly local structures. Many of these local faults appear to be closely related to the thrust faults of east-west trend acting as tear faults in the upper plates. Still others extend through both upper and lower plates and are tear faults in the sense of Hills (1963, p. 208), who applies the term to faults that tear across and are contemporaneous with the folding. As an example, he uses faults described by Jean Goguel (1947, p. 31-33) from the region of the Vanige in the Subalpine Chain of southeastern France. The north-striking faults in the vicinity of Northport are of this type, except that the deformation has been more intense than that in the Vanige; the folds have broken into thrusts. These faults and the east-west folds and thrust they tear across are superimposed on the northeast-trending anticline that extends down the Columbia River.

[Ⓞ]These rocks are considered by the writer as the eastern end of the Spirit pluton.

Some of these faults "tear" across several thrusts and are certainly related to, and contemporaneous with, the folding and thrusting.

Thrust faults and associated tears of this nature are common in the thrust belt, and although individually of no great displacement, they are in the aggregate responsible for the bend of the Kootenay Arc from the S. 20° W. trend in the Salmo district of British Columbia to a trend of S. 60° W. along and south of the International Boundary. A full discussion of this post-arc distortion of the fold belt is not possible here; but some mention should be made of the similarity of pattern of the thrust-tear structures of the Northport area and the Russian Creek-Flume Creek fault structure. The Flume Creek fault at its north end turns abruptly to continue in a S. 75° W. direction as what is known as the Russian Creek fault. Applying the thrust-tear formula to this structure, the Russian Creek fault would be a south-dipping high-angle reverse fault, and the Flume Creek fault the tear fault. The Flume Creek fault cannot be a tear fault of substantial movement, however, because the Brodie-Sullivan kink fold crosses the Flume Creek fault without offset and without kinking the fault—a fact used above to establish the relative ages of the fault and kink fold. If these relations are correctly observed and correctly interpreted, we can logically infer that any genetic relation between the Flume Creek-Russian Creek fault structure and the thrust-tear system of the fold belt requires that the postulated Flume Creek tear fault operated without apparent strike-slip movement and that the fault developed after the kink fold. We are left with the implication that the "graben" faults of Tertiary age and the tear faults of Cretaceous age both are products of movement deep in the basement rocks.

METAMORPHISM AND IGNEOUS ROCKS

All the pre-Cretaceous rocks in the area of Figure III-5 were exposed to a regional greenschist facies metamorphism that occurred during or somewhat after the time of the northeast folding. Superimposed on this background metamorphism belonging to the muscovite-chlorite subfacies are aureoles of contact metamorphism, which extend out for 2 miles or more from the Kaniksu batholith and the Spirit pluton. The regional metamorphism converted shales to slates and phyllites, and it recrystallized quartz sandstones to quartzite. The contact metamorphism had little effect on the quartzites, but coarsened the grain of the phyllites, emphasizing their planar structure and producing in the inner aureole schists and, more rarely, gneissic rocks having mineral assemblages that include biotite, andalusite, cordierite, and sillimanite. The carbonate rocks were recrystallized; and where silica was available, diopside, forsterite, wolastonite, and tremolite were formed.

The granitic rocks that produced the thermally metamorphosed sedimentary rocks underlie about one-third of the area of the Metaline and Colville quadrangles. The larger of two masses, the Kaniksu batholith, extends far outside this area, covering much of the panhandle of Idaho, an area of about 2,500 square miles. The Spirit

pluton, along whose periphery are located a number of mines, is a discordant, east-west elongate granitic body about 90 to 100 square miles in area.

The Kaniksu batholith is a complex of several granitic facies that range between quartz monzonite and granodiorite in composition. In mapping the Metaline quadrangle, Park and Cannon (1943, p. 24) recognized three

facies: one characterized by large microcline phenocrysts, a structureless rock characterized by biotite, and a two-mica rock with a well-developed planar structure. The batholith appears to have been emplaced without assimilating or granitizing its wall rock. Its visible thermal aureole does not extend to the mines of the Metaline district.

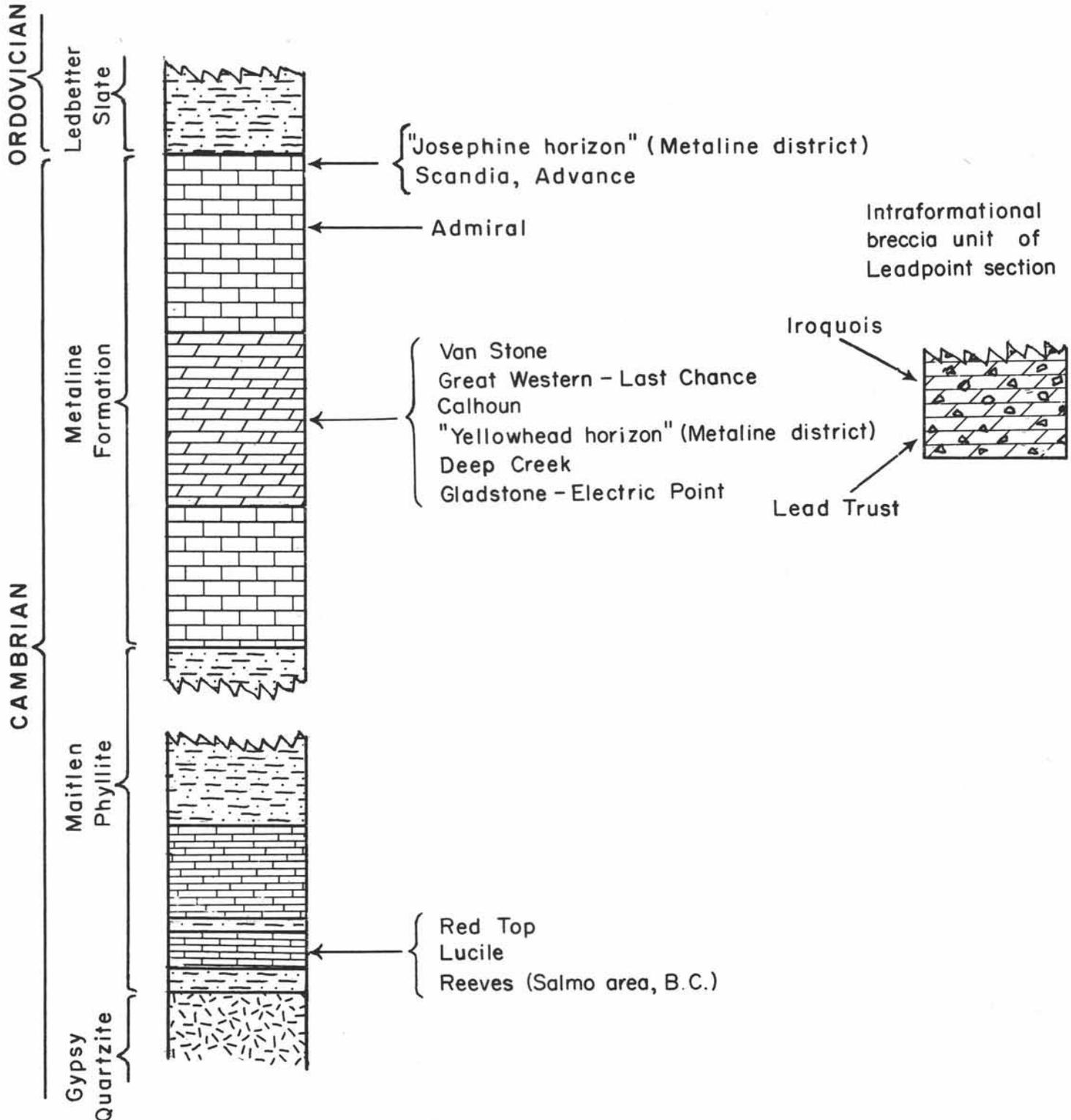


FIGURE III-6.—Columnar section showing stratigraphic hosts of ore bodies.

The Spirit pluton is predominantly a coarse-grained biotite granodiorite. It has, however, both a porphyritic and a nonporphyritic phase; but the porphyritic phase, unlike that of the Kaniksu batholith, has phenocrysts of orthoclase instead of microcline. Remnants of partly digested quartz diorite and dikes and dikelike masses of quartz monzonite and alaskite make up not more than 10 percent of the mass. Its contact aureole is described above. The Deep Creek, Van Stone, and Sierra Zinc mines are in the aureole rocks.

Although the Spirit pluton is separated from the batholith at the surface, it probably is closely related to it and part of the same igneous event, which took place about 100 m.y. ago (Yates and Engels, 1968, p. 242-247).

The other plutonic rock, the Sheppard Granite, named and first described by Daly (1912, p. 354-356), has not been radiometrically dated, but it is considered to be Eocene because it intrudes, just north of the International Boundary, conglomerate believed to be correlative with the Cretaceous Sophie Mountain Conglomerate (Little, 1960, p. 94-95). Lamprophyre dikes of the 50-m.y. event cut the Sheppard Granite and thus place a minimum limit on its age. This granite appears to be unrelated to the process of lead-zinc mineralization.

THE ORE DEPOSITS

Within the geologic framework outlined above, we can with confidence place the individual ore deposits in their

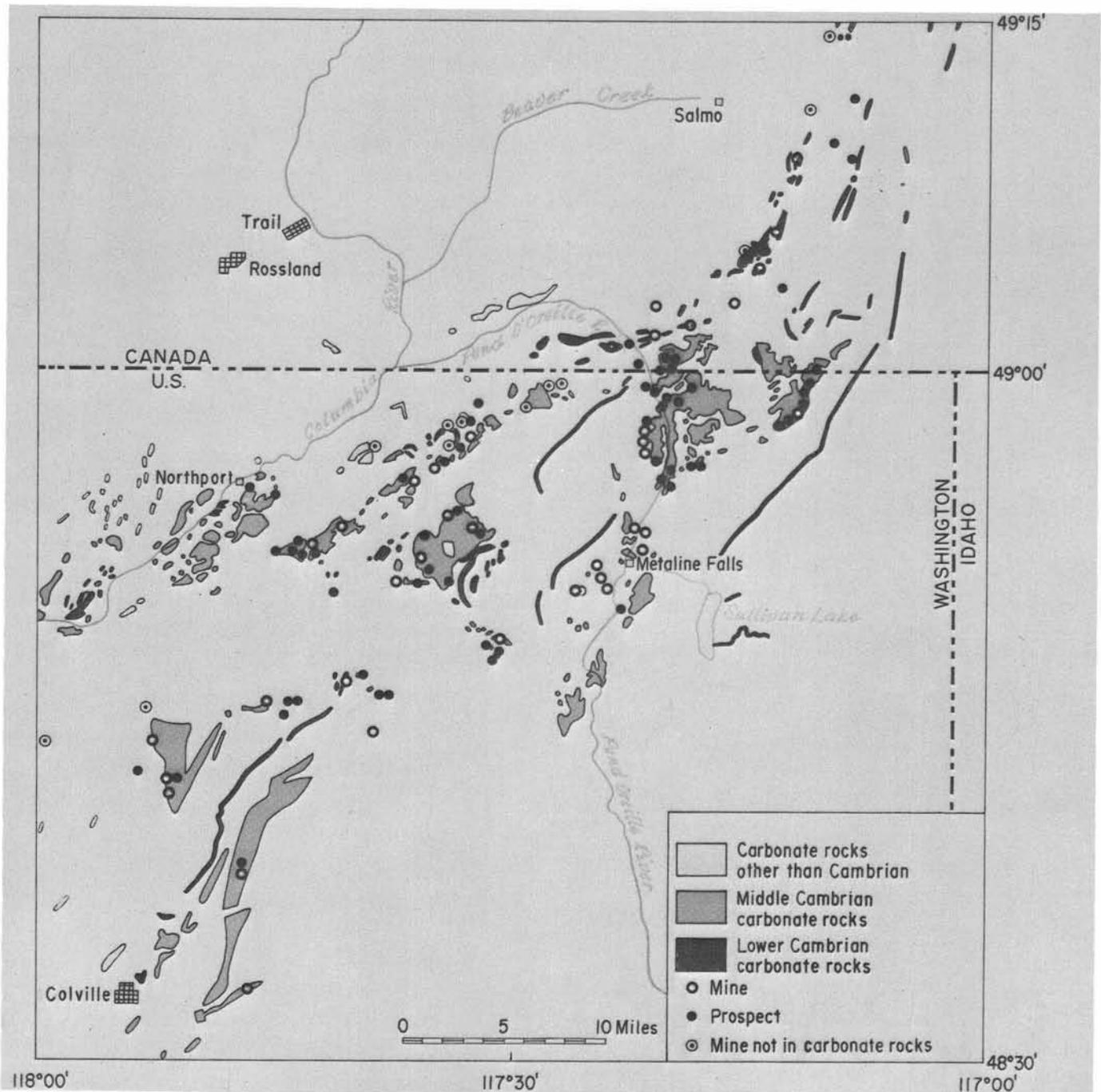


FIGURE III-7.—Map of northeastern Washington and adjacent British Columbia showing relations between mines and carbonate rocks.

proper stratigraphic structural positions; but we cannot unequivocally demonstrate that mineralization occurred at a specific period of time or by a specific process. Although the lead-zinc province lies within the Kootenay Arc, no single outcropping structure of the Arc, fold or fault, controls the province or the districts, or even groups of deposits. Nor are the deposits, if interpreted as syngenetic strata-bound deposits, to be related to any single stratigraphic horizon (see Figs. III-6 and III-7), nor are they, if interpreted as epigenetic deposits, to be related to any pluton or group of plutons (see Fig. III-8). Nor has it been proven that all deposits are of the same age, or, conversely, that different deposits are of greatly different ages.

Although the individual deposits described in this bulletin do not include all types of lead-zinc deposits that occur in the lead-zinc province, they do represent the stratiform deposits where the ores occur in favored horizons in limestone and dolomite. Further description or classification of the deposits will not be attempted. I here point out some geologic relations critical to any genetic interpretation of the lead-zinc deposits.

If we confine our attention to the deposits of the Metaline district, we can favorably entertain McConnell and Anderson's proposal (1968, p. 1476-1479) that the deposits of the Josephine Horizon are syngenetic and not the product of hydrothermal processes as proposed by Park and Cannon (1943, p. 53) and Dings and Whitebread

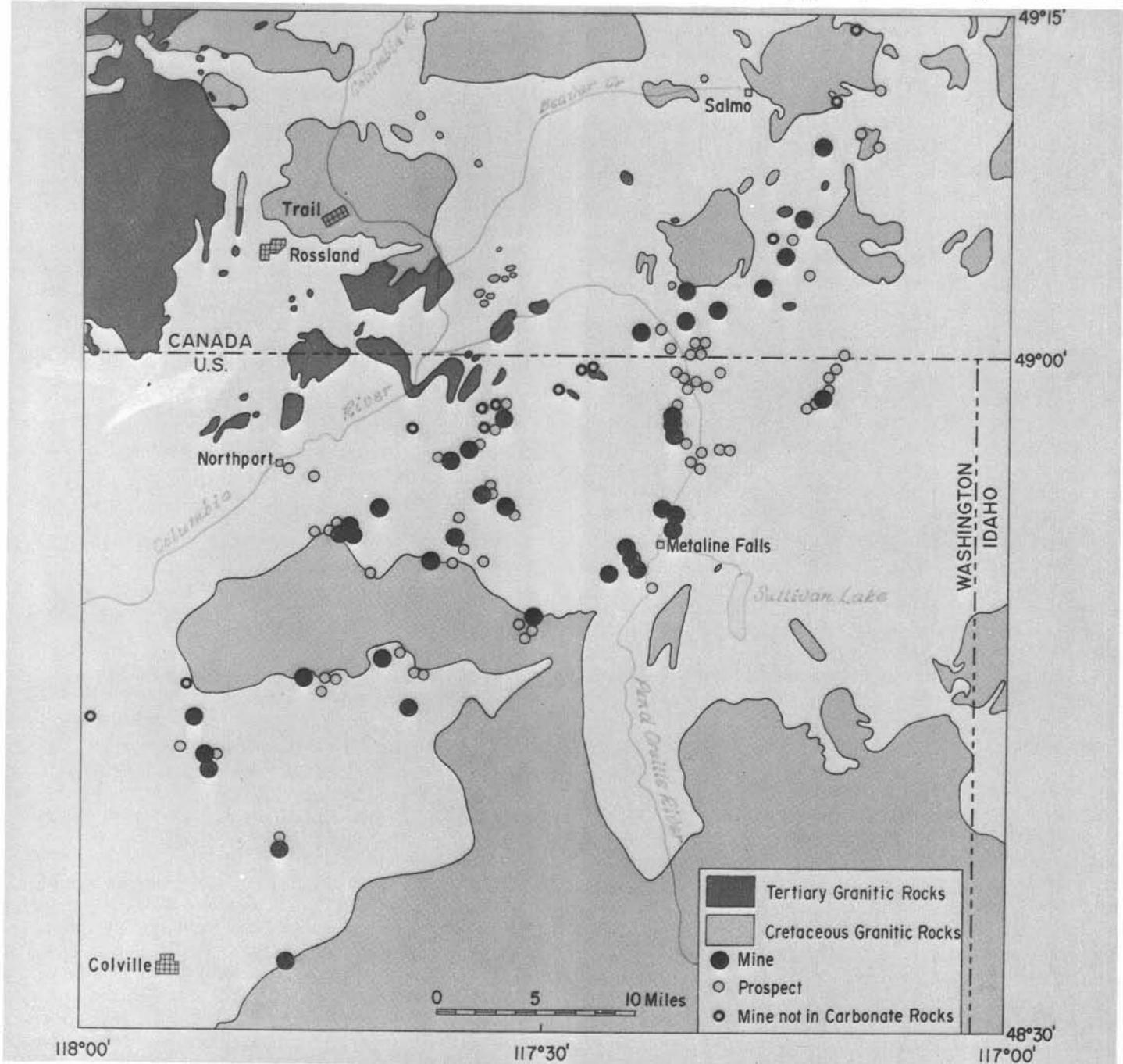


FIGURE III-8.—Map of northeastern Washington and adjacent British Columbia showing relations between mines and

(1965, p. 68-69). The Josephine Horizon is the principal, but not the only, ore horizon in this district (see Fig. III-6). To broaden our appreciation of the complexity of the problem, let us leave the Metaline district and take a look at some features in the Northport district. We find that although there is here much greater diversity in host strata and structures, geologic settings, and types of deposits, there is nevertheless so much similarity between the deposits in the two districts that there is no difficulty in believing that they are part of the same metallization event.

As can be seen in Figure III-6, the ore horizons of the Metaline district are also present in the Northport district along with several others, including the Reeves Limestone Member, the unit so productive in the Salmo district. However, not all deposits identified with certain stratigraphic units in Figure III-6 are stratiform. For example, the ore bodies of the Gladstone and Electric Point mines consisted of large masses of galena in a matrix of talc and dolomite sand that filled vertical pipes—deposits strikingly discordant. Also structurally controlled are the deposits of Red Top Mountain, those of the Red Top and Lucile mines, which had rich shoots of lead-zinc ore that contained silver-bearing tetrahedrite. These occurred in a limestone that occupies the same stratigraphic position as the ore-bearing limestone at the Reeves MacDonald mine in the Salmo district. The tetrahedrite in these deposits suggests a close kinship to the tetrahedrite ores of the quartz veins in shear zones along a belt that extends northeastward for ten miles from the Columbia River to the International Boundary.

In general, the rocks of the Northport district are more intensely deformed than those of the Metaline district, but this deformation is not reflected in the ores. It is only locally, and to a minor degree, that the ore bodies are sheared or faulted. Although the ore bodies occur in folded rocks, they are not folded, but replace folded beds.

Besides occurring in deformed rocks, the ores are found in thermally metamorphosed rocks, in particular at the Deep Creek and Van Stone mines, where the pseudomorphing of tremolite by sphalerite is common, and at numerous prospects. Although the aureole of the Spirit pluton has more than its share of deposits, one would not interpret the ores as contact metamorphic deposits in the sense that the metals were derived from the pluton. On the other hand, the suggestion has been made that these aureole deposits have been metamorphosed and the ores remobilized, but this leaves unanswered the questions: Why were not the sulfides remobilized completely out of the dolomite? Why was not the sphalerite, which ranges greatly in color and iron content, homogenized? And why is there an anomalous number of deposits near the granite contact? These questions can be answered if the metallization is interpreted as occurring after the emplacement of the pluton and before its final cooling.

The relations of the ores to the geology of both the Metaline and Northport mining districts are compatible with the conclusion reached by Sinclair (1966, p. 249-262) through a study of lead-isotope abundances in lead-zinc deposits in the Kootenay Arc. Sinclairs' conclusion is that the leads have a multistage history. He postulates that

the source rocks from which the lead ores were derived could be lower Purcell (Belt) rocks into which uranium and thorium had been introduced about 1,700 m.y. ago and lead 1,340 m.y. ago. The hypothesis requires that during the remainder of the Precambrian and the Paleozoic, the lead remained in the Purcell, and that it was not until the granite-producing orogeny of the Mesozoic that it was mobilized, mixed with the radiogenic lead developed from the uranium and thorium, and moved upward.

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IV

**GEOLOGICAL SETTING OF THE LEAD-ZINC DEPOSITS IN THE
KOOTENAY LAKE AND SALMO AREAS OF BRITISH COLUMBIA^①**

By

JAMES T. FYLES

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GEOLOGICAL SETTING OF THE LEAD-ZINC DEPOSITS IN THE KOOTENAY LAKE
AND SALMO AREAS OF BRITISH COLUMBIA

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GEOLOGICAL SETTING OF THE LEAD-ZINC DEPOSITS IN THE KOOTENAY LAKE AND SALMO AREAS OF BRITISH COLUMBIA

By JAMES T. FYLES^②

INTRODUCTION

The southeastern corner of British Columbia contains the largest and the greatest number of lead-zinc deposits in the Province. Included in it is the unique Sullivan ore body, the deposits of the Kootenay Arc, which extend beyond the Province into northeastern Washington, large deposits in the Shuswap Metamorphic Complex that as yet are not mined, and others in the east Kootenays and northern Selkirks. This lead-zinc terrain of British Columbia, is part of a larger metallogenic province extending southeast into Idaho and Montana that is recognized as economically the most important lead metallogenic province of the Cordilleran Region (see Burnham, 1959). Lead-zinc deposits in carbonate rocks are the most numerous, but not the largest deposits in the lead-zinc terrain. They are mainly in the Kootenay Arc, but a few are beyond it in the Selkirk and Rocky Mountains.

The deposits considered in this field trip may be subdivided into three types: Salmo type, Metaline type, and Bluebell type. Each type includes several deposits in southeastern British Columbia and northeastern Washington and each has characteristics linking it with similar deposits elsewhere in the world.

The Salmo type consists of essentially stratiform lenticular disseminations of pyrite, sphalerite, and galena in zones of dolomite in highly deformed Lower Cambrian limestone. This type includes the Duncan near the north end of Kootenay Lake, the Jack Pot, H.B., Jersey, and Reeves MacDonald ore bodies near Salmo and the other deposits in Stevens County, Washington.

The Metaline type includes lenticular, more or less stratiform deposits in relatively undeformed Middle Cambrian carbonate rocks. The standard deposits are those near Metaline Falls, Washington. The only known significant deposit of this type in British Columbia is the Monarch-Kicking Horse deposit in the Rocky Mountains at Field.

Deposits of the Bluebell type consist of massive and disseminated sulphides in limestone adjacent to fractures. In the Kootenay Arc, the Bluebell and Lucky Jim ore bodies are the most significant examples, but small occurrences are found elsewhere.

This paper describes the geological setting of the Salmo and Bluebell deposits and outlines some of the characteristics of each type.

STRATIGRAPHY OF THE KOOTENAY ARC

The Kootenay Arc is a curving belt of complexly deformed sedimentary, volcanic, and metamorphic rocks trending northeast for 100 miles across Washington into British Columbia, north to northern Kootenay Lake and

northwest to near Revelstoke. It has a total length of at least 250 miles. In British Columbia the Arc lies between the Purcell anticlinorium on the east and gneiss of the Shuswap Metamorphic Complex on the west and contains a thick succession of sedimentary and volcanic rocks that range in age from earliest Cambrian to late Mesozoic. The succession is essentially a conformable one, though a late Paleozoic and early Mesozoic unconformity are thought to be present and probably others exist that have not yet been found. One of the most significant markers in the succession is the Badshot Formation in the Lardeau and Kootenay Lake country and its equivalent, the Reeves Member of the Laib Formation, south of Nelson near Salmo. These limestones, which contain rare Early Cambrian archaeocyathids, are repeatedly exposed by complex folding in a belt, locally as much as 10 miles wide, along the eastern side of the Arc. Rocks to the east of this belt in general are older than the limestone and pass downward into the Precambrian. Younger rocks to the west compose a thick succession extending into the Jurassic.

The terminology used in describing the Paleozoic formations in the Kootenay Arc is given in Table IV-1.

In the Lardeau and Kootenay Lake areas the rocks belong to the Hamill, Lardeau, Milford, Kaslo, and Slocan Groups. The Hamill is quartzitic; the Lardeau has a lower calcareous section containing the Badshot Formation, overlain by a thick succession of schists and quartzites with lenticular masses of volcanic rock. In the Salmo district the lower part of the comparable succession (Quartzite Range and Reno Formations) is quartzitic, and the overlying Laib has a lower calcareous part containing the Reeves Limestone Member, beneath a thick succession of schists and minor quartzites. The overlying Nelway (Metaline), which is Middle Cambrian limestone and dolomite, and the Active (Ledbetter) Formation of Ordovician dark slate and argillite occur widely, especially in Washington, but are not found far north of the latitude of Ymir. The distribution of the Badshot, Reeves, Nelway, and Metaline Limestones, which are the most important lead-zinc-bearing formations, is shown on Figure IV-1.

The Mesozoic formations lie along the western side and within the curvature of the Arc. The Rossland Volcanic Group and the Slocan argillite, slate, and limestone are important formations in this assemblage and contain significant mineral deposits, principally in the Rossland copper-gold camp and the Slocan silver-lead-zinc camp. Granitic plutons including several small batholiths, many stocks, and sill-like masses with a variety of metamorphic

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TABLE IV-1.—Terminology and age of Paleozoic formations in the Kootenay Arc.

AGE		NORTHERN WASHINGTON		SALMO DISTRICT		KOOTENAY LAKE-LARDEAU			
			FOSSILS			FOSSILS	FOSSILS		
Mississippian -Permian				Unnamed formations		Milford Group (Mississippian)	Corals Crinoids		
	Silurian and Devonian	Limestone and argillite	Reefs Conodonts			Unconformity			
Ordovician	Ledbetter Slate	Graptolites	Active Formation	Graptolites		Undivided formations			
Cambrian	Middle	Metaline Formation	Trilobites	Nelway Formation			Lardeau Group		
	?	Maitlen Phyllite		Upper Laib					
		Reeves Member	Archaeocya- thids	Emerald Member					
				Reeves Member	Archaeocya- thids			Badshot Formation	Archaeocya- thids
				Truman Member				Mohican Formation	
	Gypsy Quartzite		Reno Formation			Hamill Group			
			Quartzite Range Formation						

and structural characteristics interrupt the continuity of the older deformed stratigraphic succession.

The Reeves Member and equivalent Badshot Formation contain the Duncan and Bluebell ore bodies, lead-zinc deposits in the Salmo district. The limestone is normally a few hundred feet thick but ranges from a few tens of feet to more than 1,000 feet, depending mainly on its structural setting. It is a grey, fine- to medium-grained crystalline limestone with some significant lithological variations related to its original character and subsequent deformation. It is remarkably uniform over a distance of 150 to 200 miles from north to south and varies gradually and systematically in lithology from east to west.

In the Salmo area the Reeves Limestone in the western or Mine Belt contains all the lead-zinc ore bodies. It is a banded grey and white or black and white fine- to medium-grained rock that weathers blue-grey. It contains

zones of fairly pure dolomite that are finer grained than the limestone and either massive grey or creamy white coloured or are mottled with black flecks, wisps, and bands. This textured dolomite contains the ore bodies, whereas the massive dolomite is only sparsely mineralized. Near the Reeves MacDonald mine the Reeves Limestone contains a lens of dark-grey siliceous dolomite.

By contrast, the Reeves Member on the Sheep Creek anticline about 5 miles east of the Mine Belt is fairly massive, fine grained, and contains thin, argillaceous lenses. Another 4 miles to the east in the Eastern Belt (see Fyles and Hewlett, 1959, p. 16) the Reeves Member ranges from a grey-weathering, white, finely crystalline limestone to a buff-weathering, grey, finely crystalline argillaceous limestone and calcareous siltstone. The most calcareous sections contain abundant archaeocyathids. On the Sheep Creek anticline and the Eastern Belt, dolomite is confined to altered zones adjacent to veins and intrusive rocks.

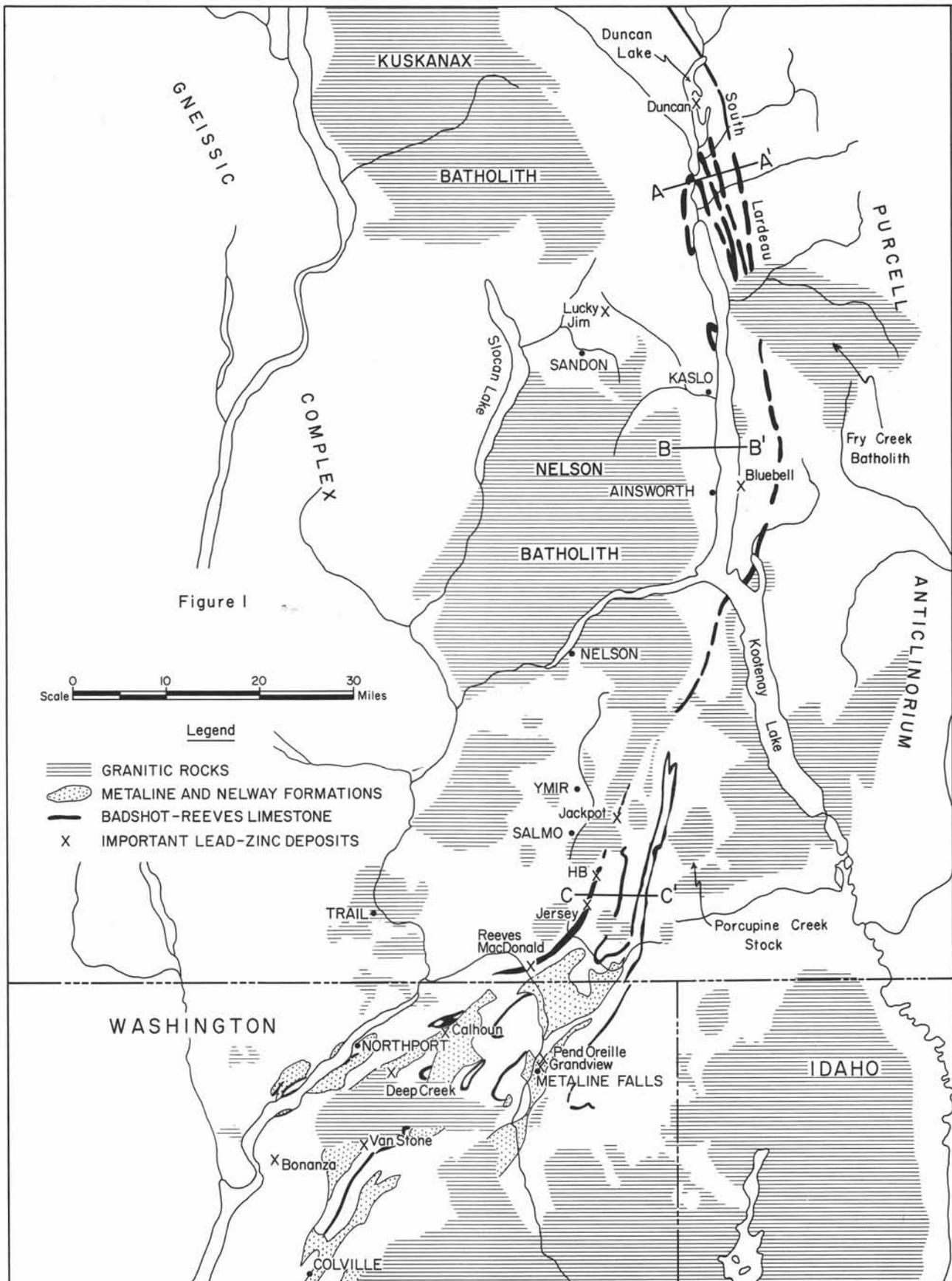
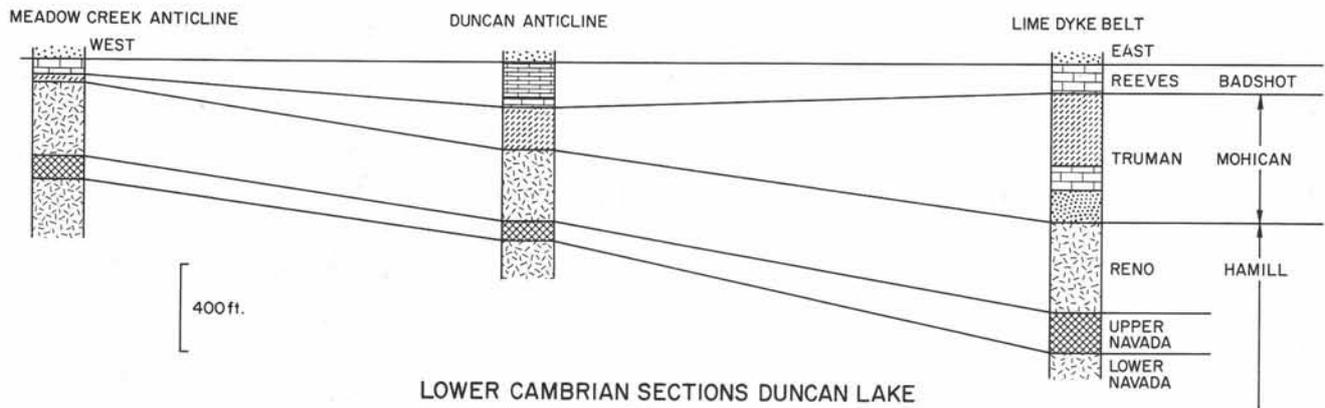
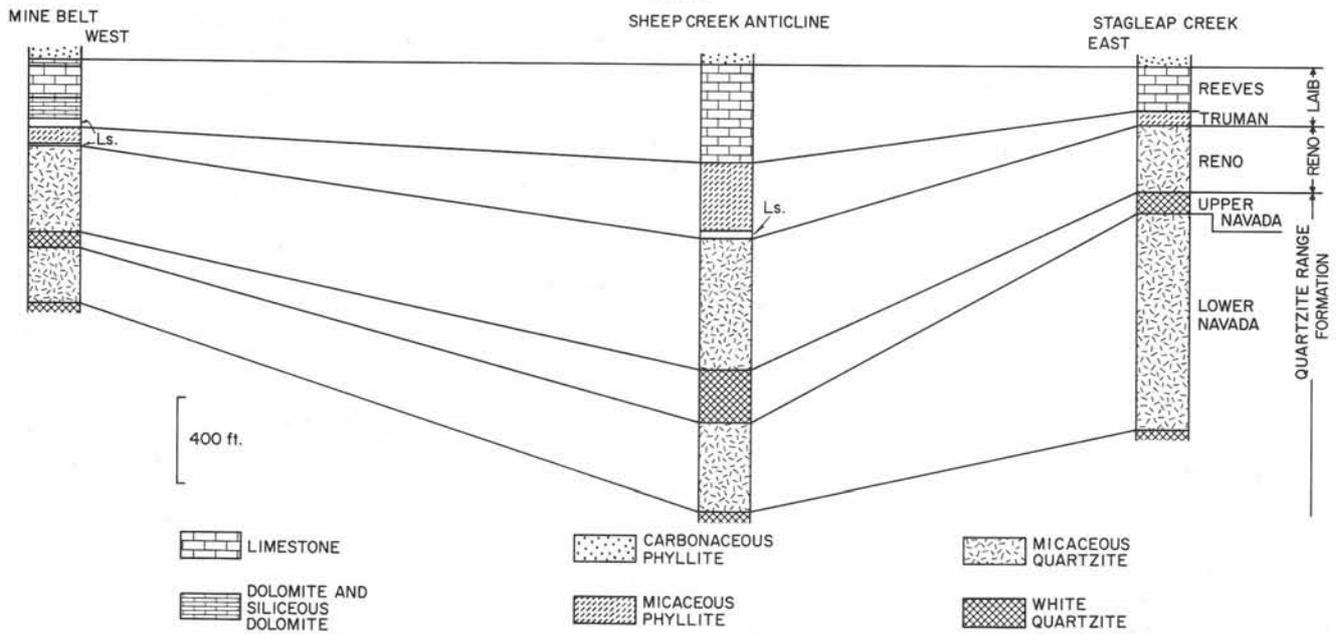


FIGURE IV-1.—Geological map of the southern part of the Kootenay Arc.



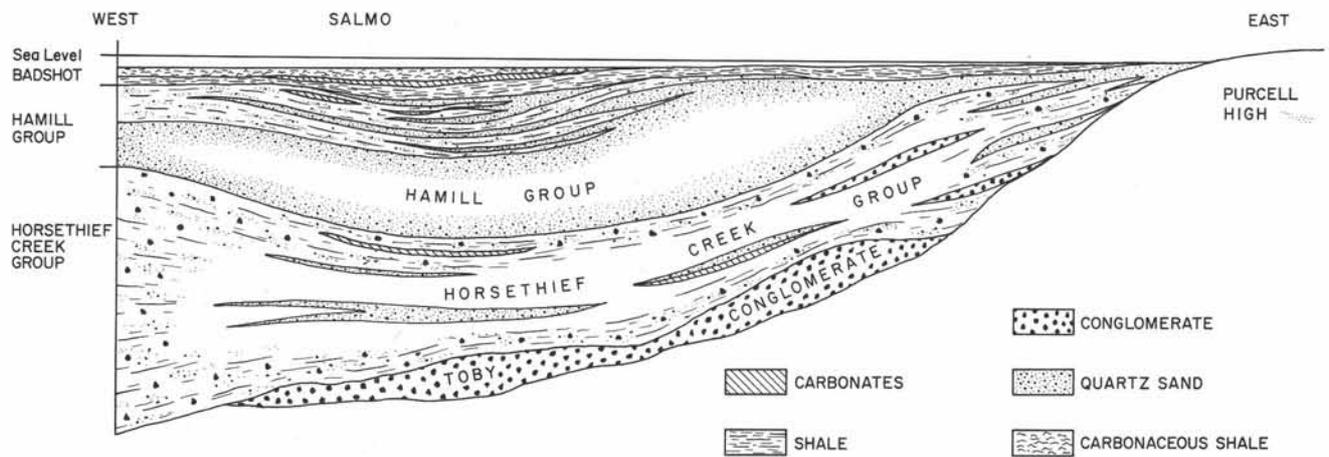
LOWER CAMBRIAN SECTIONS DUNCAN LAKE

FIGURE 2A



LOWER CAMBRIAN SECTIONS SOUTHERN PART OF THE SALMO AREA

FIGURE 2B



RESTORED SECTION LATE LOWER CAMBRIAN

FIGURE 2C

FIGURE IV-2.—Stratigraphic sections and reconstructed section of Lower Cambrian rocks in the Kootenay Arc.

These three contrasting facies of the Reeves Member extend some 20 miles northward from the International Boundary as three belts of outcrop exposed by three complex folds. The same sort of contrasting facies on comparable structures are found in the area 100 miles to the north in the south Lardeau near the north end of Kootenay Lake. In that area, the lead-zinc deposits are on the Duncan anticline in the Badshot Formation, which contains lenses of dolomite and siliceous dolomite. West of the Duncan anticline the Badshot is a relatively thin, clean crystalline limestone and to the east it becomes thicker and contains layers of quartzose calcareous siltstone.

The Badshot-Reeves Limestone is underlain by interbedded limestone, calcareous phyllite, and minor quartzite composing the Mohican Formation in the Duncan Lake area and the Truman Member in the Salmo area (Table IV-1). Calcareous quartzite grades downward into a thick succession of micaceous and white quartzite, making up the Reno and Quartzite Range Formations and Upper Hamill Group, which lie beneath the Truman and Mohican, respectively. Several members of this succession beneath the Badshot-Reeves Limestone show contrasting facies that parallel the facies changes in the Badshot-Reeves Limestone.

Rocks in the interval between the top of the Badshot-Reeves Limestone and the base of the Quartzite Range Formation and equivalent part of the Hamill Group originally formed a thick wedge of well-washed clastic sediments and clean carbonates with remarkably uniform characteristics from north to south and gradual systematic facies changes from east to west. Cross-bedding and grain-size distributions in the quartzitic rocks beneath the Badshot-Reeves Limestone suggest that a shoreline lay to the east during deposition of the sediments. No lower Paleozoic rocks are exposed west of the Kootenay Arc, so that well-founded conclusions regarding the western termination of this group of rocks cannot be made. The Badshot-Reeves Limestone marks a change from dominantly noncarbonaceous, nonvolcanic rocks with only minor carbonates below, to carbonaceous and calcareous sediments and increasing amounts of volcanic rock above. Complexities of the structure do not permit a realistic restoration of the Paleogeography. Stratigraphic sections using average thicknesses and a reconstructed stratigraphic section with an arbitrary horizontal scale are shown on Figure IV-2.

STRUCTURE OF THE SOUTHERN PART OF THE KOOTENAY ARC

FOLDS

The structure of the rocks of the Kootenay Arc in British Columbia is dominated by complex multiple folds. The most important belong to two phases of deformation referred to as Phase I, the oldest, and Phase II. Other folds belonging to later phases of folding are known, but further study is required to determine their regional significance. Phase I and Phase II folds have essentially parallel axes with a low plunge. South of the west arm

of Kootenay Lake the plunge is to the south, and along northern Kootenay Lake and Duncan Lake it is to the north. The low uniform plunge permits the regional structure to be accurately depicted in vertical cross-sections (Fig. IV-3).

The relation between the two phases of folding are most clearly displayed in the Duncan Lake area, and the same pattern of folds extends southward along Kootenay Lake and into the Salmo area.

Around Duncan Lake and the north end of Kootenay Lake, Phase I folds are isoclinal and extremely attenuated. In general, convergence of the limbs cannot be measured in the field and the fold hinges cannot be seen. The folds are recognized from a knowledge of stratigraphic relations and map patterns of rock units. These folds plunge to the north at about 10°. Their limbs and axial planes are curved and have been folded on Phase II structures.

The axes of Phase II folds plunge at about the same angle to the north as the Phase I folds. Phase II folds range in shape from very tight to relatively open, and in size from a few feet across to folds with axial planes several miles apart. They can be seen in the field and in general are outlined by the layering and by the attitudes of formational contacts. Ideally, complete Phase II folds are composed of an anticline lying east of a complementary syncline, giving the form of a reversed letter N.

In Figure IV-3, section A-A' is a diagrammatic composite section showing the structure at the north end of Kootenay Lake. Both Phase I and Phase II folds are shown, and these folds have been traced throughout the Duncan Lake area, a distance along the strike of about 25 miles, in which distance, because of the uniform plunge, a structural depth of more than 3 miles is implied.

South of the north end of Kootenay Lake, folds near the hinge zone or beneath the Meadow Creek anticline are found along the west side of the lake. Section B-B' in Figure IV-3 is a diagrammatic section, extended hypothetically across the lake to include the Badshot and Hamill rocks exposed north of the Bluebell mine.

South of the west arm of Kootenay Lake the regional west-dipping foliation gradually steepens. North of the latitude of Ymir it is essentially vertical, and from there southward it dips steeply to the east. All this region is complicated by faults and granitic intrusions, and in the Salmo area by an abrupt swing in the regional strike. Section C-C' in Figure IV-3 is across a part of the Salmo area in which the regional strike is north and the plunge of the folds is low, dominantly to the south. On the eastern edge of the section, the Reeves Limestone is repeated on the limbs of an isoclinal syncline, the Laib syncline, west of which is a complementary anticline known as the Sheep Creek anticline. These are Phase I folds, and are typical of structures in the Eastern Belt of the Salmo district. They are followed to the west by a complex deep syncline, called the Black Argillite Belt, containing incompetent rocks of the Active Formation. West of the Black Argillite Belt, rocks lying between the uppermost members of the Quartzite Range Formation and the top of the Laib Formation are complexly folded. They are in the Mine Belt and in broadest structural

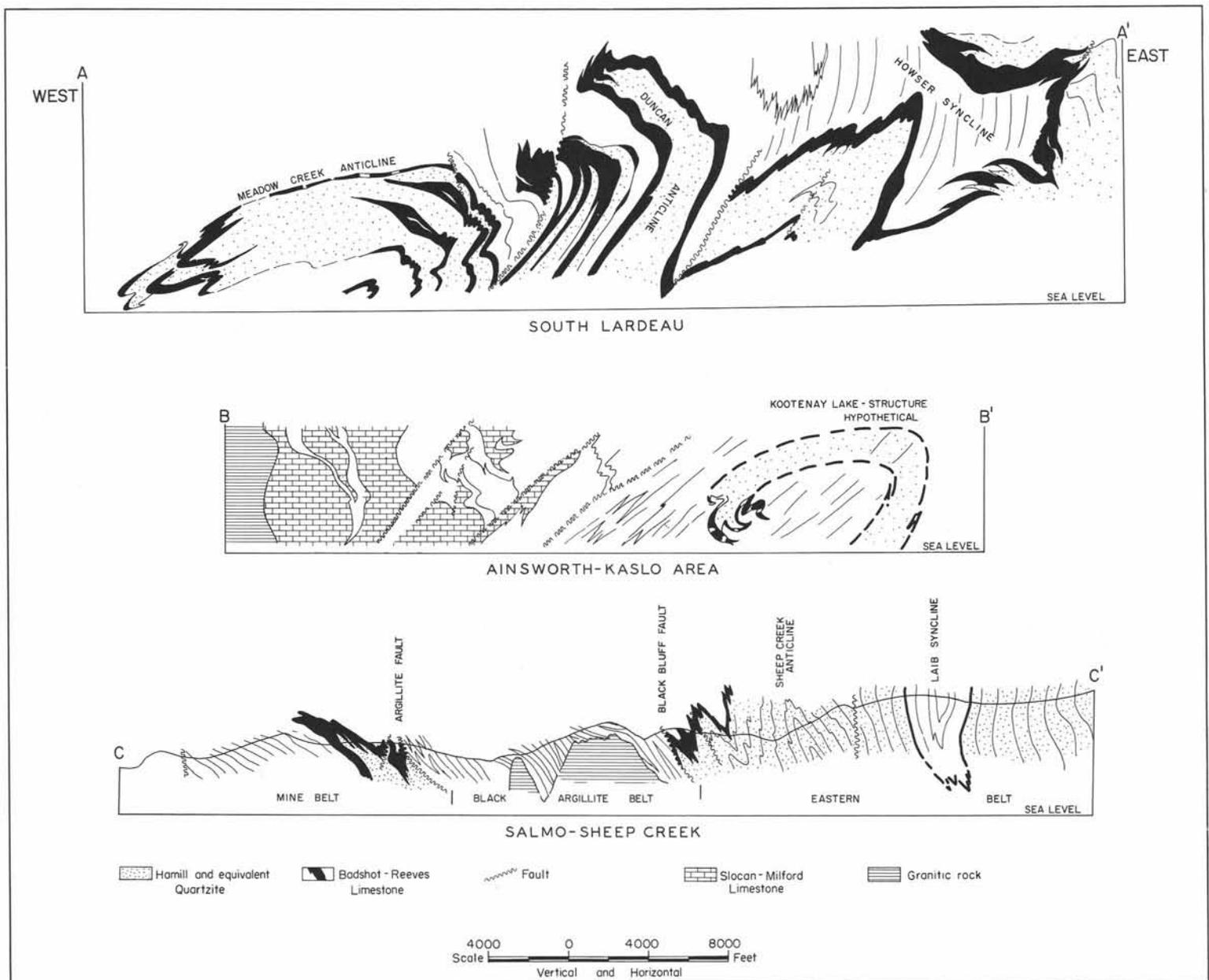


FIGURE IV-3.—Diagrammatic structural sections.

form are anticlinal. The Mine Belt and the synclinal Black Argillite Belt are a pair of Phase I folds, comparable to the Duncan anticline and Howser syncline of the Duncan Lake area.

These structures are truncated toward the southwest by faults, but the Sheep Creek anticline continues southward across the International Boundary. Ten miles north of the boundary the anticline is almost isoclinal; toward the south it becomes more open and less complicated by parasitic folds. South of the International Boundary the anticline is recognized in the Metaline Formation as a broad southwestward-trending structure broken by numerous faults. This change in apparent style of the Sheep Creek anticline as it crosses the International Boundary into structurally higher levels in the valley block of the Metaline quadrangle results in the contrast between the style of deformation of the rocks containing the Metaline lead-zinc deposits and those containing the Salmo deposits.

In the Salmo area, Phase II folds are most obvious in the Mine Belt, and Phase I folds, which are highly appressed, are relatively obscure. Phase II folds are defined by the attitudes of foliation and range from tight to relatively open. Probably the best-known Phase II folds are in the Jersey mine, where the axial planes dip steeply to the east and the axes plunge southward with the plunge of the Phase I folds.

In the southern part of the Salmo area the fold axes swing to the west and the plunge steepens. Strike faults that separate the Eastern, Black Argillite, and Mine Belts, and which in general dip to the east, also swing westward in strike and flatten in dip. In addition, north- and northwestward-trending block faults become more numerous.

Phase I and Phase II structures are more or less warped, sheared, broken by faults, or intruded by igneous and plutonic masses.

The Nelson and Kuskanax batholiths and many of the granitic stocks have local zones of intense deformation around their margins. On the northern edge of the Nelson batholith older structures are buckled downward within a mile or so of exposed granitic rocks. On the eastern edge of the batholith north of Ainsworth, Phase I and Phase II folds are warped sharply upward within a half mile of the granitic rocks. The Hidden Creek stock in the Salmo district is surrounded by a zone in which the regional strike is deflected into near parallelism with the margins of the stock. A few miles to the north the Porcupine Creek stock is ringed by upturned sedimentary rocks. It is possible that warps preceded and controlled the emplacement of the granitic masses, and that forceful intrusion further deformed the wall rocks and produced local marginal zones of faulting.

FAULTS

Phase I and Phase II folds have strike and thrust faults associated with them. A cluster of northward-trending faults along the west side of central Kootenay Lake dips westward beneath the Nelson batholith. In the Salmo district, strike faults separate the structural belts. The Black Bluff fault is east of the Black Argillite Belt, the Argillite fault is east of the Mine Belt, and the Waneta fault is east of the Mesozoic volcanics. In the northern part of the area they dip steeply to the east and to the south, and they flatten in dip and swing westward in strike with the curvature of the Arc. West of the Reeves MacDonald the Argillite fault dips to the south at 15° to 20° and resembles a thrust fault. These faults are complex structures that probably originated during Phase I folding and acted as zones of movement during all subsequent deformation. They probably acted as channelways for the migration of mineralizing fluids.

Block faults with significant displacement are uncommon north of the Salmo district, but to the south and southwest they are abundant and tend to obscure the older structural patterns.

PLUTONIC AND IGNEOUS ROCKS

Irregular, large and small masses of granitic rock occur throughout the Kootenay Arc. These include the Nelson, Kuskanax, and Fry Creek batholiths near central and northern Kootenay Lake and many smaller stocks and sill-like masses to the south. They are predominantly quartz monzonite, although the composition ranges widely. Porphyritic varieties with large phenocrysts of potash feldspar are common. It has been recently determined by potassium-argon methods that the northern part of the Nelson batholith was emplaced 160 million years ago (Nguyen, Sinclair, and Libby, 1968). The rest of the Nelson plutonic rocks, which have not been dated, are generally considered to be Late Jurassic or Early Cretaceous.

The Jersey and Emerald stocks exposed in the Jersey mine are part of the Nelson plutonic suite. Sills of felsite, fine-grained leucogranite, quartz feldspar porphyry, and granite pegmatite occur widely, and locally are abundant. In the Salmo area and along Kootenay Lake they follow

the most prominent foliation in the deformed rocks and appear to have been folded. Along Kootenay Lake they are closely associated with the Nelson plutonic rocks, and where the geological relationships are clear they are older than, or the same age as, the associated plutons.

Small Tertiary intrusions are common in the southern part of the Kootenay Arc and west of it. These include stocks of augite biotite monzonite common in the Salmo area and irregular masses of the Sheppard granite near the International Boundary.

All the rocks are cut by lamprophyre dykes which follow fractures, faults, or prominent foliation planes and range from small discontinuous masses to bodies several tens of feet wide and several thousands of feet long. They are dark greenish-grey or brownish rocks commonly with subhedral phenocrysts of biotite, feldspar, hornblende, or augite. These dykes are undeformed, cut the plutonic rocks, and probably are mainly Tertiary. Radiometric dates have not been reported for these dykes.

METAMORPHISM

The Kootenay Arc contains rocks ranging from a low to a high grade of regional metamorphism. Although mapping of metamorphic zones is still incomplete it has been possible in areas of detailed work to map metamorphic isograds based on the first appearance with increasing metamorphism of the classic index minerals. An elongate metamorphic belt centred on Kootenay Lake extends along the Purcell trench for 15 to 100 miles. Rocks in the biotite and higher grades form a belt probably about 15 miles wide from east to west and at least 75 miles long. Central Kootenay Lake and the Bluebell mine are within the area enclosed by the kyanite isograd and in which the rocks belong to the lower amphibolite facies of regional metamorphism. The garnet isograd follows the western side of Kootenay Lake and passes close to the Duncan lead-zinc deposit. Most of the Salmo area is between the garnet and biotite isograds in rocks of the greenschist facies.

Zones of regional metamorphism are truncated by the plutons, each of which has a marginal zone of contact metamorphism. At the Jersey mine near the Jersey and Emerald stocks, for example, calcareous rocks are metamorphosed to medium- to coarse-grained garnet-diopside skarn. Siltstones and argillites are thermally metamorphosed to various types of biotite hornfels. At the Jack Pot property on the northern edge of the Hidden Creek stock, coarsely crystalline limestone contains clusters of tremolite, wollastonite, and forsterite. These zones of thermal metamorphism extend a few hundred feet to as much as half a mile from granitic contacts which commonly are sharply defined. A few plutons are associated with wide zones of gneiss and granitized metasedimentary rocks.

LEAD-ZINC DEPOSITS

Detailed descriptions of many of the lead-zinc deposits in the Kootenay Arc have been given in various publications, and up-to-date accounts by geologists working

with the deposits discussed on this field trip are included in this volume. The following notes summarize salient features of selected deposits emphasizing similar and contrasting characteristics, many of which are important in understanding the origin of the deposits.

The Salmo and Duncan deposits are disseminations of pyrite, sphalerite, and galena with or without pyrrhotite in a calcareous host rock. The dominant gangue is the host rock or its alteration products and consists mineralogically of dolomite, calcite, quartz, and rarely barite. Detailed studies indicate that the quartz and dolomite formed by silicification and dolomitization of limestones before sulphide mineralization and at least in part before deformation. Much of the calcite appears to have developed from the dolomite at the time the sulphides were deposited. Certain dolomitized and silicified zones were selectively replaced by the sulphides. Silicification and dolomitization were much more widespread than sulphide mineralization, and their ultimate control is not known. Sulphide mineralization, however, is localized by fold structures.

CONTROL OF MINERALIZATION

The Salmo and Duncan ore bodies are controlled by Phase II structures, mainly folds, but locally by faults and breccia zones associated with them. At the Jersey mine these folds are upright and fairly open with axes plunging at low angles to the south (Bradley, Figs. VIII-2 and VIII-3, this guidebook). Ore in "A" zone follows the crest of an antiform for 4,000 feet. The "B" zone follows the trough of a synform for almost 1,000 feet, and the "C" and "D" zones are along the flanks of an upright antiform with continuous mineralization for at least 1,500 feet.

At the Reeves MacDonald, Phase II folds plunge 50° to the south and have modified isoclinal Phase I folds, which have a more gentle plunge to the southwest. The Reeves ore body and the faulted extensions to the east have the form of, and lie within, an isoclinal Phase I syncline steepened in plunge by Phase II structures. They have a total length parallel to the plunge of the Phase II folds of at least 4,000 feet.

The Duncan zones of mineralization, which have not been mined, are associated with Phase II folds plunging about 10° to the north. The longest zone outlined by drilling and underground work is more than 3,000 feet long parallel to the plunge of the folds.

Although in hand specimens and under the microscope some sulphides appear deformed, the ore bodies are not folded by Phase II structures. On macroscopic scales mineralization has feathery terminations indicative of incomplete replacement of folded layers.

Some mineralization is controlled by zones of brecciated dolomite. Part of the Reeves ore body consists of sulphides, mainly pyrite, enclosing rounded and irregular disoriented fragments of dolomite. A layer rich in galena in the "A" zone of the Jersey mine consists of angular fragments of dolomite surrounded by sulphides. In the H.B. mine, gently dipping tabular, relatively high-grade zones called cross-zones, because they lie between more extensive disseminated ore bodies, are composed of breccia. The textured dolomite itself is a type of breccia.

In it dark-grey, carbonaceous wisps surround angular and irregular masses of light-grey dolomite, producing a lineation parallel to the axes of Phase II folds. The distribution and attitude of the breccia mineralization and of the textured dolomite indicate their close association with the Phase II deformation.

The Bluebell ore bodies by contrast are more or less massive replacements of limestone controlled by cross-fractures. The limestone is 100 to 150 feet thick, strikes northward, and dips, on the average, 35° to the west. The limestone belongs to the Badshot Formation. Very similar deposits occur in stratigraphically higher limestones on the western side of Kootenay Lake in the Ainsworth camp. The ore bodies are generally of higher grade, richer in silver, and more complex in mineralogy than those of the Salmo type. They have formed by replacement outward from a joint system that is subsequent to both Phase II deformation and to regional metamorphism.

AGE OF MINERALIZATION

Studies of lead isotopes by Sinclair (1966, p. 251) indicate that the lead of the Kootenay Arc deposits had a relatively complicated history before being emplaced in the present mineralized zones. The age relations determined geologically concern the latest mineralization process, which took place well after most of the host rocks were deposited. The criteria for relative age are commonly difficult to interpret and may lead to conflicting conclusions. They show various stages in a complicated mineralization process. Gross relations lead to the following conclusions:

The Salmo and Duncan deposits are relatively old, and the Bluebell is relatively young.

(a) Salmo deposits are older than a suite of lamprophyre dykes that are undeformed, of very similar composition and texture, and have been regarded as a con-sanguineous suite of Tertiary age. Absolute age determinations that might establish this age and relative ages within the suite have not yet been made. The Bluebell mineralization is later than these dykes.

(b) At least one ore body at the Jersey mine is cut by a granitic dyke, which is an offshoot of a granitic stock, the Jersey stock, belonging to the Nelson plutonic suite.

(c) At the H.B. mine some galena and sphalerite have been found along cleavage planes of tremolite that formed by thermal metamorphism near granitic stocks close to the Jersey stock. Thus the galena is younger than the thermal metamorphism, indicating that some of the galena at the H.B. was deposited during granitic intrusion.

(d) On a regional basis, ore bodies of the Salmo type are associated with structures that are older than the Nelson plutonic rocks.

(e) Ore textures studied and illustrated by Muraro (1966, p. 245) clearly show that the sulphides of the Salmo deposits have undergone penetrative deformation.

The Nelson granitic rocks, which include the Nelson and Kuskanax batholiths and a number of related stocks mainly in the southern part of the Arc, as now exposed, seem to have been incidental to the mineralization process. Early workers regarded these granitic rocks as the

source of the lead and zinc, but detailed studies have not found evidence that any one deposit or group of deposits is genetically related to a specific pluton. The Salmo deposits appear to be older than the exposed granitic rocks.

CONCLUSION

In the foregoing, essential features of the important lead-zinc deposits in the Kootenay Arc in British Columbia have been summarized. From the observed characteristics it is clear that the deposits have had a complex history, and evidence for their origin becomes more obscure the farther back through this history we try to go. Relatively young fracture-controlled replacement such as the Bluebell and associated veins clearly formed by deposition of minerals from hot water saline solutions channelled by fractures in brittle rocks which had previously undergone plastic deformation and metamorphism. Mafic dykes in the Bluebell and Ainsworth areas that follow the same patterns of fracturing as the veins show that this mineralizing process is closely associated with magmatic activity.

On the other hand, the older concordant Salmo type deposits have shapes controlled by pre-Nelson fold structures. The sulphides have been deformed, they appear to replace folds and special types of deformed dolomite, and locally they are intergrown with silicate gangue minerals formed by thermal metamorphism associated with Nelson intrusions. From these relations it is concluded that during deformation there was widespread movement of the metals in hydrothermal solutions and that during granitic intrusion there was remobilization in zones of contact metamorphism. The widespread mineralization does not bear any spatial relation to the Kootenay Lake zone of high-grade regional metamorphism, and from this it is concluded that characteristics of the enclosing sedimentary rocks such as their original composition and water content were dominant in influencing the mineralizing process. Although our knowledge of the original stratigraphy is very incomplete, the association of the Salmo type deposits with dolomitic and siliceous parts of the clean facies of the Badshot-Reeves limestone and with an "offshore" facies in the underlying sequence of quartzitic rocks suggests that ancient sedimentary processes were important in controlling the regional distribution of the

metals. Finally, lead isotope studies by Sinclair suggest that the lead of the Kootenay Arc deposits went through several stages of introduction and redistribution, beginning with the introduction of uranium and thorium in the upper crust 1,700 million years ago.

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V

STRUCTURAL ENVIRONMENT OF THE SALMO TYPE LEAD-ZINC DEPOSITS

By

A. S. MACDONALD

STRUCTURAL ENVIRONMENT OF THE SALMO TYPE LEAD-ZINC DEPOSITS

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STRUCTURAL ENVIRONMENT OF THE SALMO TYPE LEAD-ZINC DEPOSITS

By A. S. MACDONALD^①

INTRODUCTION

This contribution is a progress report on a current research project and not a definitive account of the structural environment of Salmo type lead-zinc deposits. Specifically, it is an analysis of minor structures, mapped mainly at the surface in the immediate vicinity of the Reeves MacDonald and Jersey mines (Fig. V-1), pre-

homogeneous domains. However, sampling bias is inevitable due to exposure variability and perhaps also due to the variable orientation of structures relative to the surface.

STRUCTURAL ANALYSIS

PHASE I STRUCTURES

The foliation consists of bedding (S_0) more or less modified by a nearly coincident cleavage (S_1) (Figs. V-2g and V-3e,f).

Phase I folds are recognized, on both microscopic and minor scales, as complex isoclinal structures of similar style (Figs. V-2h,i,j and V-3g,h,i). Where undeformed, they are overturned toward the west and plunge south to southwest (Figs. V-2b and V-3b). Locally abundant near the hinge zones of the major folds (Salmo River and Jersey anticlines), they are sporadic in occurrence elsewhere, apparently having been obliterated by the transposition of bedding on the limbs of these folds, by subsequent deformation, and by local recrystallization accompanying contact metamorphism in the Jersey mine area.

A relatively obscure lineation (L_1), paralleling the minor fold axes, can be discerned locally in the non-carbonate rocks. It is marked by bedding/cleavage intersections, mica alignment and faint disjoint crinkling, often difficult to distinguish from L_2 lineation unless in association with it.

The linear structures (F_1 fold axes and L_1 lineations) plunge at various angles either toward the southwest or the northeast, reflecting reorientation during subsequent folding (Figs. V-2b and V-3b). However, the strong maximum concentrations of these elements in the stereoplots indicate a high directional stability, which may be interpreted as the original orientation. Note also that the fold axial surfaces have the same general orientation as the main foliation, confirming their genetic association with the S_1 penetrative cleavage (Fig. V-3a and b).

This cleavage, since it is weakly developed in the fold closures, appears to have been enhanced by the later widespread development of transposed bedding, together with other extension features such as boudinage and bedding faults (Figs. V-2f and V-3g).

PHASE II STRUCTURES

Phase II folds and lineations are ubiquitous and comprise the most obvious minor deformation structures in both areas.

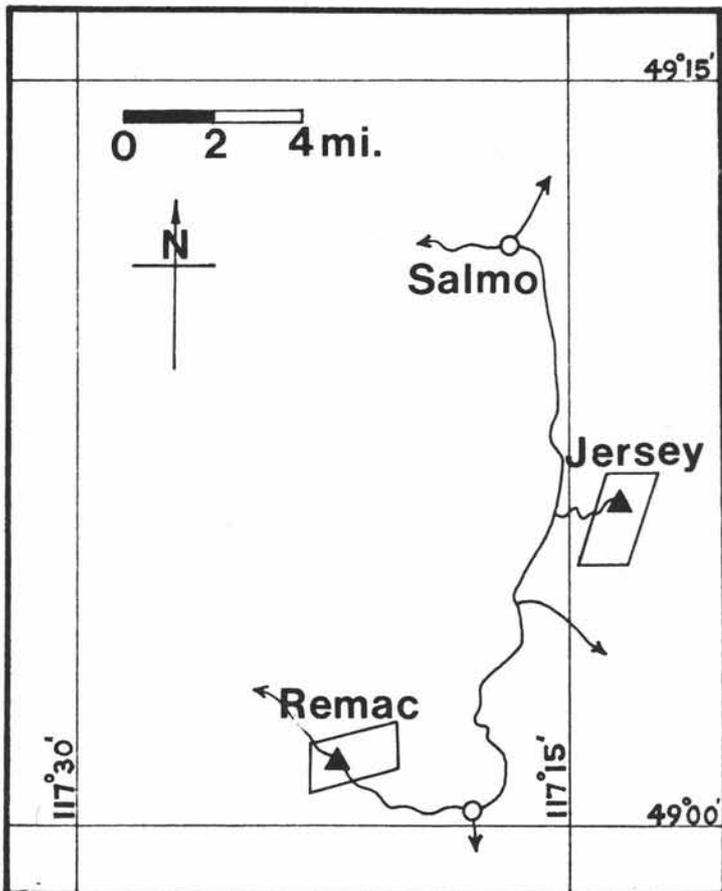


FIGURE V-1.—Location of areas investigated.

sented with some conclusions regarding the significance of these structures in the structural evolution of the host rocks and the associated sulphide deposits.

The attitudes of S-planes, minor folds, and lineations were mapped and plotted on an equal area net and contoured using the Mellis method. Different fold generations were distinguished in the field on the basis of their style and orientation. Difficulties in distinguishing between generations were encountered, particularly with lineations, because of the near parallelism of Phase I and Phase II axes of folding.

The relatively small areas investigated are considered, for the purposes of analysis, as representing statistically

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REEVES MACDONALD MINE AREA

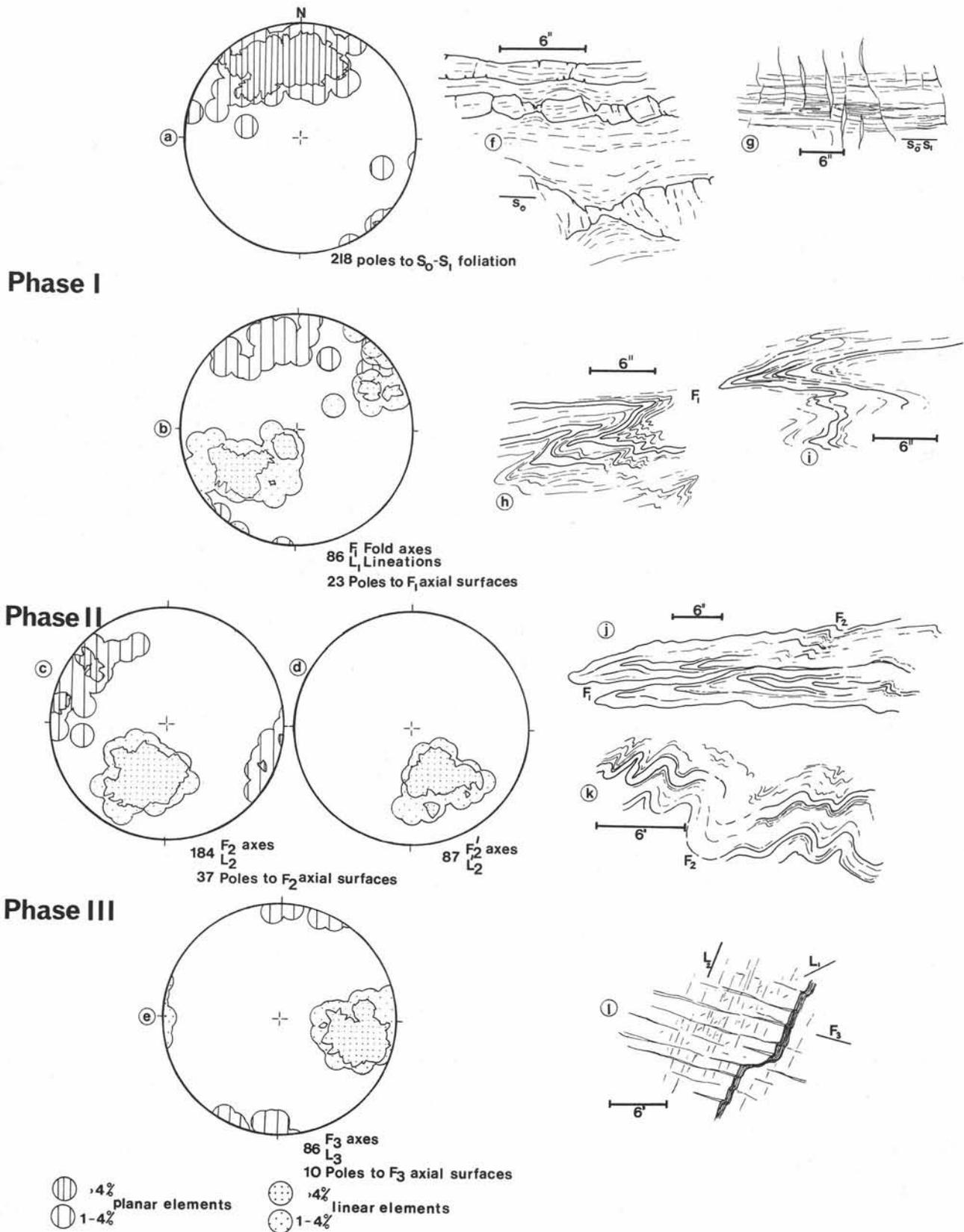


FIGURE V-2.—Contour diagrams and sketches of structural features in the Reeves MacDonal mine area.

JERSEY MINE AREA

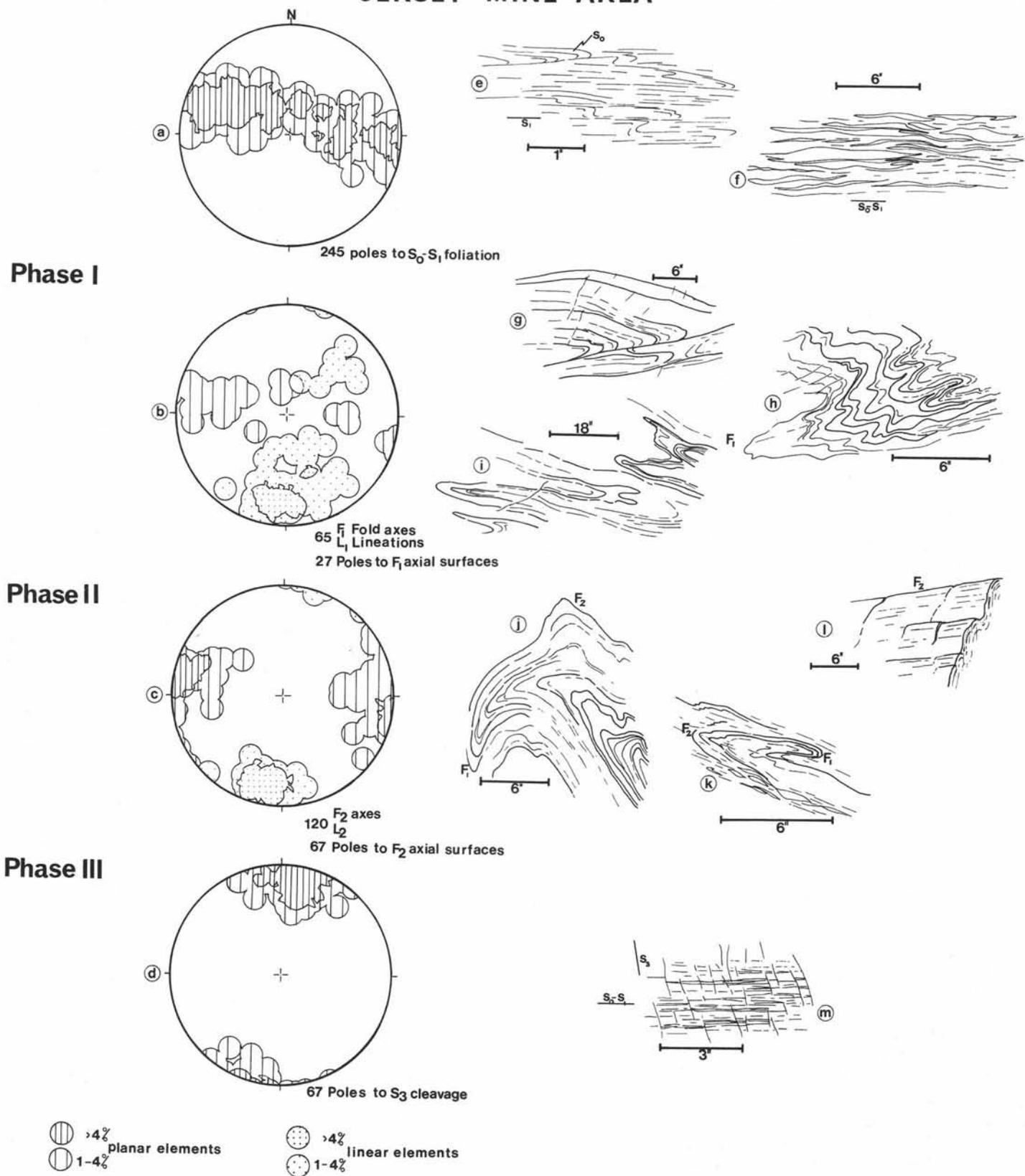


FIGURE V-3.—Contour diagrams and sketches of structural features in the Jersey mine area.

The folds are moderately tight to open, asymmetric, sinistral folds (i.e., Z-shaped in plan) (Figs. V-2k and V-3i). They are upright and have steeply dipping axial surfaces. Thus, at Jersey, they tend to be readily distinguished from the more recumbent Phase I folds. Unlike Phase I folds, lithology markedly affects the style, so that competent siliceous units display more rounded hinges (Fig. V-2k). Striations are also developed across some fold crests perpendicular to the axes, indicating flexural-slip folding. In addition, measured perpendicular to the beds, there is thickening of the limbs toward the crests, and measured parallel to the axial surfaces, thicknesses of the beds tend to a minimum at the crests, further indicating that the flexural-slip folding has been modified by flattening (Ramsay, 1967, p. 466).

A very strong axial lineation (L_2) is developed in all lithologies and is marked by: (1) the linear alignment of micas; (2) crinkling of the S_0 - S_1 foliation; and (3) a distinctive rodding of the quartzitic rocks.

The associated S_2 planes likewise vary with lithology from a fine crenulation cleavage in pelitic rocks to a discrete axial plane cleavage in the more massive or competent rocks. The former is apparent in thin section but is difficult to map except as a linear expression (L_2) on the main foliation planes. The fold axial surfaces and associated cleavage are subvertical.

Interference fold structures, produced by the refolding of Phase I folds about Phase II axes, are not uncommon in either area, but the style differs between the two areas because of the different angular relationships between F_1/F_2 orientations (Figs. V-2j and V-3j,k).

Phase II structures have highly consistent orientations within the limits of the areas mapped. However, the plunge of linear elements changes north and south of the Jersey mine area, indicating the existence of a culmination which is presumed to be the effect of broad cross-folding (Phase III?).

At the Reeves MacDonald mine, Phase II folding is complicated by a second episode (F'_2) of open, flexural-slip folding with an associated axial lineation (L'_2). These plunge steeply south to southeast, making distinction from earlier F_2 and L_2 elements difficult except where both structures occur together (Fig. V-2c and d).

PHASE III STRUCTURES

Phase III structures differ in form between the Reeves and Jersey but display the same orientation (Figs. V-2e and V-3d). At the Reeves they take the form of faint lineations, close-spaced kink-bands and occasional open dextral folds (i.e., N-shaped in plan) consistently plunging to the west-southwest (Fig. V-2l). A weak strain-slip cleavage can be distinguished locally. No major Phase III structure was recognized within the map area.

In comparison, at the Jersey mine a closely spaced east-west fracture cleavage affects most competent units (Fig. V-3m). There is no obvious association with minor folding as at the Reeves, but there appears to be a geometric relation with the broad cross-folds, interpreted as affecting the plunge of Phase II linear elements. The fractures are frequently dilated and filled with quartz

or calcite. They cut contact metamorphosed spotted schists and skarn and are seen to cut grossularite bands produced in the highest grade of metamorphism (hornblende-hornfels facies) affecting the Truman member. It is not clear to what extent these fractures predate intrusion of the Dodger and Emerald stocks.

Calcite-filled tension gashes, cutting the Reeves Limestone, are common at the Reeves mine, particularly underground where they are frequently associated with coarse, slightly darker sphalerite and galena. They occur in linear swarms or as short parallel gashes, usually cutting across the foliation and dipping north to northeast at 20° to 50° . Their relation to earlier S-planes is not clear, but in part they may represent dilated S-planes together with a variety of joint planes.

STRUCTURAL SYNTHESIS

Phase I and II folds are approximately homoaxial. The two phases may in fact represent essentially continuous folding, with the change in style from similar to flexural-slip due to changing orientation of the structure relative to the stress field and/or to waning stress, so that the ductility contrast between the different lithologies came into play.

The effect of Phase II has been to refold the major Phase I structures and to produce local steepening of the Phase I axes and a general tightening of the folds accompanied by the development of, or at least the enhancement of, transposed bedding, bedding faults, and possibly sliding of major lithological units.

The effects appear to decrease southward along the Kootenay Arc, this being reflected in the reduced dispersion of S_0 - S_1 poles in Figure V-2a as compared with Figure V-3a.

Both sets of folds change direction southwestward roughly paralleling the curvature of the Arc, but a divergence of the folds also develops, the Phase I folds being oriented more to the southwest.

Modal orientations for the various structural elements, determined from the stereoplots, are summarized below:

	Phase I		Phase II		Phase III	
	F_1 Dirn./ Plunge	S_1 Strike/ Dip	F_2 Dirn./ Plunge	S_2 Strike/ Dip	F_3 Dirn./ Plunge	S_3 Strike/ Dip
JERSEY	$190^\circ/20^\circ$	$20^\circ/50^\circ\text{E}$	$195^\circ/15^\circ$	$15^\circ/\text{c.}90^\circ$?	$105^\circ/\text{c.}90^\circ$
REEVES			$200^\circ/50^\circ$	$25\text{-}40^\circ/\text{c.}90^\circ$	$105^\circ/30^\circ$	$105^\circ/\text{c.}90^\circ$
MacDONALD	$230^\circ/35^\circ$	$70^\circ/65^\circ\text{S}$	$160^\circ/50^\circ$?		

INTERPRETATION OF SULPHIDE/STRUCTURE RELATIONS

Dolomite zones within the Reeves Limestone member provide the host to mineralization, but the main structural control for ore localization, according to a number of geologists (notably Fyles and Hewlett, 1959), is Phase II folding, which is variously interpreted as preceding or in part synchronous with mineralization. The extent to which the sulphides have been deformed is a matter of some debate (Muraro, 1966) as is the significance of the epigenetic aspects of the deposits such as feathery terminations to ore zones, cross-cutting veinlets, etc.

MAJOR STRUCTURES INVOLVING SULPHIDES

At the Reeves mine, Phase II folding appears to have steepened the plunge of Phase I structures, as described above, without producing any comparable major Phase II folds. Thus the Reeves syncline, toward the hinge zone of which the ore is located, is almost certainly a Phase I structure, steepened and tightened by Phase II deformation. Comparison of the orientation of this structure as given by Addie, 233° at 48° (see p. 83, this volume), with the modal orientations of Phase I and II structures tabulated above shows the direction to be closest to that of the Phase I structures and the plunge to be that of the Phase II structures.

At the Jersey, most of the ore is along a series of Phase II folds; the "lead band," however, is concentrated under the nose of the A-zone skarn roll, which from its isoclinal style and recumbent attitude conforms to a Phase I structure. The "lead band," with its calcite augen and fragmented dolomite inclusions in a galena matrix, has the features of a cataclastic flow breccia. Bradley (this volume) points out that the "lower lead band" in the F-zone has a somewhat comparable structural setting, although verging toward the east. If involved in the

transposition deformation described above, such horizons should have behaved as the loci for sliding.

MINOR STRUCTURES INVOLVING SULPHIDES

In most of the folds, whether Phase I or II, involving sulphides, mineralization can be explained as an essentially post-deformation replacement, which, in general, outlines the fold structures. However, there are at both properties a number of structures, particularly the development of boudinage involving pyrite, which suggest that the sulphides may have undergone penetrative deformation.

At the Reeves, the pyrite-rich hanging wall, as observed in the Reeves glory hole and in the East MacDonald ore zone, displays a variety of such structures. Massive pyrite finely interbedded with dark graphitic layers is found in broken and brecciated folds, isolated within fine-grained galena-rich sulphides or in highly transposed dolomite (Fig. V-4). The sharp detail of the boundaries and the transition into breccia appear to preclude pyrite replacement. Close examination suggests that the folded layers are disrupted by boudinage, in which the graphitic layers behaved as the more competent mate-



FIGURE V-4.—Folded and disrupted bedded pyrite and graphite from Reeves MacDonald glory hole (magnification: $1\frac{1}{2}x$).

rial and failed by extension. Such extension following on compression is frequently produced during progressive deformation (Ramsay, 1967, p. 114). Every gradation can be found between this fine disruptive boudinage and superficially massive pyrite layers containing disoriented, finely fragmented graphitic layers as abundant inclusions. This association bears resemblance to sulphide-graphite schists found in a number of deformed sulphide deposits in Sweden and Finland (Marmo, 1960).

The style of the least disrupted folds is that of Phase I similar folds. Within the same vicinity, thin disseminated pyrite layers within dolomite also outline similar folds but display no hint of internal deformation.

At the Jersey, it is pyrite itself that is boudinaged within other sulphides or dolomite. Such pyrite boudins are found within sphalerite and dolomite in the D-zone, commonly within dolomite in the C-zone, and within galena in the G-zone. Sphalerite at the Jersey mine is typically subhedral in thin section, but in specimens from the A-zone skarn roll it displays an elongate to strongly schistose grain fabric locally developed within galena. Sphalerite in specimens from the lower A-, E-, and J-zones displays a similar fabric within more or less intensely deformed carbonate. The only internal deformation effect displayed by galena is a common curvilinear cleavage produced by translation gliding.

Such examples of deformation textures could be interpreted as representing either relict or local imposed fabric. However, since boudinage is a common feature throughout the Reeves Limestone and is related to transposition preceding or accompanying Phase II folding, the former seems more probable.

Pyrite that displays relict textural effects is not unusual and contrasts with the softer sulphides in which recrystallization or annealing may obliterate all signs of deformation, producing an apparently undeformed, allotriomorphic fabric (Vokes, 1969).

EFFECTS OF METAMORPHISM ON SULPHIDES

At the Jersey, there is good reason to suspect that widespread annealing recrystallization of the sulphides has resulted from contact thermal metamorphism by the Emerald and Dodger stocks to albite-epidote-hornfels and locally to hornblende-hornfels facies. According to Winkler (1967, p. 73) the transition between these facies generally occurs at 500° to 530° C. Calcite in the Reeves Limestone commonly appears to be recrystallized and increased in grain size with partial obliteration of earlier fabric. There is little such evidence at the Reeves mine, where the nearest intrusion is approximately 2 miles to the north, but the occurrence of more iron-rich sphalerite in cross-cutting fractures and calcite-filled veinlets is suggestive of migration under a thermal regime.

Lamprophyre dykes cut the ore zones in both mines, apparently transforming, as do granite dykes at Jersey, pyrite to pyrrhotite and producing more iron-rich sphalerite in adjacent marginal sulphides and in sulphide xenoliths. Assuming sulphides were deformed by Phase I folding, then the effects of regional metamorphism to greenschist facies must also be considered. The Reeves

area lies within the chlorite zone, and the Jersey within the biotite zone. The transition between the two zones is put by Winkler (1967, p. 99) at a temperature of 450° to 470° C. The effects of this metamorphism upon the sulphides are as yet undetermined. However, some of the epigenetic features of the ore, already mentioned, could have been produced during Phase I deformation by the local transport of sulphides under the influence of high fluid pressures and directed stress prevailing under such metamorphic conditions.

The obvious differences in texture, grain size, and composition of the sulphides between the areas have been related in the past to contact metamorphic effects at Jersey, but these differences may be hybrid.

CONCLUSIONS

A structural evolution for the host rocks of Salmo-type deposits is proposed, in which Phase I and Phase II folding are seen to be essentially homoaxial and the product of continuous progressive deformation. The significance of Phase III structures is not entirely clear; they are tentatively related to late-orogenic deformation accompanying granitic intrusion in the area.

Deformation recognized in the sulphides can be related to the same structural framework, suggesting that the sulphides have been involved in Phase I deformation and hence in regional metamorphism.

The relict nature of the sulphide fabric is ascribed to general recovery following upon regional metamorphism and/or to annealing recrystallization promoted by subsequent thermal metamorphic events (McDonald, 1967).

Deformation of the sulphides by no means proves a syngenetic origin, but it is perhaps implied. The association of bedded pyrite and graphite with a slightly siliceous dolomite facies of the Reeves Limestone is the only link yet recognized with an original sedimentary environment. Around this tenuous evidence several, probably equally valid, genetic models could be set up.

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VI

THE METALINE DISTRICT, PEND OREILLE COUNTY, WASHINGTON

By

GEORGE G. ADDIE

THE METALINE DISTRICT, PEND OREILLE COUNTY, WASHINGTON

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THE METALINE DISTRICT, PEND OREILLE COUNTY, WASHINGTON

By GEORGE G. ADDIE^①

LOCATION AND PREVIOUS WORK

The Metaline mining district is in Pend Oreille County in the northeastern corner of the State of Washington. It is about 10 miles south of the International Boundary, near the town of Metaline Falls (Yates, Fig. III-2, this guidebook).

A comprehensive study of the district was made by Park and Cannon (1943), and the complex structures within the graben and the ore bodies were later studied by Dings and Whitebread (1965). These publications were followed by McConnel and Anderson's (1968) detailed account of the geology of the mines that placed emphasis on emerging new concepts of origin of the ores. Much of the following discussion of the general geology of the district has been summarized from these reports.

HISTORY AND PRODUCTION

Lead and zinc minerals are said to have been discovered in the Metaline district as early as 1896. Exploitation was delayed in part because of the district's remoteness from transportation. First recorded production was in 1906, but even after rail transport became available development was slow, and until 1929 production was small and erratic. The first concentrator was built at the Grandview mine in 1937. A small flotation plant was installed at Pend Oreille Mines & Metals Company's West Side mine in 1930, and an enlarged mill was built in 1937. Substantial production was not achieved until 1938, when nearly a quarter of a million tons of ore was mined. Pend Oreille's 2,400-ton-per-day East Side mill was constructed in 1951. About this time a highly mechanized system of mining was adopted, and in 1957 production reached nearly 1 million tons of ore. Present production is at the rate of 200,000 tons per year, all from the East Side mine, about 1½ miles north of Metaline Falls.

From 1906 through 1969 the district has produced 657,109 ounces of silver, 196,652 tons of lead, and 418,202 tons of zinc. A substantial amount of cadmium has also been produced. The grade of the ore mined has varied with the mining method used. Before mechanization of the mines in 1950 the average grade was about 0.043 ounce silver per ton, 1.38 percent lead, and 3.60 percent zinc. After mechanization the average grade was about 0.036 ounce silver per ton, 0.84 percent lead, and 2.31 percent zinc (Weissenborn, 1966, p. 139). Recently more selective mining methods have been used, and they have resulted in a higher average grade of ore than that quoted in 1966.

Production has been almost entirely from the Pend Oreille, Grandview, and Metaline mines. Only the Pend Oreille mine is presently in operation. Figure VI-1 is a plan of the Pend Oreille and Grandview mines, the two principal producers. Figure VI-2 shows sections of these mines.

STRATIGRAPHY OF THE DISTRICT

The stratigraphic sequence in the Metaline district has been summarized by Dings and Whitebread (1965, p. 5) as follows:

The rocks in the Metaline district range in age from Cambrian, or possibly older, to Tertiary. Quaternary deposits of surficial material mask large areas. The formations of Paleozoic age include, in ascending order, the Monk Formation (Cambrian?), Gypsy Quartzite (Cambrian), Maitlen Phyllite (Lower or Middle Cambrian), Metaline Limestone (Middle Cambrian), Ledbetter Slate (Ordovician), and an unnamed group of rocks, largely dark-gray to black argillite of Silurian and Devonian ages. Small areas of poorly consolidated rock of the Tiger Formation of Tertiary age occur locally. The Quaternary deposits include silts and sands of lacustrine origin, and glacial and glaciofluvial material consisting of sand, gravel, cobbles, and boulders. Small deposits of alluvium occur at a few places. Igneous rocks are rare, although wide areas of the Kaniksu batholith (chiefly quartz monzonite that grades into granodiorite) crop out a few miles distant. In the mapped area a few dark and generally narrow dikes of probably Late Cretaceous or Early Tertiary age are widely scattered. Olivine trachybasalt occurs at one locality northwest of Ledbetter Lake as a partly dissected flow resting unconformably on Silurian or Devonian strata.

Precambrian rocks of the Priest River Group, the Shedroof Conglomerate, and the Leola Volcanics crop out east of the productive part of the area under discussion (Fig. VI-3).

The Metaline Limestone is the host rock for the important ore deposits of the Metaline district. It has been divided into three principal lithologic units: a lower thin-bedded limestone shale sequence 1,000 to 1,200 feet thick; an intermediate light-gray bedded dolomite 3,500 feet thick; and an upper gray massive limestone not everywhere present in the district. Total thickness of the formation is given by Dings and Whitebread as about 5,500 feet and by McConnel and Anderson as from 4,500 to 6,500 feet. The Metaline is essentially the equivalent of the Nelway Formation of British Columbia.

STRATIGRAPHY OF THE "JOSEPHINE HORIZON"

The host rock for all the important known ore bodies is an irregular zone known as the "Josephine Horizon" that occurs at the top of the Metaline Limestone. The base of the zone is 35 to 200 feet stratigraphically below the overlying Ledbetter Slate. Actually, the major tonnage that the mines have produced has been extracted from an interval within about 70 feet of the slate.

The "Josephine Horizon" as considered by past authors, forms a distinctive unit that is characterized by its ir-

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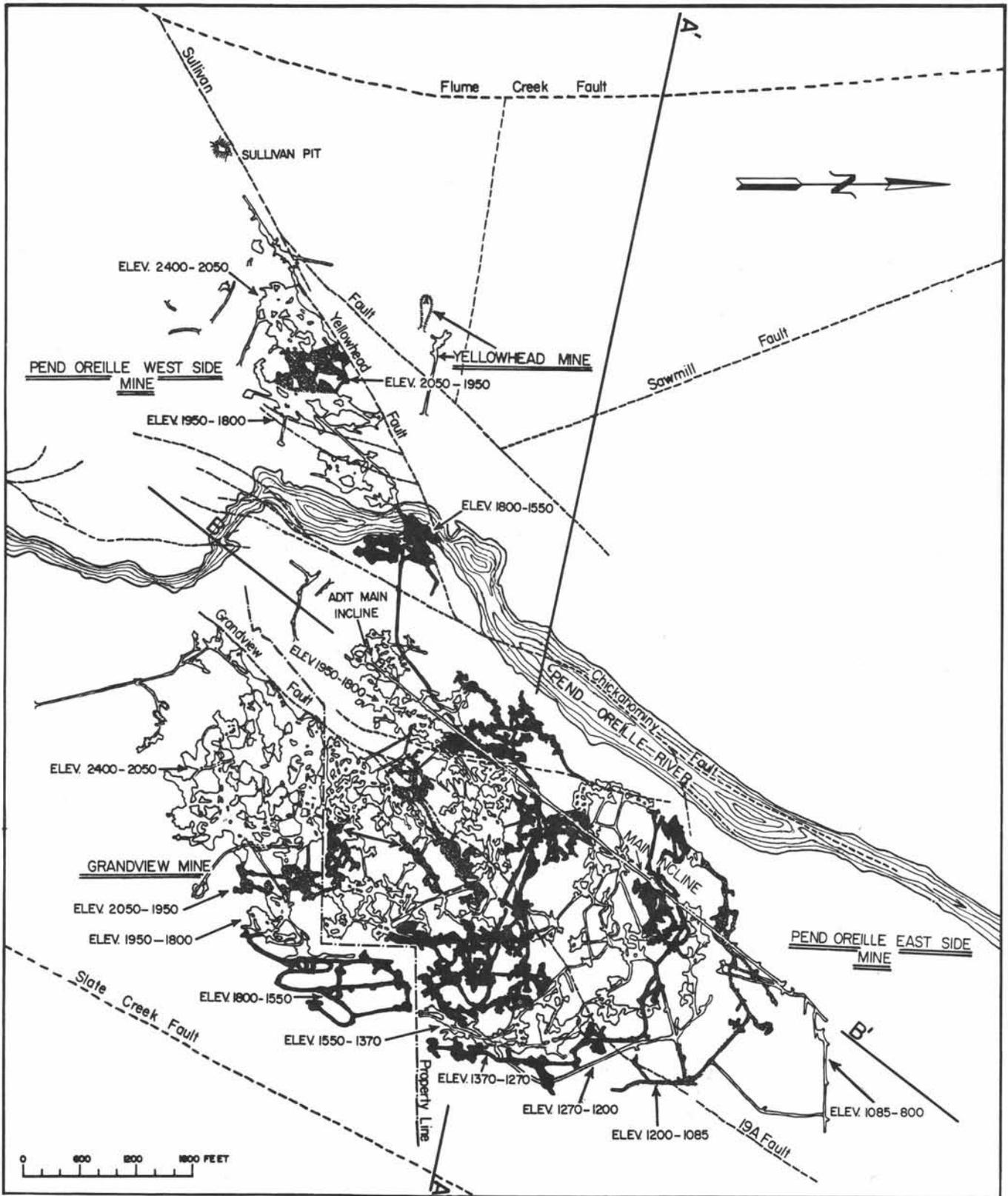


FIGURE VI-1.—Pend Oreille-Grandview mine workings. (Published with permission of the American Institute of Mining, Metallurgical, and Petroleum Engineers, Inc. from Graton-Sales Volume II, Ore deposits of the United States, 1933-1967.)

regular brecciation and the diversity of rock types within it. It crops out in the vicinity of the Metaline, Grandview, and Pend Oreille mines (Fig. VI-3). It is exposed near Slate Creek and east of the Pend Oreille River along the Lead Hill fault, as well as west of the Pend Oreille River north of Slate Creek and east of the river north of Z Canyon. It is found in most drill holes that have penetrated beneath the Ledbetter Slate west of the river from south of the Metaline mine to the northernmost drill holes west of Boundary Dam. As described by McConnell and Anderson (1968), this complex rock unit, which lies immediately below the Ledbetter Slate, forms an irregularly brecciated "stratum" ranging in thickness from a few feet to over 200 feet. Five types of fragments and blocks form the breccia and are present in highly variable proportions. The types are:

- 1) Medium- to light-gray moderately fine-grained dolomite, some of which may show banding or lamination.
- 2) Light-gray coarse-grained nonbedded dolomite.

- 3) Abundant "zebra rock"² (Fig. VI-4).
- 4) Dense gray sublithographic limestone.
- 5) Black jasperoid.

Cementing the breccia is a matrix of dense, black, carbon-rich dolomite or dark jasperoid, or a mixture of the two. Fragments within the matrix are usually disoriented and range in size from a fraction of an inch to blocks 8 to 10 feet on a side. Within the "Josephine Horizon" there are, in addition to the breccia, discontinuous beds and lenses of bedded and nonbedded dolomite and gray limestone.

The "Josephine Horizon" is regarded by Dings and Whitebread as hydrothermally metamorphosed Metaline Limestone and is shown as such on Figure 1 of Professional Paper 489 (Dings and Whitebread, 1965). McConnell

²"Zebra rock" is a coarsely crystalline dolomite conspicuously striped alternately dark and light gray. Stripes are 1/16- to 1/4-inch thick and may be as much as several feet long. Stripes are nearly parallel and are irregularly curved or gently undulating. Another form exhibits spheroidal forms up to several inches in diameter that commonly are found in clusters several feet across. Occasional moundlike agglomerations attain dimensions of many feet. McConnell and Anderson (1968) state the gross forms and distributions of the "zebra rock" suggest an algal origin.

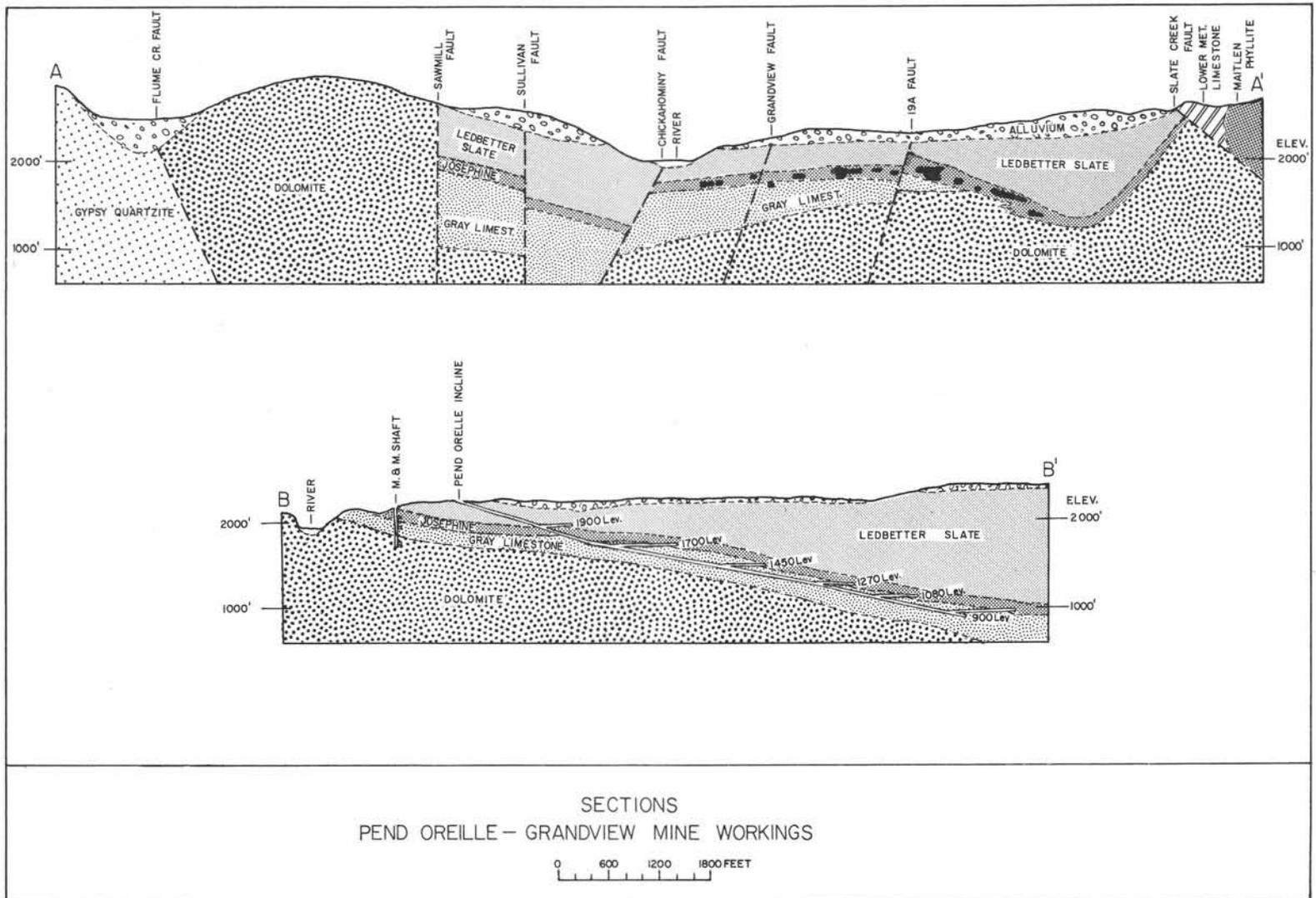


FIGURE VI-2.—Sections. Pend Oreille-Grandview mine workings. (Published with permission of the American Institute of Mining, Metallurgical, and Petroleum Engineers, Inc. from Graton-Sales Volume II, Ore deposits of the United States, 1933-1967.)

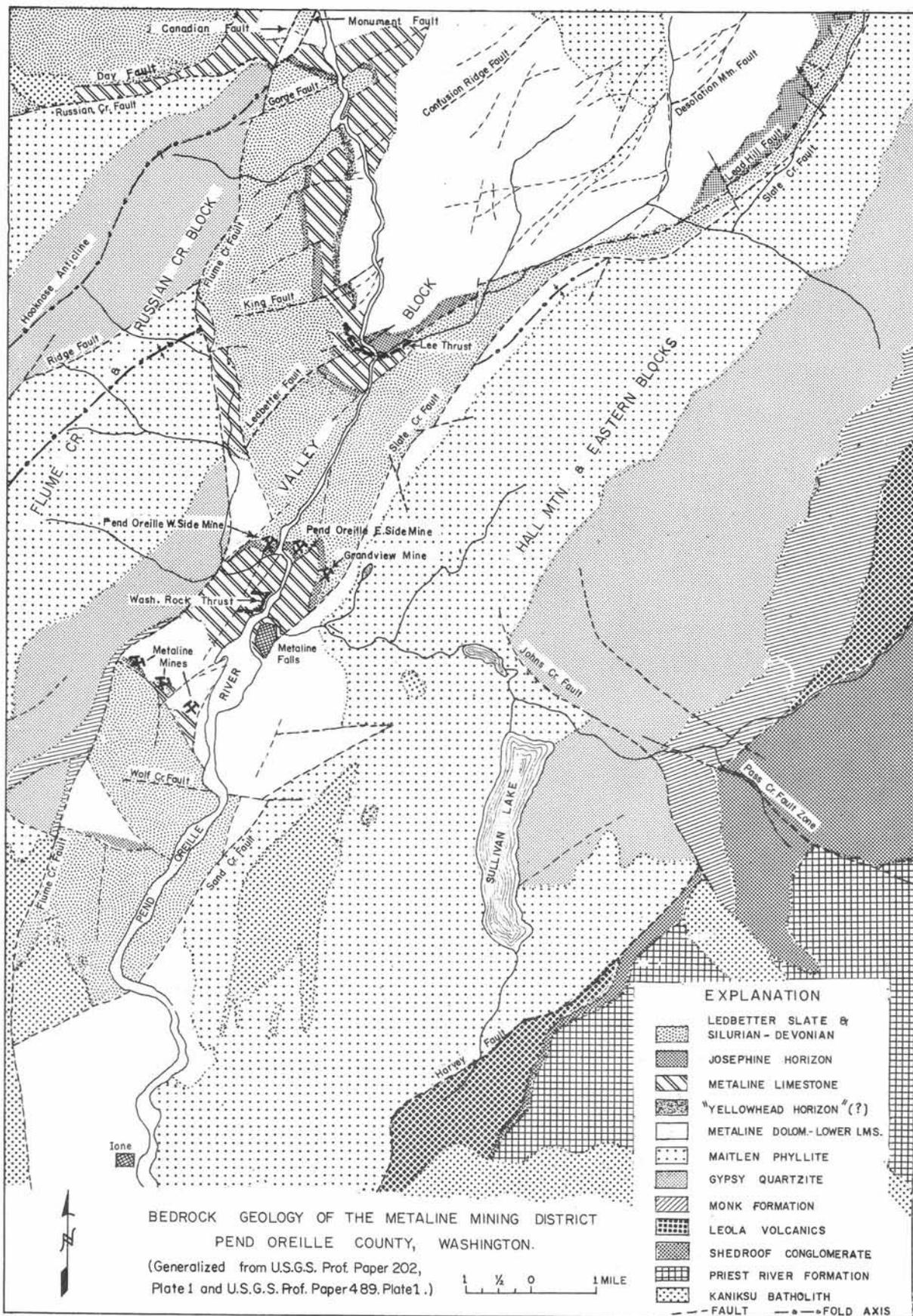


FIGURE VI-3.—Bedrock geology of the Metaline mining district, Pend Oreille County, Washington. (Published with permission of the American Institute of Mining, Metallurgical, and Petroleum Engineers, Inc. from Graton-Sales Volume II, Ore deposits of the United States, 1933-1967.)

nel and Anderson (1968), on the contrary, regard the "Josephine Horizon" as an irregular and discontinuous original stratigraphic unit overlying the gray limestone unit of the Metaline Formation. They believe that it developed as a result of sedimentary and erosional processes in a reef environment. These different interpretations have an important bearing on concepts of ore genesis.

The writer considers the "Josephine Horizon" to consist of two units: the hanging wall zone and the ore zone. The hanging wall zone consists of dark, sometimes black, dolomite, usually slightly mineralized with sphalerite and with abundant gash veins. The ore zone is a brecciated jasperoid zone that ranges from 20 to 50 feet in thickness.

Although the ore zone has been described as a jasperoid breccia, occasionally one finds thin beds of argillite within it. The argillite has been crumpled, contains some boudinage, and includes textures that look very much like thixotropic deformation (Fig. VI-5).

Lateral facies changes in the host rock that at many places nearly coincide with the margins of ore bodies suggest that rock facies had some control in ore deposition.

The ore zone in the present mining area is underlain by a limestone unit known locally as the Gray Metaline Limestone. To the east the gray limestone changes and its place is taken by gray dolomite. Drill-hole data suggest that the ore zone continues as a slightly mineralized light-gray, sucrose-textured crystalline dolomite about 50 feet below the Ledbetter Slate. At the point where the gray limestone gives way to the crystalline dolomite, the "zebra rock" or "algal" material comes into prominence (Fig. VI-6). At this point, the amount of the ore sulphides decreases sharply and the amount of pyrite increases sharply.

The gray dolomite facies is relatively deficient in ore but is the locale for pyritic zones. A second, slightly stronger "mineral horizon" seems to be present about 100 feet below the Ledbetter in the same kind of dolomite.

The narrow zone across which the facies change occurs trends about N. 25° E., which is about the same as the trend of the crest of the "Grandview Anticline" discussed below under "Structure."

The writer believes that a "special facies" was important in the deposition of the ore. Maucher and Schneider



FIGURE VI-4.—"Zebra rock." C 12 area, Pend Oreille mine.

(1967) made a similar suggestion for Middle Triassic Alpine lead-zinc deposits.

STRUCTURE

In describing the regional structure, Dings and Whitebread (1965, p. 6) wrote:

The geologic structures are generally moderately complex, although locally they are extremely complicated. Structures include folds, faults, and widespread fracture zones with prevailing northeasterly trends. The dominant structure is a wedge-shaped graben—the valley block or graben—which along its western border is marked by a fault with a stratigraphic throw of 10,000 to 12,000 feet. The stratigraphic throw of the fault that bounds the graben on the east is not precisely known, although it is locally a minimum of 6,000 to 7,000 feet. Several stages of deformation are recognized.

In the Metaline district the productive ore bodies are mainly restricted to this wedge-shaped, structurally depressed block or graben between the Slate Creek and Flume Creek faults. The graben is about 16 miles long and nearly 9 miles wide at its maximum width just south of the International Boundary (Fig. VI-3).

Within the mine area the most significant structure is the Grandview Anticline (Dings and Whitebread, 1965, p. 35). It plunges about 10° N. 25° E. Whether the axial plane is overturned or not is not known.

ORE DEPOSITS

The ore bodies in the "Josephine Horizon" have been described in great detail by McConnel and Anderson (1968), and only a brief description will be given here. The ore bodies range from pods and lenses a few feet thick and 10 to a few tens of feet wide and long, to masses 3,000 feet long, 100 feet thick, and 300 feet wide. The ore bodies are highly irregular, and many are elongated in channel-like shapes that pinch and swell, and coalesce irregularly with nearby ore pods (Fig. VI-1). The longest elongated bodies plunge northeastward at 10° to 20° .

The ore consists of an intimate mixture of sphalerite and galena in a wide range of proportions. This range is sufficient that certain parts of the ore bodies can be characterized as lead-rich areas, others as zinc-rich areas. Silver minerals have not been identified. Pyrite is the common gangue sulfide. Other gangue minerals are quartz, calcite, dolomite, barite, and paligorskite.

Sphalerite occurs chiefly as fine disseminations in the matrix of the breccias, and also rims breccia fragments. In "zebra rock," sphalerite may form narrow bands along one edge of white dolomite streaks. Coarser grained sphalerite with quartz forms irregular veinlike streaks and patches in the matrix and between breccia fragments. It may occur also within white, coarse-grained dolomite fragments that are fragments of the breccia. Sphalerite also replaces fine-grained gray limestone, either as coarse-grained aggregates or as fine-grained masses in which individual grains cannot be detected in hand specimens. The color of sphalerite ranges from reddish brown to pale yellow.

Most of the galena is distributed irregularly throughout the breccia as stringers, pods, and segregations. It is ordinarily medium grained, but large cubes are found occasionally. In places it fills faults and fractures in the breccia matrix.

Small, irregular bodies and scattered grains of sphalerite and galena are found throughout the entire stratigraphic thickness of the Metaline Limestone, except the very basal part. All the presently mined ore bodies, however, are in the upper 200 feet of the Metaline Limestone. Most ore bodies are found in the breccia. Most of the breccia contains at least traces of sphalerite and galena, but not all of it is ore. Very few ore bodies are in contact with the Ledbetter Slate.

A small tonnage of more pyritic ore was mined some years ago from a second zone, known locally as the "Yellowhead Horizon," 1,100 to 1,200 feet stratigraphically below the Ledbetter Slate. This zone is exposed in only a few places and has been little explored in the Metaline district. It has been a productive ore zone to the west in parts of Stevens County, Washington.

In the Metaline district no ore bodies have been found in the Maitlen Phyllite, which underlies the Metaline Limestone, although important ore bodies are found in carbonate rocks in Stevens County, Washington, and in its Canadian equivalent, the Laib Formation in adjoining parts of British Columbia. Minor amounts of ore have been found at the Oriole mine in a limy zone near the base of the Gypsy Quartzite (Yates, this guidebook), which stratigraphically underlies the Maitlen. The Oriole is west of the graben outside the main producing area of the district.

GASH VEINS

In the hanging wall of the ore zone are prominent gash veins that end at the top of the ore zone (Fig. VI-7). They strike east-west and dip 80° S. The veins consist of quartz and calcite, the quartz crystals being oriented perpendicular to the vein walls. A few of these veins contain small amounts of sphalerite and galena at or near the upper part of the ore horizon. Figure VI-8 shows a gash vein whose wide end where it meets the top of the ore zone consists of galena that shows flow structure. At the Reeves MacDonald mine, a few miles north of the International Boundary, some similar gash veins normally contain abundant sphalerite and galena, suggesting remobilization.

The writer thinks that the gash veins may be tensional features genetically related to the Phase II folding described by J. T. Fyles in an article in this guidebook. If, on the other hand, the gash veins are tensional features genetically related to the Grandview anticline, then the anticline must be overturned to the west.

At the Reeves MacDonald mine it has been found that the plunge of the ore bodies is normal to the plane of the gash veins. If it is found that the same structural relations are present in the Metaline district, the ore bodies plunge 10° due north, i.e., normal to the gash veins, not 10° N. 25° E., parallel to the plunge of the Grandview anticline.

Such apparently simple structural control could be disrupted if the trend of the "special facies" diverges significantly from either of the above plunges.

LEAD ISOTOPES

Sinclair (1966) studied the isotopic composition of a number of galena specimens from mines in the Kootenay Arc. He also used analyses of 5 specimens from the Metaline district. Because of the isotopic compositions, he concluded that leads with different histories were involved. He implied that the lead in the present-day deposits had its source in Precambrian Purcell (Belt) rocks, and he postulated that the lead was introduced into the Cambrian host rocks by hydrothermal solutions, probably during the Middle Jurassic Coast Range Orogeny.

HYPOTHESIS OF ORIGIN

HYDROTHERMAL

Dings and Whitebread (1965, p. 68) favor a hydrothermal origin. They state:

The ore bodies of the Metaline district are believed by the authors to have been deposited chiefly by hydrothermal solutions genetically related to the magma that in part formed the Kaniksu batholith and smaller outlying bodies of rock. The ores were probably introduced into the carbonate rock late in the igneous history, after partial or complete solidification of the batholith, as postulated by Park and Cannon (1943, p. 53). They noted the similarities between the replacement deposits in the Metaline Limestone and the replacement bodies in the igneous-metamorphic zone (situated beyond the area mapped for the present investigation), and they tentatively



FIGURE VI-7.—East-west tension gash veins in hanging wall dolomite. Pend Oreille mine.

assumed that both types were deposited during one period of mineralization.

The detailed features of the individual ore bodies, such as irregular brecciation, abundance of closely associated alteration products, and inconsistent and confusing paragenesis of the principal minerals, indicate that the mode of formation was moderately complex and probably extended over a considerable period of time. One of the principal factors causing these features is a repeated settling of the graben block during mineralization, resulting in the development of crushed zones, fractures, and faults which periodically afforded channelways for the migrating hydrothermal fluids.

Sinclair (1966, p. 253) also favors a hydrothermal origin for at least the lead.

McConnel and Anderson (1968, p. 1479-1480) favor the introduction of the metals during the sedimentary and diagenetic stages of the host rocks, with later mobilization as a result of intrusion of the nearby Kaniksu batholith.

REMOBILIZATION

Galena and sphalerite in gash veins (Fig. VI-8) is evidence that some remobilization has occurred. The apparent structural control of the ore bodies suggests that if the ore minerals were syngenetic they have at least in



FIGURE VI-8.—Gash veins filled with calcite and replaced at top of ore zone by galena with flow structure.

part been remobilized. And Sinclair's (1966, p. 253) interpretation that Kootenay Arc leads, including lead from the Metaline district, are the product of the mixing of two leads of different parentage further suggests remobilization.

CONCLUDING REMARKS

Although the origin of the ores is far from proved, the writer favors the following model: A "special facies," similar to that described by Maucher and Schneider (1967), contained syngenetic pyrite that was later replaced, perhaps during diagenesis, by sphalerite. Much later the lead was introduced, probably by hydrothermal solutions. Still later the zinc and lead were slightly remobilized due to tectonism.

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VII

THE REEVES MacDONALD MINE, NELWAY, BRITISH COLUMBIA

By

GEORGE G. ADDIE

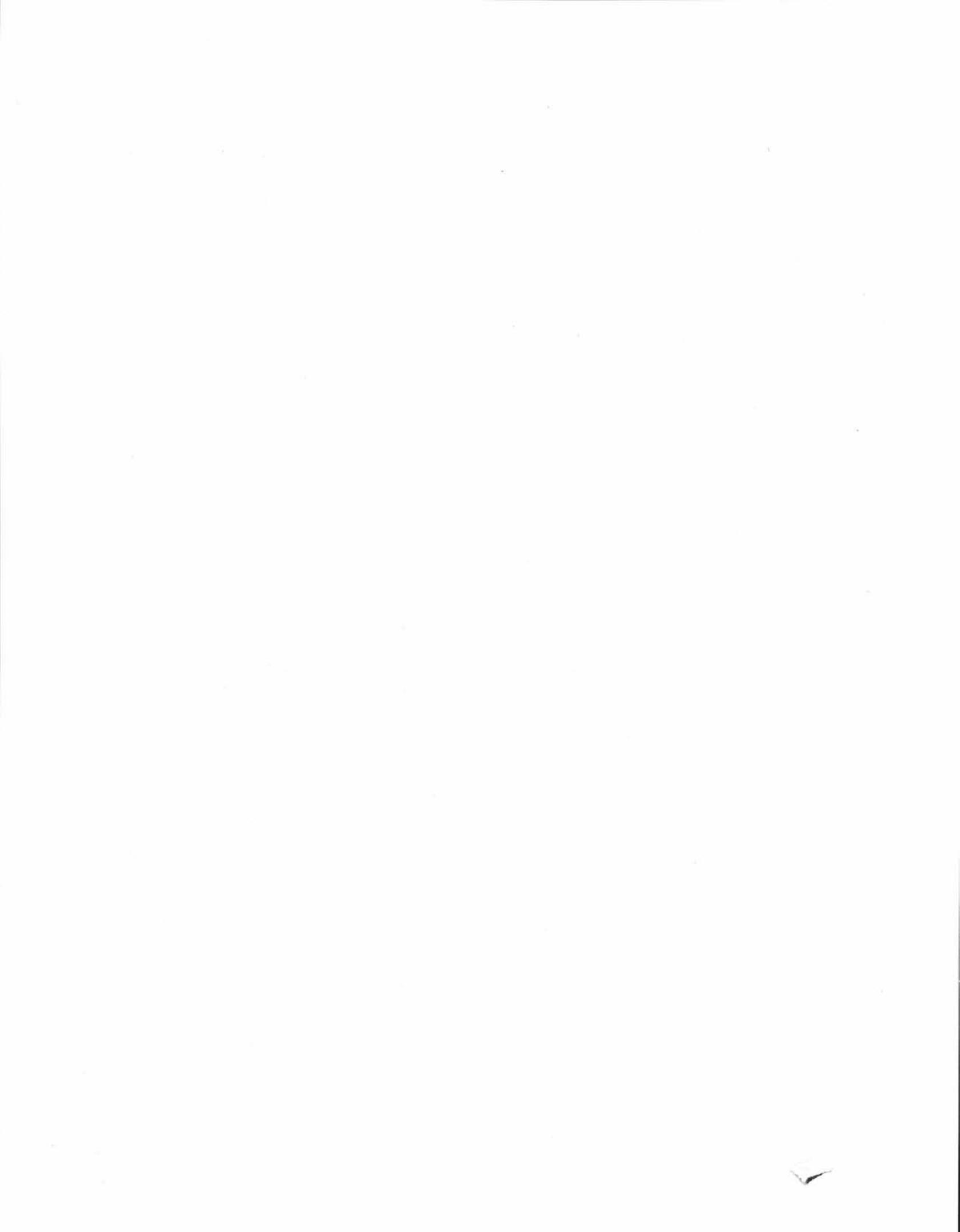
THE REEVES MACDONALD MINE, NELWAY, BRITISH COLUMBIA

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THE REEVES MacDONALD MINE, NELWAY, BRITISH COLUMBIA

By GEORGE G. ADDIE^①

INTRODUCTION

The Reeves MacDonald mine is on the Pend Oreille River about 1¾ miles north of the United States-Canada boundary and about 1 mile southeast of the mouth of the Salmo River (Fyles, Fig. IV-1, this volume). The mine is in the Kootenay Arc, a major geologic structure that is discussed elsewhere in this guidebook. The zinc-lead ore bodies at the Reeves MacDonald mine are sulphide replacement deposits in dolomite zones in the Reeves Limestone Member of the Laib Formation.

PRODUCTION

From 1949 through 1969, a total of 6,279,435 tons of ore was mined. Concentrates from the ore contained 437,718,698 pounds of zinc and 121,496,143 pounds of lead.^②

STRATIGRAPHY

In the vicinity of the mine the oldest rock exposed is quartzite of the Quartzite Range Formation. This formation is conformably overlain by quartzite of the Reno Formation. The Reno Formation is conformably overlain by the Laib Formation. All three formations are Early Cambrian. The Quartzite Range and Reno Formations have been correlated with the top part of the Gypsy Quartzite in the Metaline mining district to the south in northeastern Washington, and the Laib Formation has been correlated with the Maitlen Phyllite in the same area (Fyles and Hewlett, 1959, p. 17).

The Laib Formation has been divided into four members (Fyles and Hewlett, 1959; Yates, Fig. III-4, this volume). The bottom member, about 100 feet thick, comprises interbanded schist and limestone and is known as the Truman Member. The next overlying member, about 300 feet thick, is the Reeves Limestone, a banded grey and white limestone with dolomitic lenses. In some areas almost the entire thickness of the Reeves has been dolomitized. The third member, about 300 feet thick, is a black calcareous argillite called the Emerald Argillite. The top member, the Upper Laib Member, is missing in the immediate mine area. The next overlying formation, the Nelway Formation of Middle Cambrian age, is the correlative of the Metaline Limestone in the Metaline district. It is absent in the vicinity of the Reeves MacDonald mine. To the south of the mine area a high-angle reverse fault of regional extent has brought rocks of the Active Formation, of Ordovician, Silurian(?), and Devonian(?) age (Fyles and Hewlett, 1959, p. 34) into contact with the Laib Formation. The Active Formation strati-

graphically overlies the Nelway Formation, and is the Canadian equivalent of the Ledbetter Slate.

Rocks near the Reeves mine display a low degree of metamorphism. The limestone has been recrystallized to a coarser grain size, and chlorite and muscovite have been developed in the underlying argillaceous and quartzitic strata. The rocks are highly deformed and in places display intense crumpling and drag folding. All noncalcareous rocks, except white quartzite, possess a well-developed cleavage that parallels the bedding. A few miles north of the mine, rocks around granitic stocks are intensely thermally metamorphosed, but this type of metamorphism has not been found in the vicinity of the mine.

STRUCTURE

In the mine area the formations strike N. 70°-80° E. and dip 50°-70° S. The most prominent structural feature in the area is the isoclinal Salmo River anticline, whose axial plane strikes N. 70°-80° E. and dips 60° S. The strata on the north limb of the anticline are overturned (Fig VII-1). Steeply plunging, smaller isoclinal folds that roughly parallel the Salmo River anticline have been superimposed on both limbs of the large anticline. The Reeves ore body occupies one of the larger synclinal troughs developed on the south limb of the Salmo River anticline (Fig. VII-2).

A series of faults that strike approximately north-south and dip 25°-40° E. offset the formations in the mine area. Apparent relative movement on the faults was left-lateral in plan (Fig. VII-1) and normal in the east-west section shown on Fig. VII-3. Five of these faults have been named; from east to west they are: No. 4, O'Donnell, B. L., Reeves, and MacDonald. Each of these faults cuts and offsets what seems once to have been a continuous ore body. On each fault the ore has been shifted downward and eastward. A reconstruction of the Reeves ore body gives a plunge length of at least 5,000 feet before faulting.

ORE BODIES

Figure VII-4 shows a typical level plan for the Reeves ore body. The ore body is located entirely within a dolomitized zone of the Reeves Limestone. The limestone is underlain to the south by the interbanded schists and limestone of the Truman Member. East from the Reeves ore body the Reeves Limestone narrows progressively in plan in such a manner as to suggest a westward-plunging attenuated syncline. The ore body has a synclinal form and is known to plunge 48° S. 53° W. The synclinal form of the Reeves ore body parallels the tight syncline in the Reeves Limestone.

^①Chief Geologist, Reeves MacDonald Mine, Ltd.

^②Records of the British Columbia Department of Mines and Petroleum Resources show that, in addition to the lead and zinc reported above, the Reeves MacDonald mine to the end of 1968 yielded 582,075 ounces of silver, 2,504,769 pounds of cadmium, and 39,830 pounds of copper (The Editors).

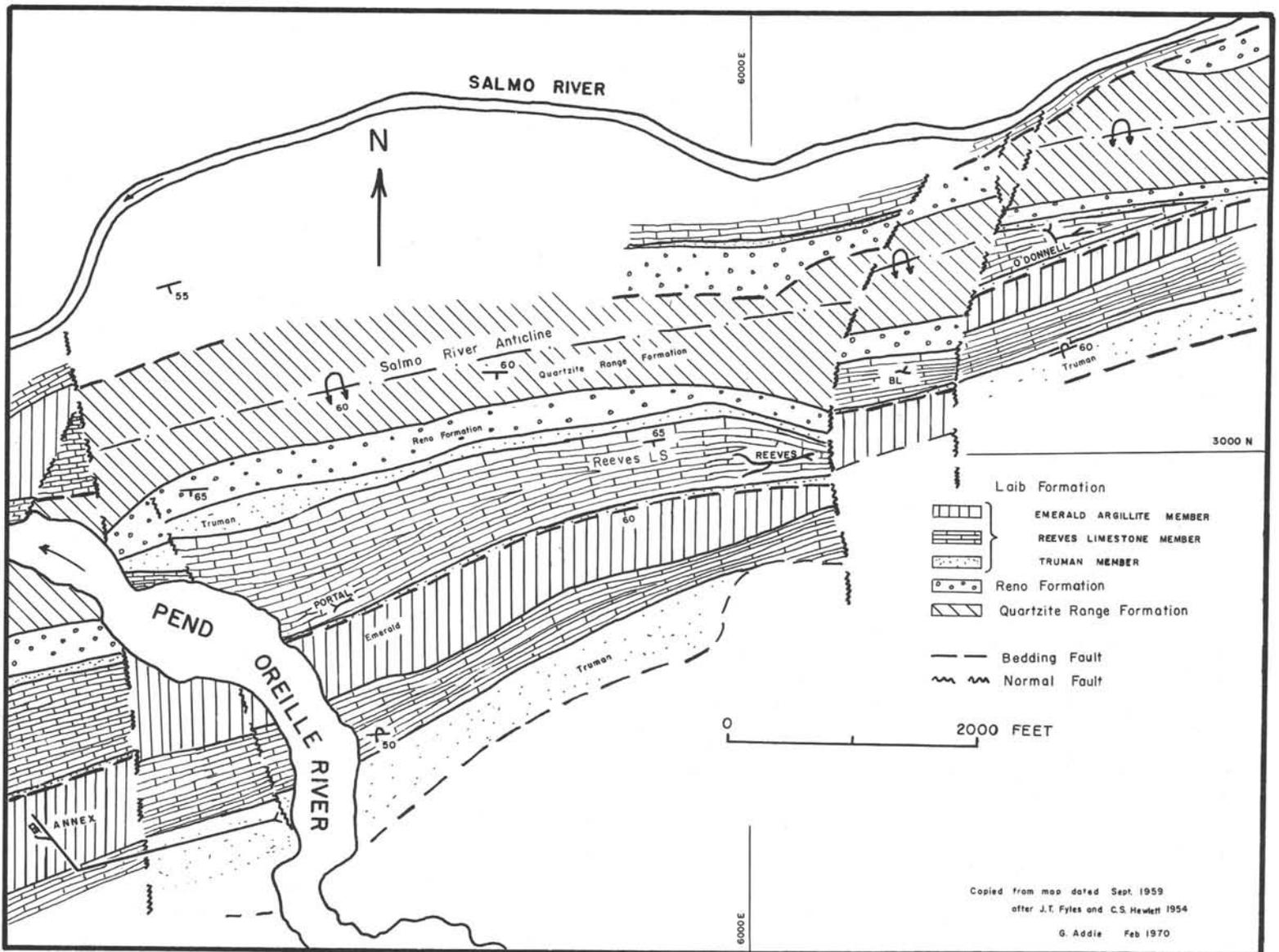


FIGURE VII-1.—Surface geology of the Reeves MacDonald mine.

The main part of the ore body in the trough of the syncline is about 300 feet long. At its west end the mineralized zone progressively narrows and finally pinches out in an envelope of dolomite. At its east end the main part of the ore body becomes 75 to 80 feet thick before it splits into two branches, the hanging wall branch and the footwall branch. These branches are 25 to 30 feet thick and extend to the east about another 300 feet. The shape of the alteration halo of dolomite around the ore body closely resembles the shape of the ore body. The area between the hanging-wall and footwall branches is known as "the split," and is occupied by well-banded blue-grey and white limestone.

Considerable drag folding occurs along the footwall branch of the ore body. The plunge of the drag folds and the plunge of "the split" are about the same as the plunge of the main part of the ore body and the plunge of the Salmo River anticline, 48° S. 53° W. The near parallelism of the plunges of these several features results in outlines

of the ore body as seen in plan that are very similar over a large vertical distance, and this feature enables the horizontal outline of the ore body to be projected down the plunge for several hundred feet with remarkable accuracy.

The principal sulphides in the ore are pyrite, honey-coloured sphalerite, and galena. They replace medium-to dark-grey, banded dolomite. The bulk of the ore has a banded appearance, and locally it may be brecciated. A zone of breccia ore extends along the footwall part of the ore in the trough of the syncline and continues along the footwall of the footwall branch. This breccia ore has angular fragments of limestone and dolomite scattered through a matrix of massive sulphides, mainly pyrite. Lenses of unmineralized light-grey dolomite 2 to 5 feet thick and 10 to 20 feet long are not uncommon in the ore zone, and thin bands of argillite, showing thixotropic deformation, also occur in the ore zone and in its hanging wall (Fig. VII-5).

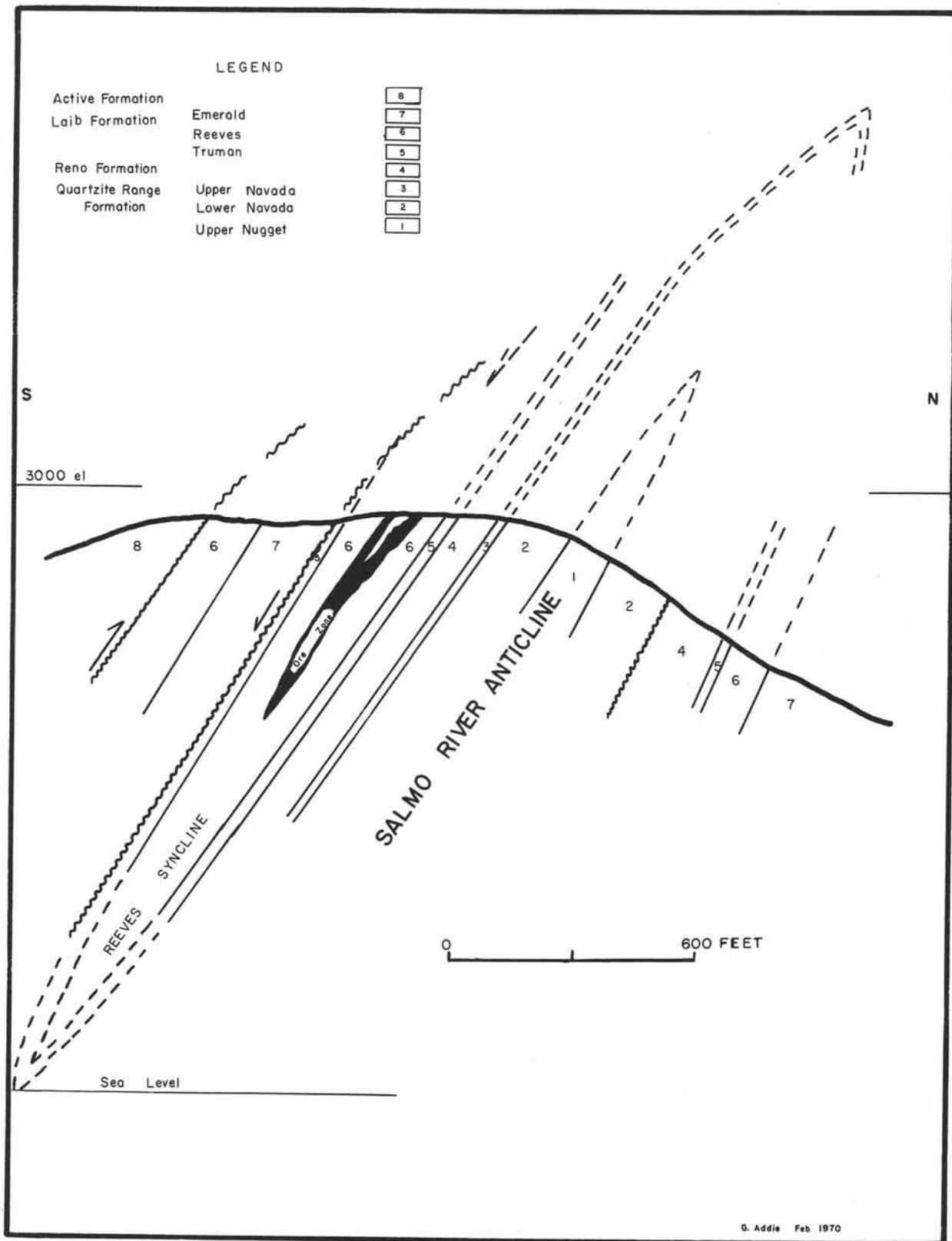


FIGURE VII-2.—Diagrammatic north-south section 6000 E., looking west.

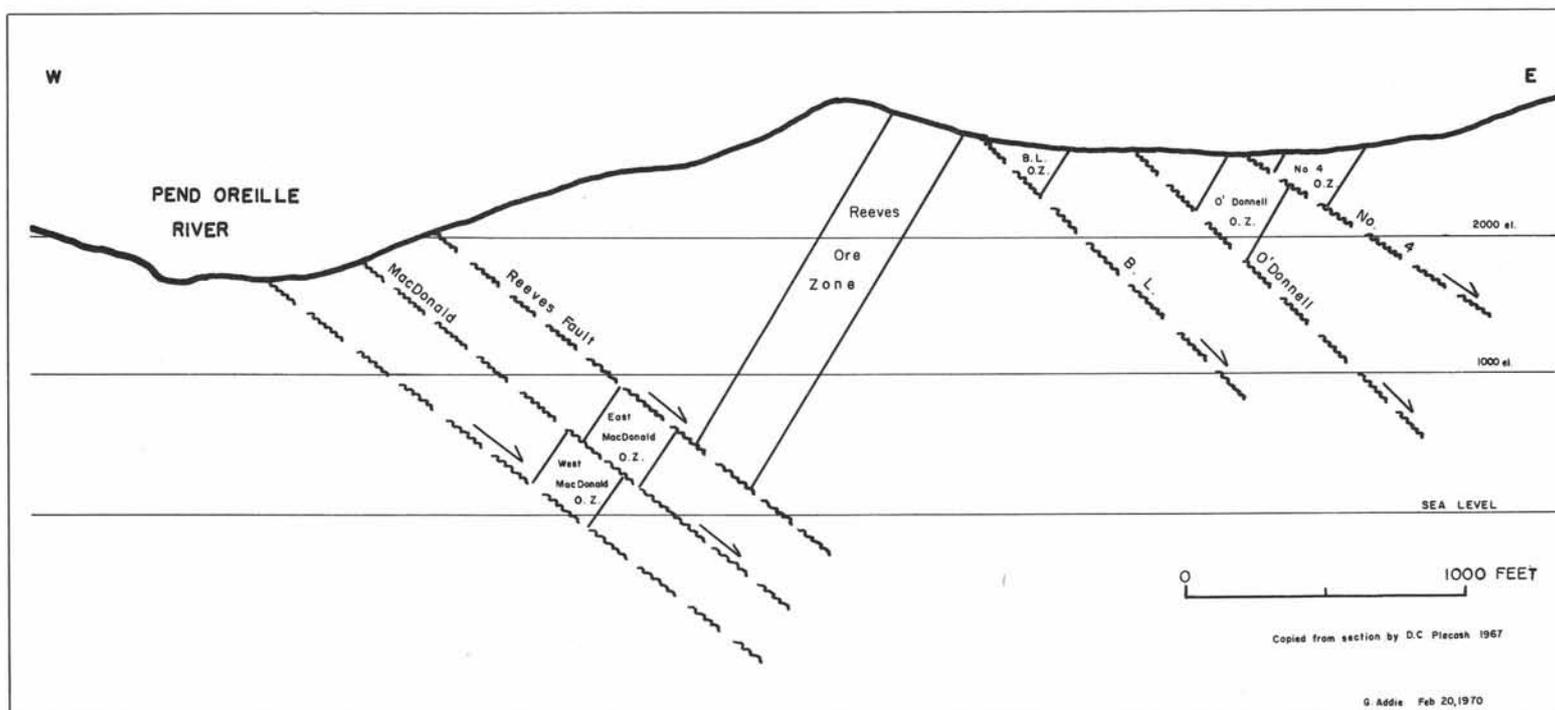


FIGURE VII-3.—Diagrammatic east-west projection of Reeves MacDonald mines.

The No. 4, O'Donnell, B. L., Reeves, East MacDonald, and West MacDonald ore zones (Fig. VII-3) have essentially the same shape, size, plunge, and grade. Whereas oxidation of the ore minerals in the No. 4 and O'Donnell ore zones extends to considerable depth below the surface, sulphides in the other ore zones show very little oxidation.

GASH VEINS

Gash veins that fill tension openings occur within, and are younger than, the main ore body (Fig. VII-6). Locally, they are well mineralized with galena and sphalerite; the sphalerite is reddish brown, not honey yellow as in the replacement ore. The veins strike about N. 35° W. and dip about 40° NE. With this attitude, the plane of the gash veins is essentially normal to the plunge of the main ore body.

LEAD ISOTOPES

In a study of the isotopic composition of lead in galenas from several mines in the Kootenay Arc, Sinclair (1966, p. 249) recognized anomalous leads that have been interpreted to have gone through complex histories. He implied that the leads were derived from pre-existing leads in Lower Purcell (Belt) strata and he postulated that "by local melting and eventual production of hydrothermal fluids" these different anomalous leads were introduced into their host rocks and deposited "during Coast Range Orogeny, probably about Middle Jurassic time" (Sinclair, 1966, p. 253).

GENETIC HYPOTHESES

HYDROTHERMAL

Sinclair's interpretation of the lead isotope data states that the lead was deposited from hydrothermal solutions. Sinclair does not postulate an origin for any of the other metals in the ore. And perhaps the close structural control of the ore body also argues for a hydrothermal origin for the ore.

SYNGENETIC

The presence in and near the ore zone of argillite that shows thixotropic deformation textures suggests that the ore occurs in and is genetically related to a "special facies" (Maucher and Schneider, 1967, p. 71) of the Reeves. A facies control of ore deposition is also suggested by the fact that replacement zinc-lead deposits in the mine belt occur exclusively in the Reeves Limestone Member of the Laib Formation or in its stratigraphic equivalents. And perhaps it can be argued that the facies control is a syngenetic stratigraphic control rather than an epigenetic control.

REMOBILIZATION

It seems quite clear that the present position of the ore is closely controlled by structure. It can be argued, however, that the zinc was deposited syngenetically and was later remobilized into its present position in the Reeves structure. At the Reeves MacDonald, Sinclair (1966, p. 252) recognized possibly remobilized galena in

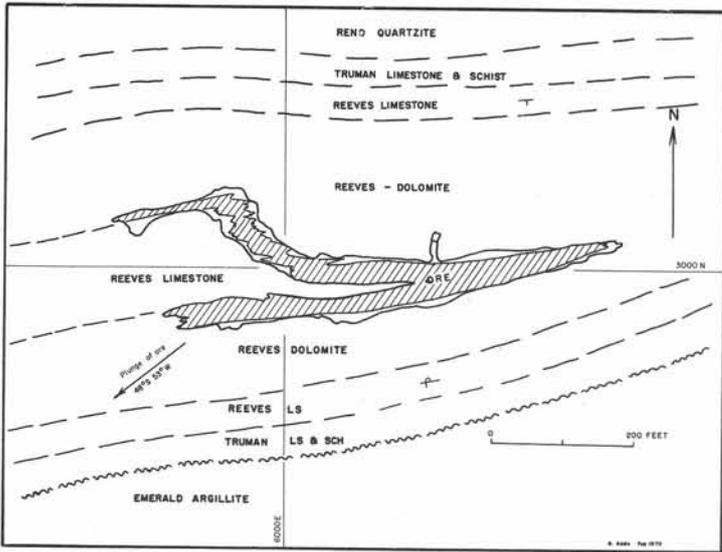


FIGURE VII-4.—Reeves ore zone—typical level.

a small vein that cuts and is younger than the massive part of the ore body. This is one of the gash veins described earlier.

If the lead were deposited from hydrothermal solutions, as suggested by the lead isotope data, then the lead was deposited after the postulated syngenetic remobilized zinc, and a second remobilization was necessary to fill the galena-bearing gash vein that cut the massive ore.

COMMENTS

The obvious stratigraphic control on lead-zinc deposits in the Canadian part of the Kootenay Arc argues for the possibility of syngenetic origin. I should like to suggest a "special facies" control, in which syngenetic pyrite is replaced by penesyngenetic sphalerite that was later mobilized to form the present ore bodies at the Reeves MacDonald mine. The lead isotope data and the close structural control indicate that hydrothermal processes may have been involved. Accordingly, it is suggested that the final stage of mineralization is an overprint by hydrothermal solutions that brought in the lead.



FIGURE VII-5.—Bands of argillite in ore zone showing thixotropic deformation. Reeves MacDonald mine.

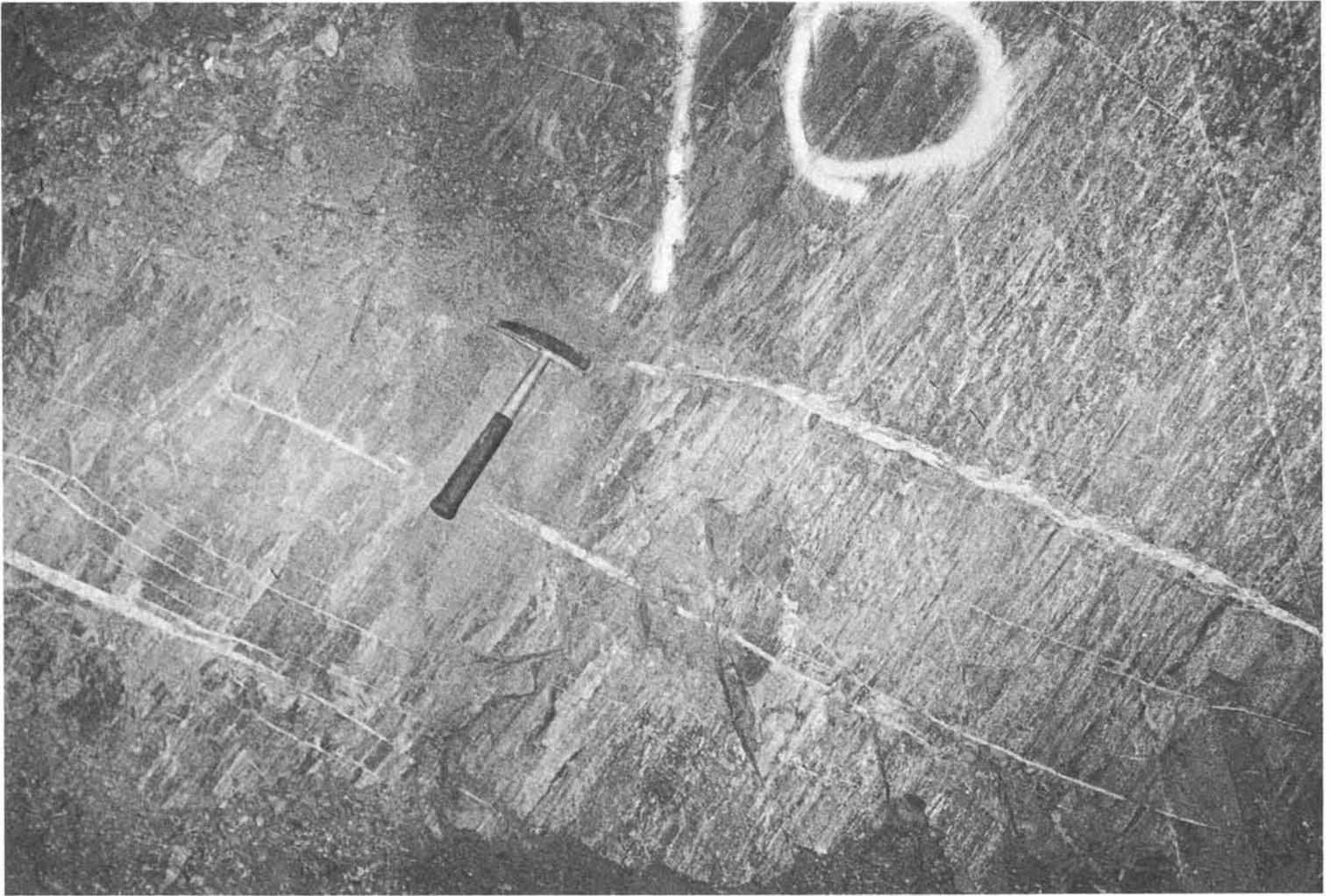


FIGURE VII-6.—Gash veins within ore body. Reeves MacDonald mine.

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(Mississippi Valley type deposits)—A symposium: *Econ. Geology Mon. 3*, p. 71-89.

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VIII

GEOLOGY OF THE JERSEY LEAD-ZINC MINE, SALMO, BRITISH COLUMBIA

By

O. E. BRADLEY

GEOLOGY OF THE JERSEY LEAD-ZINC MINE, SALMO, BRITISH COLUMBIA

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GEOLOGY OF THE JERSEY LEAD-ZINC MINE, SALMO, BRITISH COLUMBIA

By O. E. BRADLEY^①

ABSTRACT

The Jersey lead-zinc mine is located in the Nelson Mining Division of southeastern British Columbia. It lies 8 miles south of Salmo and 22 miles east of the City of Trail.

The deposit forms part of the Kootenay Arc, a north-trending belt of lime-bearing sedimentary rocks of Early Cambrian age. The Kootenay Arc is favourable to lead-zinc mineralization and includes the Reeves MacDonald, Jersey, H.B., Bluebell, and Duncan Lake mines.

The ore bodies are contained within dolomite in the Reeves Limestone, part of a member of the Laib Formation. Ore mineralization consists of fine-grained sphalerite and galena with pyrite, pyrrhotite, and minor arsenopyrite present as gangue sulphides.

The ore bodies have a trend of N. 15° E. and a plunge of 10° southerly over a distance of 6,000 feet. Maximum east-west width is 2,000 feet. The dominant structure in the mine area is the Jersey Anticline, a complex isoclinal fold, whose axis trends about N. 15° E. and plunges at low angles to the north and south. The axial plane of the Jersey Anticline dips approximately 45° E. The ore bodies lie on the normal limb of this anticline.

A series of secondary folds [open anticlines and synclines] occur on the upper limb of the Jersey Anticline. These folds, with amplitudes rarely more than 50 feet, have been used to delineate ore zones from "A Zone" on the west side to "J Zone" on the east side.

The Black Argillite Fault (a reverse fault) bounds the mine area on the east side and downthrows the Reeves Member to the east. The Dodger and Emerald granite stocks underlie the mine area.

Several post-ore transverse faults, mostly normal in movement, cut the ore bodies. Net slip on these faults decreases to the north. Movement is in the order of 15 to 40 feet.

Five ore bands, ranging in thickness from 1 to 30 feet, are recognized in the mine. These ore bands from top to bottom are: Upper Lead band; Upper Zinc band; Middle Zinc band; Lower Zinc band; and Lower Lead band.

In the "A Zone," the ore bands are very close together and in places have been mined as a unit as much as 80 feet thick. Throughout the rest of the mine these bands have been mined separately or in combinations.

The Upper Lead band is from 1 to 4 feet thick. It occurs in the "A Zone" and at the north end of the mine. The Lower Lead band, which also occurs throughout the "A Zone," is very similar but extends somewhat farther south in the mine. Both the Upper and the Lower Lead bands are sinuous in plan view and, except in the "A Zone," are rarely more than 100 feet wide in east-west dimension. The Upper, Middle, and Lower Zinc bands occur either singly or in combinations throughout the

entire mine. Lead-zinc ratios are in the order of 1 to 4. Bands of mineralization are parallel to sedimentary banding in the host rock.

Cross-cutting features exhibited by dykes associated with the Dodger Stock suggest emplacement of ore before granitic intrusion.

LOCATION

The Jersey lead-zinc mine is on Iron Mountain in the Nelson Mining Division of southeastern British Columbia. It lies 8 miles south of the village of Salmo and 22 miles east of the city of Trail (Fyles, Fig. IV-1, this guidebook).

Elevations of the ore zones range from 4,000 to 4,500 feet.

HISTORY

First records of ore shipped from the Jersey area date back to 1908, when John Waldbeser produced 426 tons of lead ore, then valued at \$7,000. In 1910, Iron Mountain Limited, a subsidiary of Pacific Coast Steel of San Francisco, was formed to operate the property. Waldbeser was retained as manager. The property was worked intermittently between 1910 and 1941 by Iron Mountain Limited. In 1941, detailed surface mapping led to the discovery of tungsten mineral within well-defined beds of skarn. This skarn lay conformably below the lead-zinc-bearing dolomite. The Canadian Government purchased the tungsten claims through Wartime Metals Corporation and operated the property for a short period during World War II as the Emerald Tungsten Project. In 1947, the original 41 mineral claims and fractions were bought by Canadian Exploration Limited. Diamond drilling, under the direction of Harold Lakes, primarily for tungsten exploration, indicated approximately 70,000 tons of good-grade lead-zinc ore. The tungsten mill was converted to a lead-zinc mill, and production from the Jersey mine began in March of 1949 at a rate of 8,000 tons per month. From March 1949 to the present time the Jersey mine has operated continuously and produced more than 7 million tons of ore with an average grade of 1.8 percent lead and 4.1 percent zinc. Current production rate is 40,000 tons per month.^②

As of January 31, 1970, the published reserves of the mine were 219,880 tons of ore at 1.0 percent lead and 3.5 percent zinc. It is planned to suspend operations when the reserves are exhausted.

^①Senior Geologist, Canadian Exploration Limited.

^②Records of the British Columbia Department of Mines and Petroleum Resources show that to the end of 1969 the Jersey mine produced 25,687,496 pounds of zinc, 9,023,605 pounds of lead, 207,086 pounds of cadmium, and 28,241 ounces of silver. (The Editors.)

REGIONAL GEOLOGY

The Jersey ore bodies lie within the Kootenay Arc (Fyles, Fig. IV-1, this guidebook), a north-trending limestone belt favourable to lead-zinc mineralization. The Kootenay Arc includes the Reeves MacDonald, Jersey, H.B., Bluebell, and Duncan Lake mines. Mineralization associated with this belt is contained in the tightly folded Reeves Limestone Member of the Laib Formation and its stratigraphic equivalent, the Badshot Formation, both of Early Cambrian age.

The part of the Kootenay Arc extending from the United States border, near Nelway, north to a point east of Ymir is known as the Mine Belt. The Mine Belt includes the Reeves MacDonald, Jersey, H.B., Jackpot, and Oxide properties. The dominant structure of the Mine Belt is the north-trending "Jersey Anticline," a complex isoclinal fold whose axis trends about N. 15° E. and plunges at low angles to the north and south. The axial plane of the Jersey Anticline dips approximately 45° E.

The Jersey ore bodies lie on the normal limb of the Jersey Anticline parallel to the banding in the sediments (Fig. VIII-1). The Jersey ore bodies have a trend of N. 15° E. and a plunge of 10° southerly over a horizontal distance of 6,000 feet. Maximum east-west width is 2,000 feet.

The Reeves Limestone is 400 to 500 feet thick in the mine area. Lead-zinc mineralization occurs mainly in a dolomite layer near the base of the Reeves that ranges from 25 to 100 feet in thickness within the mine area. Reeves limestone and dolomite range from a blue-gray banded type to a white massive variety. The chief visual distinction between the two is grain size; the dolomites are generally finer grained than the limestones.

The Truman Member of the Laib Formation conformably underlies the Reeves, and forms the mine footwall rocks. This member consists of a hard, dense, reddish-green skarn and a brown, greasy-textured argillite.

Contact relations in the mine suggest the origin of the skarn to be an alteration of argillite. The skarn is characterized by tungsten and minor molybdenum mineralization. The previously mined Dodger, Emerald, and Feeney tungsten mines (Pastoor, Fig. IX-1, this guidebook) underlie stratigraphically, and in some places, geographically, the lead-zinc ore bodies.

The mine is bounded on the east by the Black Argillite (Iron Mountain) Fault, a steeply dipping reverse fault, which down-faults younger beds on the east side (Fig. VIII-1).

The Dodger and Emerald stocks, offshoots of the Nelson Batholith, underlie the mine area.

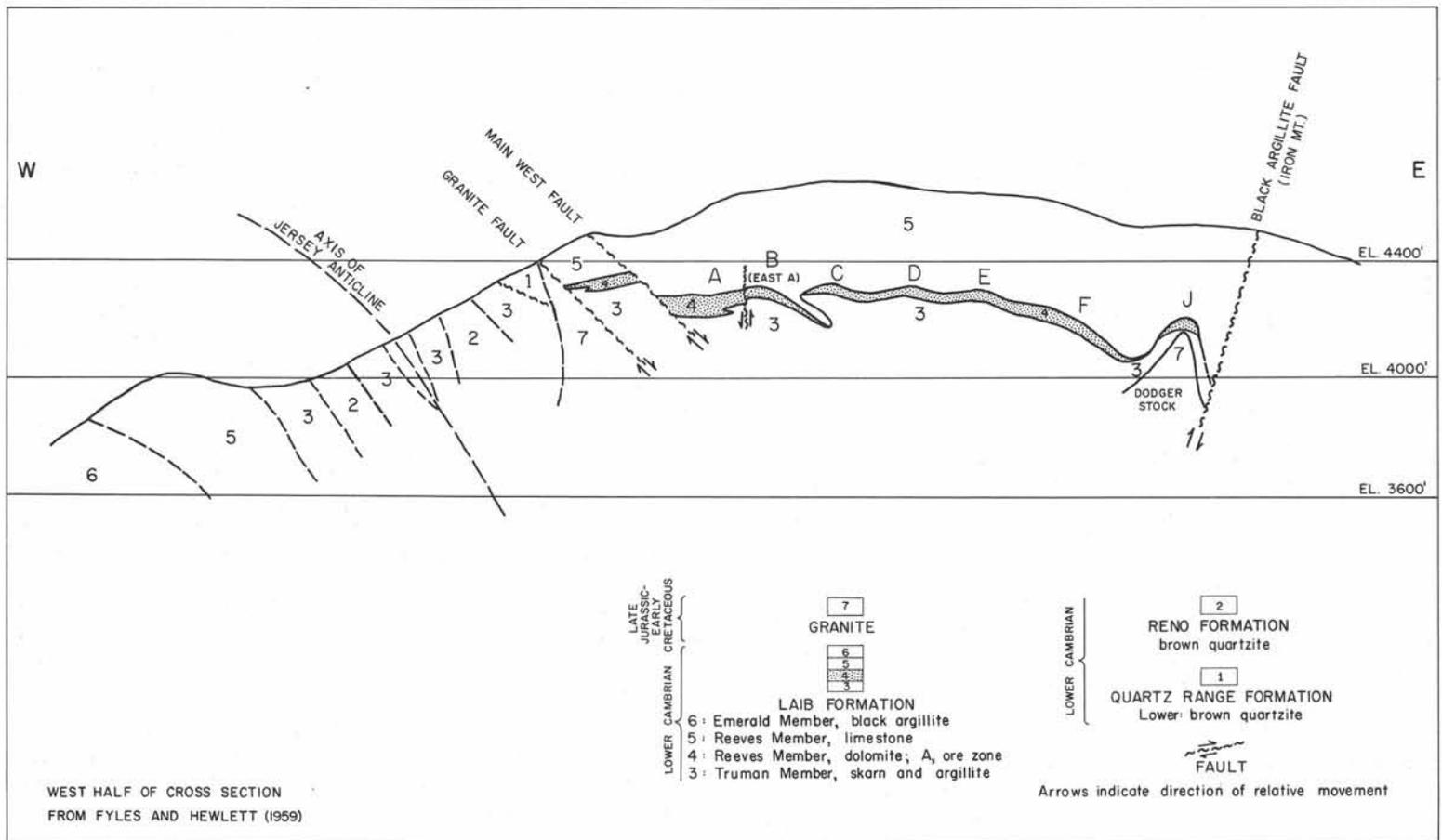


FIGURE VIII-1.—Diagrammatic cross section, Jersey mine. (West half reproduced from British Columbia Department of Mines Bulletin No. 41.)

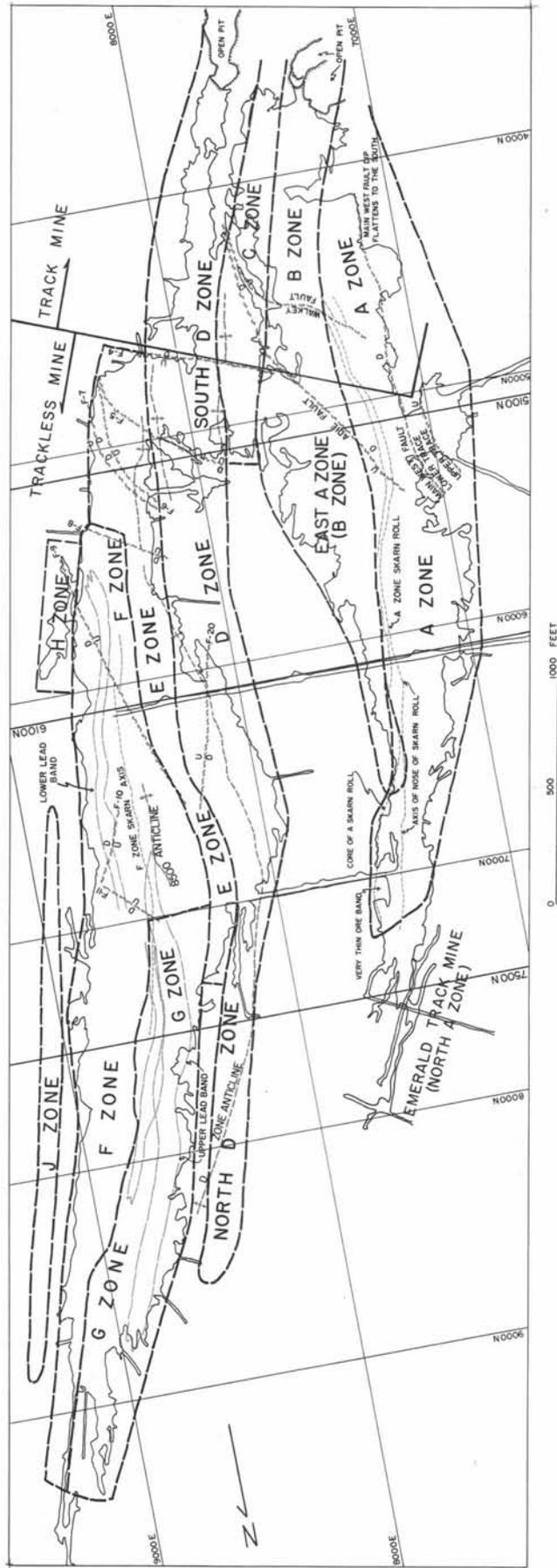


FIGURE VIII-2.—Plan of Jersey mine showing ore zones, skarn rolls, structure, and lead bands in F and G ore zones.

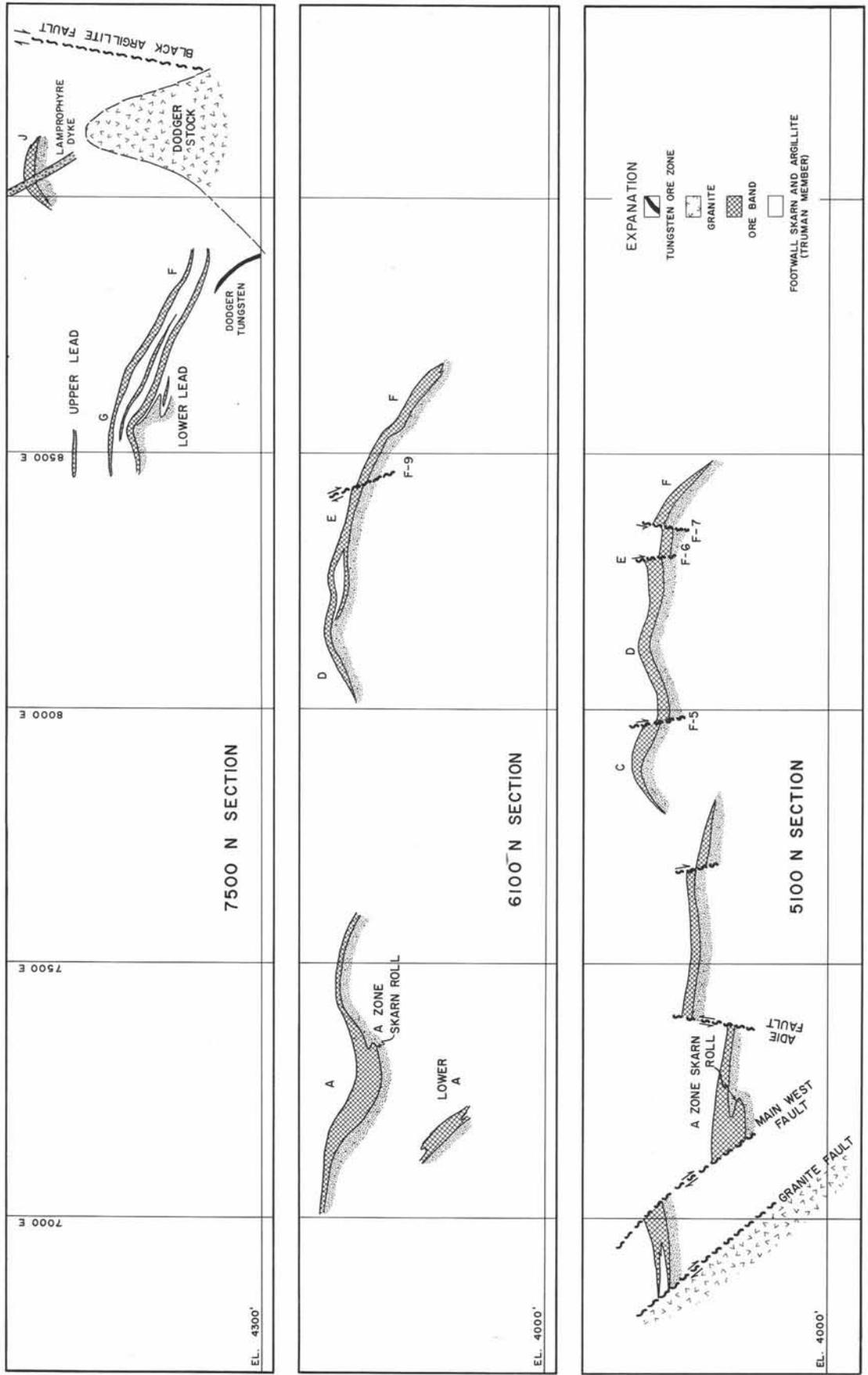


FIGURE VIII-3.—Cross sections viewed north, Jersey mine.

MINE GEOLOGY

Secondary fold structures along the normal limb of the Jersey Anticline have been used to outline ore zones. These folds are most commonly symmetrical anticlines and synclines, having amplitudes rarely more than 50 feet. The axes of these folds trend slightly east of north. Ore zones have been designated A to J from west to east (Fig. VIII-2).

The mine is cut by several transverse faults, mostly of normal movement (Figs. VIII-2 and VIII-3). Movement appears to be principally dip slip and is in the order of 15 to 40 feet. These faults are mainly rotational types, hinged on the west side of the mine. Transverse faults appear to have a common zone of origin as longitudinal faults in the southeast section of the mine and gradually swing westward in a horsetail type of structure (Fig. VIII-2). Movement on faults decreases to the north. These faults are all post ore. Faults have been numbered F-4 to F-21, increasing numbers to the north.

F-4, a near-vertical transverse fault, separates the track and trackless mines (Fig. VIII-2).

Lamprophyre dykes occur throughout the mine, especially along faults and the crests of folds (Fig. VIII-3). These dykes are younger than both the ore and the faults, and do not influence the position of ore horizons. They are occasionally useful to delineate fault zones, but can seriously dilute ore when they form the hanging wall, as they are very incompetent. The dykes range in thickness from less than 1 inch to 20 feet.

Ore mineralization consists of fine-grained sphalerite and galena, with pyrite, pyrrhotite, and minor arsenopyrite as gangue sulphides. Cadmium is present in the sphalerite, and silver accompanies the galena. Iron content of the sphalerite is approximately 6 percent. The grade of the probable reserve of the mine is 1.0 percent lead and 3.5 percent zinc.

Five ore bands, ranging in thickness from 1 to 30 feet, are recognized in the mine. These ore bands from top to bottom are: Upper Lead band; Upper Zinc band; Middle Zinc band; Lower Zinc band; and Lower Lead band. Any ore band containing more lead than zinc has been designated as a lead band.

A ZONE

The A Zone is the westernmost ore body in the mine, the north end of which was originally mined as the Emerald track mine (Fig. VIII-2). This zone crops out on the west slope of Iron Mountain. The A Zone is divided into three distinct areas, the west, central, and east areas.

The west area is a steep, east-dipping, two-band structure. A thin lead band (1 to 4 feet thick) underlies a zinc band of similar thickness. These bands are separated by 5 to 10 feet of gray-white dolomite. The Main West Fault, shown on Figures VIII-1, VIII-2, and VIII-3, is a north-trending structure dipping 20° to 35° E., and has a normal throw of about 125 feet. Approximately 70,000 tons of ore has been mined from west of this fault.

Farther west, the Granite Fault, shown on Figures VIII-1 and VIII-3, has brought intrusive rocks up on the west side but does not interfere with any known ore horizons.

The central area of the A Zone has a combined ore thickness approaching 80 feet. A unique feature of the central A Zone is the A Zone skarn roll, a recumbent isoclinal fold open to the west at the base of the Reeves. This fold contains high-grade ore that is particularly high in lead. Footwall rocks of the Truman Member form an envelope around the lead- and zinc-bearing Reeves dolomite. The west limit shown on Figure VIII-2 is the trace of the nose of the fold, while the east limit shown represents the core of the fold. The axis of the fold trends north, paralleling the A Zone. In the south part of the mine the ore contained in the skarn roll is thick, but has limited east-west dimension. Proceeding north, the situation reverses.

B ZONE

The east area, or B Zone, is principally a zinc horizon, ranging in thickness from 1 to 10 feet and dipping 0° to 20° E. This zone is very pyritic, and occasionally the lead-zinc ore is masked by massive pyrite. Samples of this pyritic ore show a grade of 1.4 percent lead and 3.8 percent zinc. The quantity of iron (20 to 30 percent), however, significantly lowers the amount of sphalerite and galena recoverable. The A Zone east area dips below the C zone, and has been traced north and east to a point 70 feet below the west side of the D Zone at coordinate 7000 N. (Fig. VIII-2).

Total production from the A Zone from 1949 to 1970 was approximately 2,225,000 tons of 2.9 percent lead and 2.8 percent zinc.

C ZONE

The C Zone lies immediately east of the A Zone and is delimited by a north-trending anticline (Fig. VIII-2). The crest of this anticline is occupied and paralleled by a large lamprophyre dyke. Ore in the C Zone is generally thin on the crest of the fold and thickens on the flanks. This zone does not persist to the north.

D AND E ZONES

The D and E Zones occupy the central part of the Jersey mine (Fig. VIII-2). Two anticlines of small amplitude east of the C Zone trend north and delimit these zones (Fig. VIII-2). Ore mineralization tends to parallel the fold axis and is seldom more than 20 feet thick. Faults F-5, F-6, and F-7 displace ore, mainly vertically, in the order of 20 to 30 feet (Figs. VIII-2 and VIII-3). Oxide associated with F-5 directly overlies the D Zone and ranges between 5 and 20 feet in thickness. Both zones are predominantly zinc bearing. In most areas the ore has been mined as a single unit, but in places three distinct bands of mineralization have been recognized and mined as such where sill thicknesses have permitted. North of 4470 crosscut (Fig. VIII-2), D Zone ore ceases to parallel the west anticline and trends northeastward across the structure at a low angle. Mineralization fades out in a coarsely crystalline limestone in the west flank of the D Zone anticline.

Ore in the D Zone is underlain by Truman skarn and argillite (Fig. VIII-3). The argillite is dark brown and locally exhibits a boudinage structure.

In a few isolated areas the argillite forms what has been termed a "dessication breccia," rectangular fragments of brown argillite in a black limy matrix.

F ZONE

The F Zone lies on the east, moderately steeply dipping, flank of the trackless mine. All five ore bands are represented in this zone. Usually two or more zinc bands have been mined as a unit in this area. A monoclinical roll of footwall rocks trends north as shown in Figure VIII-2, abruptly cutting off the lower zinc band. An upper lead and lower lead band, shown on Figures VIII-2 and VIII-3, occur above and below the zinc horizons. The trends of these bands are closely parallel. Combined grades from the upper and lower lead bands are similar, but the lead-zinc ratios differ, i.e.:

Upper..... 3.3 percent lead, 3.0 percent zinc
Lower..... 4.0 percent lead, 2.3 percent zinc.

A unique feature of the F Zone is the skarn that encloses the lower lead band, roughly resembling the A Zone skarn roll, but open to the east. Zinc ore in the east part of the F Zone is locally overlain by massive arsenopyrite.

G ZONE

The G Zone in the mine is a misnomer and actually is the north extension of the E Zone. Mineralization in the G Zone has a greater vertical extent than that encountered in the E Zone and, as in the F Zone, all five ore bands are present. Typical G Zone lead ore is associated with boudins of pyrite. A large low-grade section of zinc ore occurs at the north end of this zone, overlain by oxidized remnants of the upper zinc and lead horizons.

H AND J ZONES

In the steeply dipping east part of the mine it is not uncommon to have a flattening of ore banding and, in some instances, a dip reversal.

The H Zone (Fig. VIII-2), a two-band zinc structure, lies within this flattened area, referred to as the Dodger trough.

H Zone ore is complicated by sills and dykes from the Dodger Stock.

The J Zone, a recent addition to the mine, is a double-band, tilted anticlinal structure occurring above the top of the Dodger Stock (Figs. VIII-2 and VIII-3). Diamond drilling has indicated a narrow north-trending zone of mineralization extending for a distance of 1,600 feet that closely parallels the trend and plunge of the other ore bodies in the mine. Ore grade is approximately 1.0 percent lead and 2.5 percent zinc. J Zone mineralization is probably directly related to the east part of the F Zone and has been pushed up about 300 feet by the emplacement of the Dodger Stock. Steep dips and occasional lead-zinc mineralization encountered between the F and J

Zones in diamond drilling tentatively confirm this correlation theory. Mineralization in the J Zone is underlain by Truman footwall rocks.

MINERALIZATION CONTROL

Dr. J. T. Fyles, of the British Columbia Department of Mines and Petroleum Resources, suggested that the dolomitization of the Reeves Limestone is structurally controlled by secondary folding. Ore mineralization is more abundant in the troughs of the secondary folds than on the crests. Open textures developed in the dolomite may have been favourable to lead-zinc mineralization. Granitic intrusions followed dolomitization and mineralization, as evidenced by post-ore dykes in the north and west areas of the mine. The range of possible ages of mineralization is wide. It may be said that it approximates or post-dates secondary folding and pre-dates granitic intrusions.

CONCLUSION

Although the grade of the probable reserve of the Jersey trackless mine is close to 5 percent combined lead and zinc, the mining grade is less. This is due to wide distribution of low-tonnage probable ore blocks that require the addition of much of the marginal reserve in order to maintain an economic production tonnage.

The gentle plunge and low dips of the ore are particularly amenable to trackless mining.

ACKNOWLEDGMENTS

The west half of Fig. VIII-1, "Diagrammatic cross-section of the Jersey mine," was reproduced from British Columbia Department of Mines Bulletin No. 41 with the permission of Dr. J. T. Fyles, of the British Columbia Department of Mines and Petroleum Resources.

I wish to acknowledge the most welcome assistance of Mr. Lyle Andrews, of Canex Aerial Exploration, in the preparation of the accompanying illustrations.

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IX

**GEOLOGY OF THE INVINCIBLE TUNGSTEN ORE ZONE OF
CANADIAN EXPLORATION LIMITED, SALMO, BRITISH COLUMBIA**

By

D. W. PASTOOR

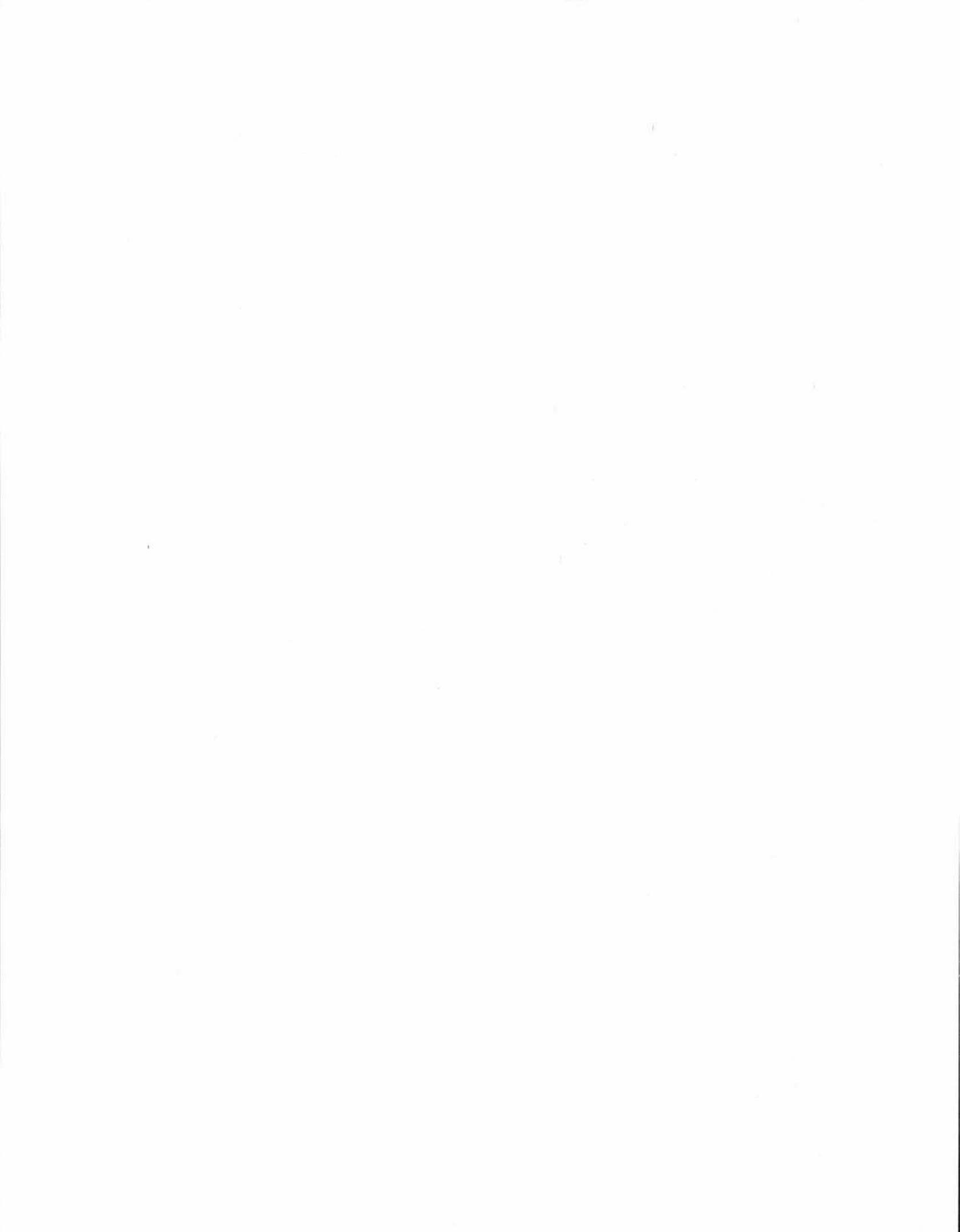
GEOLOGY OF THE INVINCIBLE TUNGSTEN ORE ZONE OF
CANADIAN EXPLORATION LIMITED, SALMO, BRITISH COLUMBIA

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GEOLOGY OF THE INVINCIBLE TUNGSTEN ORE ZONE^① OF CANADIAN EXPLORATION LIMITED, SALMO, BRITISH COLUMBIA

By D. W. PASTOOR^②

LOCATION

The Invincible tungsten ore zone (Fig. IX-1) is 3,100 feet northeast of the old Emerald tungsten mine and 250 feet northwest of the north end of the producing Jersey lead-zinc mine, on Iron Mountain in the Nelson Mining Division of southeastern British Columbia (Fyles, Fig. IV-1, this guidebook).

The Invincible ore zone ranges in elevation from 3,320 feet to 3,500 feet.

HISTORY

In 1941, tungsten was discovered on the property of Iron Mountain Limited, which company had operated the property as a lead-zinc mine intermittently between 1910 and 1941. Because of the strategic use of tungsten, the

^①Although the Field Conference is concerned primarily with the zinc-lead deposits, this brief description of the Invincible tungsten ore zone is included because of its proximity to the zinc-lead deposits of the Jersey mine, and its possible genetic relation to these deposits (The Editors).

^②Mine Geologist.

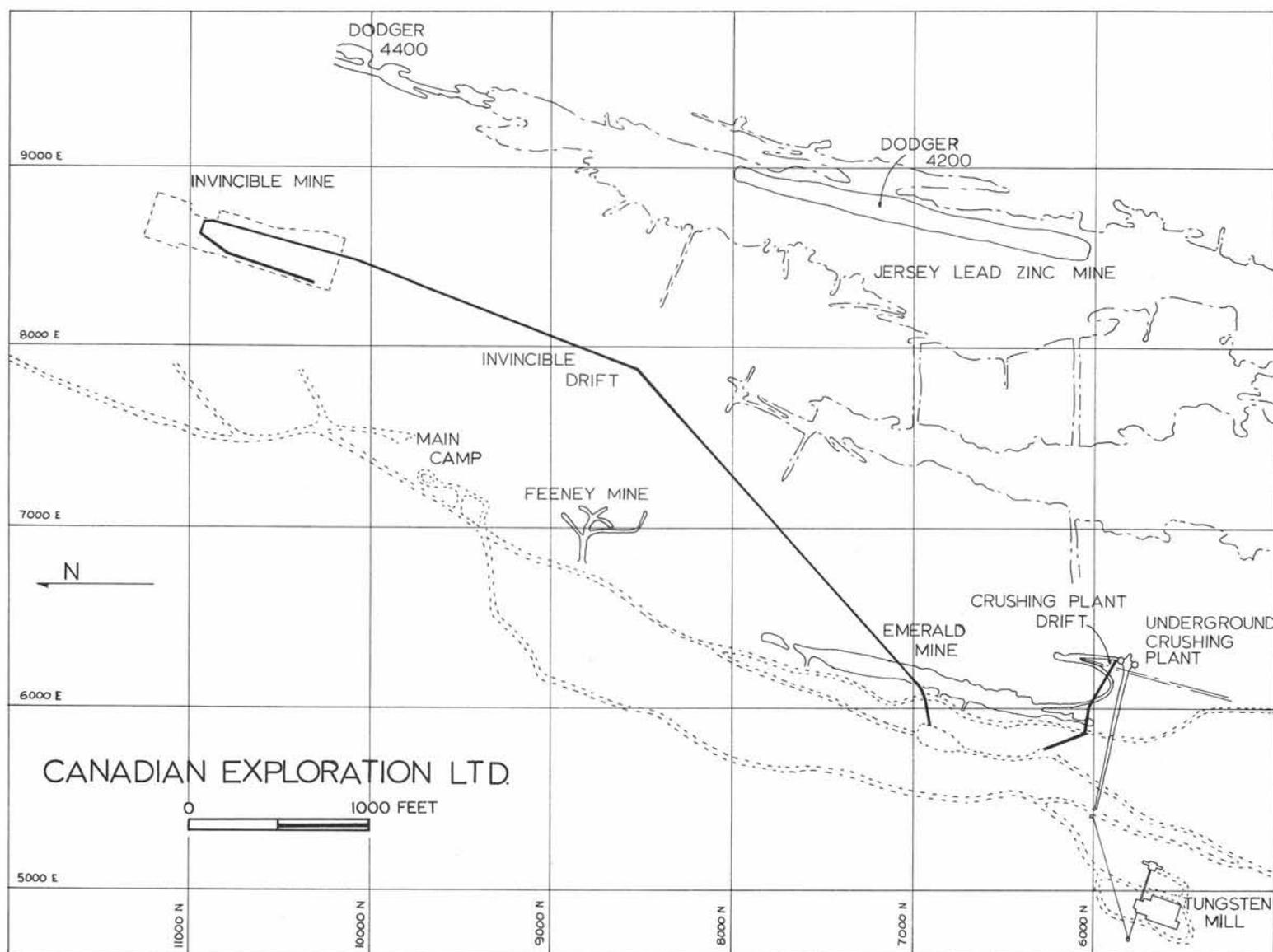


FIGURE IX-1.—Location map of mines on company property.

property was purchased by the Canadian Government through Wartime Metals Corporation and operated by them as the Emerald Tungsten Project. In the summer of 1943, the mine and mill were brought into production and operated for six weeks, then the tungsten mine was closed.

Early in 1947, Canadian Exploration Limited purchased the property and undertook a diamond-drilling program for tungsten, which proved 60,000 tons of good-grade lead-zinc ore. The tungsten mine was phased out of production in the latter part of 1948 and the mill converted to lead-zinc recovery. Lead-zinc production started in March 1949.

In the early part of 1951, the Canadian Government purchased two blocks of ground from Canadian Exploration Limited, covering the Emerald tungsten ore and the partly developed Dodger tungsten ore zone, on the east of the Jersey lead-zinc mine (Fig. IX-1).

In September 1952, Canadian Exploration bought back the ore blocks, and operated the Emerald tungsten mine until July 31, 1958.

An extensive surface-mapping program outlined another tungsten ore zone approximately 300 feet northeast of the Emerald mine, and this became known as the Feeney mine (Fig. IX-1). This mine operated from 1951 to 1955 and produced 60,000 tons of tungsten ore, grading 0.92 percent tungstic oxide.

Underground exploration in the Dodger area proved up two tungsten ore zones. These became the Dodger 4400 and the Dodger 4200 mines (Fig. IX-1).

The Dodger 4400 was operated from 1952 to 1957 and produced 137,000 tons of tungsten ore, with a grade of 0.56 percent tungstic oxide. The Dodger 4200, operated from 1954 to 1957, produced a total of 158,000 tons of ore, grading 0.60 percent tungstic oxide.

Surface mapping had also shown that there was a possibility of a trough structure, similar to the Emerald and Feeney mine troughs, north of the Feeney mine. A surface diamond-drill program was initiated on February 25, 1953. This program was suspended in April of that year without reaching the projected ore horizon, because

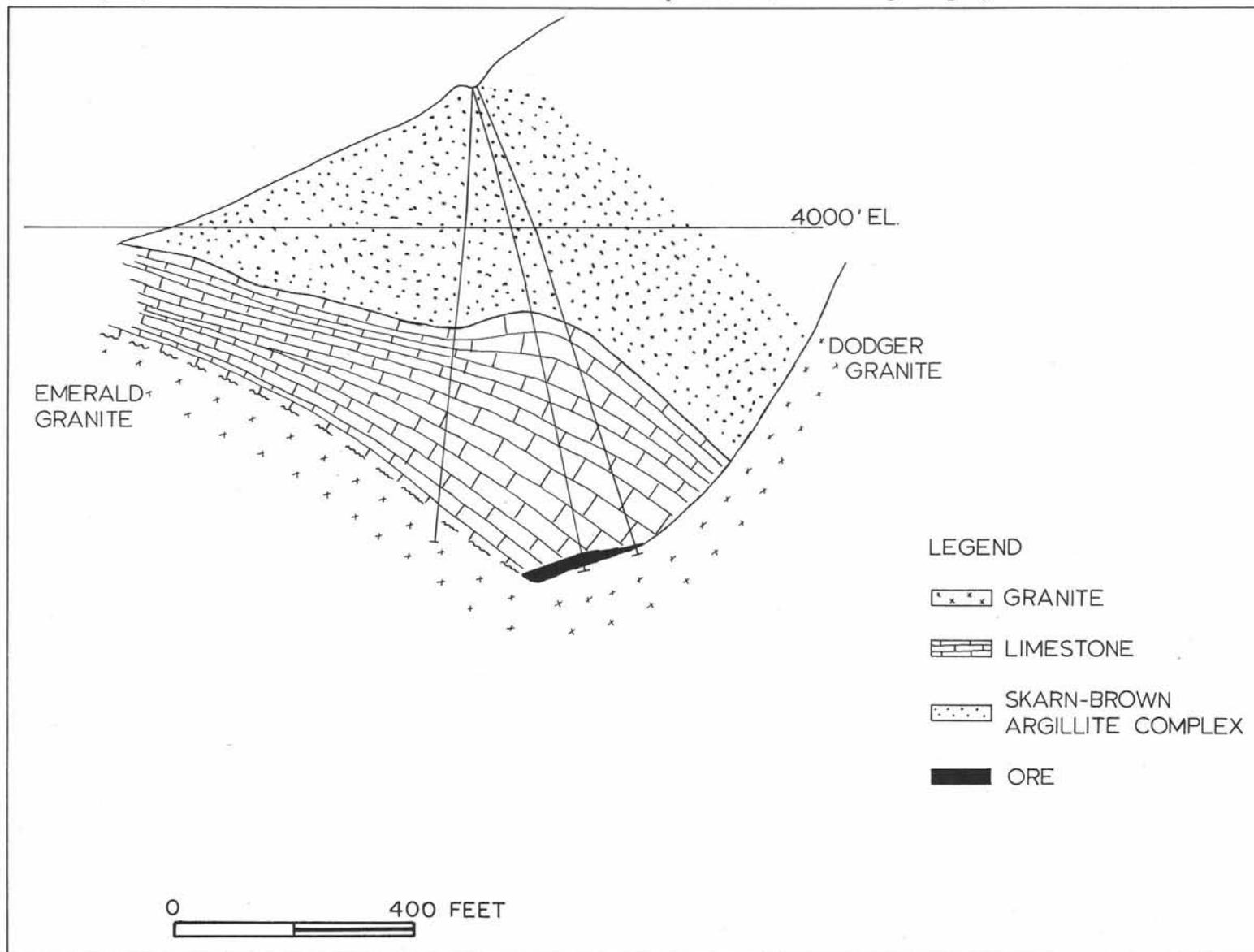


FIGURE IX-2.—Cross section south zone.

of highly fractured and vuggy ground that made drilling extremely difficult.

A second drill program was started in August 1954 at the same location as the initial attempts. The original hole was drilled to completion and intersected 10 feet of tungsten mineralization of 0.75 percent tungstic oxide ore. By the summer of 1956 enough ore had been outlined to warrant feasibility studies. That winter it was decided to bring the Invincible ore zone into production, and a shaft site was selected. The shaft pilot hole was collared in February 1957, and on March 7 the hole intersected 26 feet of tungsten ore, so a new shaft site was selected. The spring of 1957 saw the price of tungsten concentrate drop drastically, and the project was postponed.

In 1969, after a total of 34,462 feet of diamond drilling, another feasibility study was made, and late that fall it was again decided to bring the property into production. The Invincible ore zone will be developed by driving a 6,000-foot-long 16-foot by 19-foot drift to facilitate the use of trackless equipment. The ore will be trucked directly

to the internal crushers through another 650-foot-long drift (Fig. IX-1). The internal crushers are presently used for the ore mined in the Jersey lead-zinc mine, which will be closed when the tungsten mine is producing.

GEOLOGY

The Invincible ore zone is divided into two zones, separated by a 650-foot-long area containing granite cross-dykes.

The south zone is approximately 800 feet long and is lying nearly flat in a troughlike structure formed by westward-dipping Dodger granite and eastward-dipping Emerald granite (Fig. IX-2). The intrusive Dodger granite contact dips approximately 45° west. The Emerald granite underlies a fault surface and dips approximately 35° east. The trough is terminated at both ends by areas of "high-granite" or cross-dykes.

The north zone is approximately 1,100 feet long and has a gentle plunge to the south. The trough structure is

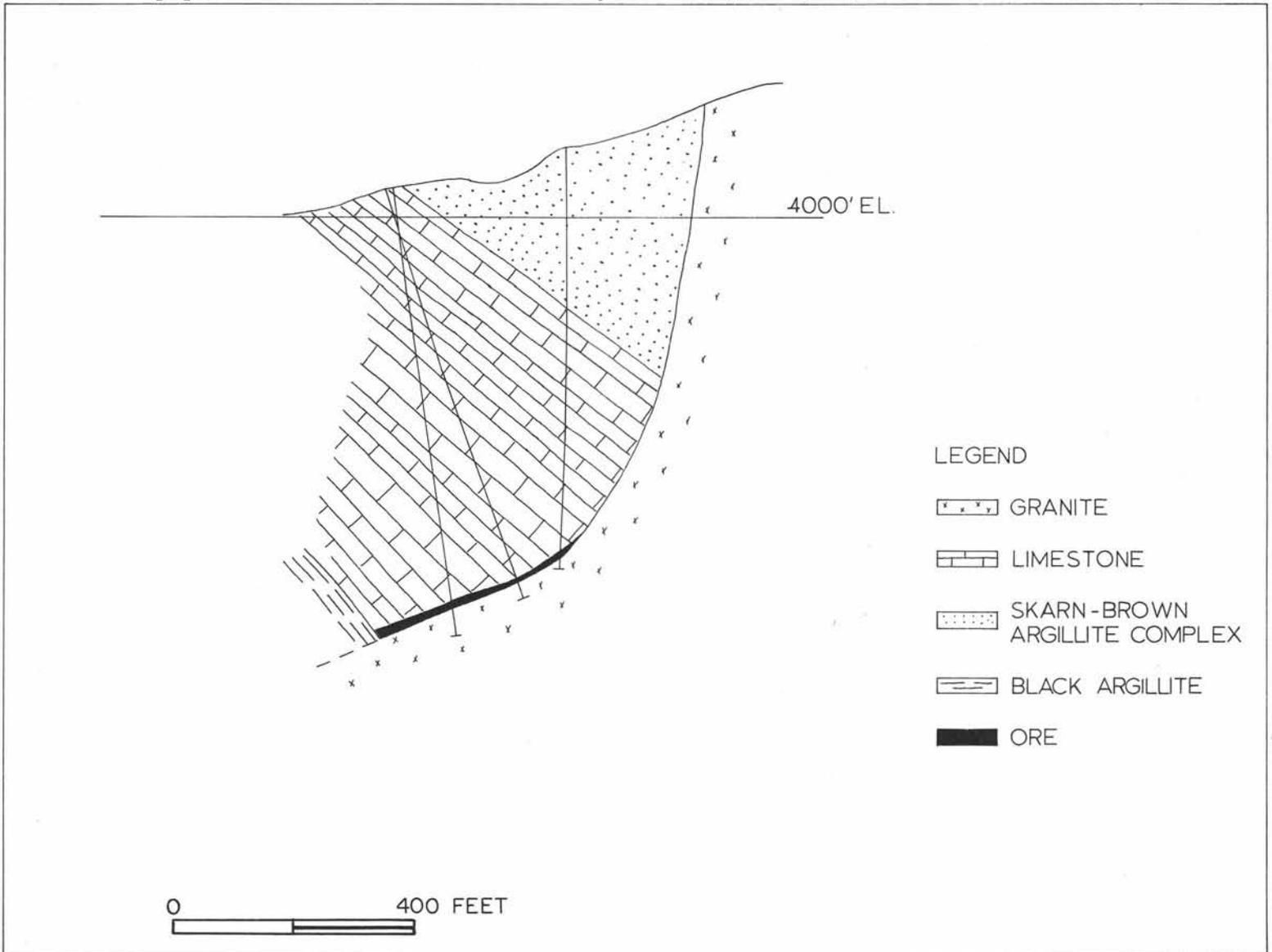


FIGURE IX-3.—Cross section north zone.

formed by westward-dipping Dodger granite and the eastward-dipping contact between Emerald black argillite and limestone (Fig. IX-3). The black argillite-limestone contact dips about 40° east, and the granite dips about 35° west.

In both zones, the tungsten ore occurs where the limestone is in contact with the Dodger granite. The ore is contained in quartz-sulphide replacement of the limestone. It is localized in general by fracturing, faulting,

and brecciation in the limestone, and in detail by minor fracturing and the bedding of the limestone.

To date, 278,000 tons grading 0.78 percent tungstic oxide has been outlined by surface diamond drilling.

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X

**SOME UNIQUE GEOLOGICAL FEATURES AT THE BLUEBELL MINE,
RIONDEL, BRITISH COLUMBIA**

By

FRANK G. SHANNON

SOME UNIQUE GEOLOGICAL FEATURES AT THE BLUEBELL MINE,
RIONDEL, BRITISH COLUMBIA

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SOME UNIQUE GEOLOGICAL FEATURES AT THE BLUEBELL MINE, RIONDEL, BRITISH COLUMBIA

By FRANK G. SHANNON^①

LOCATION AND HISTORY OF PRODUCTION

The Bluebell mine is on the east shore of Kootenay Lake, 50 miles north of the Canada-United States border, and 100 miles west of the British Columbia-Alberta border (Fyles, Fig. IV-1, this guidebook). The nearest city is Nelson, 30 miles to the west. Kootenay Lake is about 75 miles long and about 2 miles wide. Bluebell mine is almost directly across the lake from Ainsworth, an early, now almost dormant, silver-lead-zinc mining camp. Both the Ainsworth and Bluebell areas were involved in the earliest mining recorded in British Columbia in 1882, but were well known prior to 1882 to the early fur traders, partly because of the outcrops on the shorelines of the lake. Production was intermittent from 1895 to 1927 under various owners. From 1952 to the present time the mine has been under the management of Cominco, Ltd.

Production has been:

Ore		Pb %/ton	Zn %/ton
By the pre-Cominco owners	550,000 tons	6.5	8.2
By Cominco	4,250,000 tons	5.2	6.2

Other products are: silver, about 1 oz. per ton, copper about 0.10 percent per ton, and cadmium 0.025 percent per ton.

Descriptions of the Bluebell mine have been published previously (Irvine, 1957), and the main features of the deposit described in this paper are taken from that publication. Detailed descriptions are based on observations made by the writer in recent years. Westervelt (1960) added considerably to our knowledge of the mineralization and described the occurrence of knebelite. Ohmoto and Rye (1970) have increased our knowledge with their studies, which are continuing, on fluid inclusions in crystals and also on the isotopes of hydrogen, oxygen, and carbon.

The Bluebell mine has four areas of geological interest that are unique. The information about them has been developed over a long period of time and, because they are abnormal, they may be of use to geologists elsewhere. These areas of geological interest are:

- (1) Mineralization and mode of occurrence
- (2) Structural control of mineral distribution
- (3) Oxidation at depth under a sulphide cap
- (4) Thermal springs with heavy emission of CO₂ gas.

REGIONAL GEOLOGY AND MINE STRATIGRAPHIC SECTION

The ore outcrops are conspicuous on a peninsula about 1 mile long by half a mile wide on the east shore of

Kootenay Lake. Metasedimentary rocks of sillimanite grade of regional metamorphism dip 35° westward under the lake (Fig. X-1a). The Bluebell Limestone, which contains the ore bodies, has recently been correlated with the Badshot Limestone, which is thought to be Early Cambrian (see Fyles, Table IV-1, this guidebook).

STRATIGRAPHIC SECTIONS NEAR THE MINE

Starting at the lakeshore and proceeding eastward from hanging wall through footwall, the stratigraphic section consists of:

400 FEET OF HAMILL QUARTZITES

This top member adjacent to the lake is the hanging wall for the mine. White quartzite bands are interlayered with brown platy mica schist and some quartz feldspar pegmatite bands, lenses, and deformed blobs indicating shearing movement. Quartz augen and garnet porphyroblasts partly replaced by chlorite are common. Tourmaline occurs rarely. There is one lens of limestone (locally called upper limestone) 80 feet above the Bluebell Limestone in the Hamill Quartzites. The maximum thickness is 18 feet, maximum strike length 3,500 feet, and maximum dip length below outcrop at surface is 800 feet. This limestone may be correlated with limestone in the Mohican Formation in the Lardeau district (Fyles, 1964, p. 22).

100 TO 150 FEET OF BLUEBELL LIMESTONE

Mostly white, alternating fine- and coarse-grained crystalline limestone with some grey bands, some phlogopite partings, and disseminated biotite and graphite in the lower part. Some of the coarse-grained grey beds are dolomitic. No fossils have been noted. This limestone is tentatively correlated with the Badshot Limestone (Crosby, 1968; Fyles, this guidebook).

700 FEET OF LOWER INDEX FORMATION—EARLY PALEOZOIC

Graphitic grey and black schistose argillite or argillaceous quartzite forms the footwall for the mine. The beds are partly calcareous, contain fine pyrite, and are somewhat feldspathic. The formation shows a great deal of shearing on planes parallel or subparallel to the bedding. They locally contain lenses of impure limestone.

500 FEET OF INDEX FORMATION

Hornblende schist with some interlayered quartz mica schist and limy schist succeed the lower Index Formation.

1,600+ FEET OF INDEX FORMATION

Quartz-calc-silicate schist, feldspathic and calcareous in some places, with amphibolite and coarse-grained white pegmatite sills, in places containing lenses of impure limestone, form a thick sequence east of the mine.

^①Mine geologist, Cominco Ltd., Bluebell mine, Riondel, British Columbia.

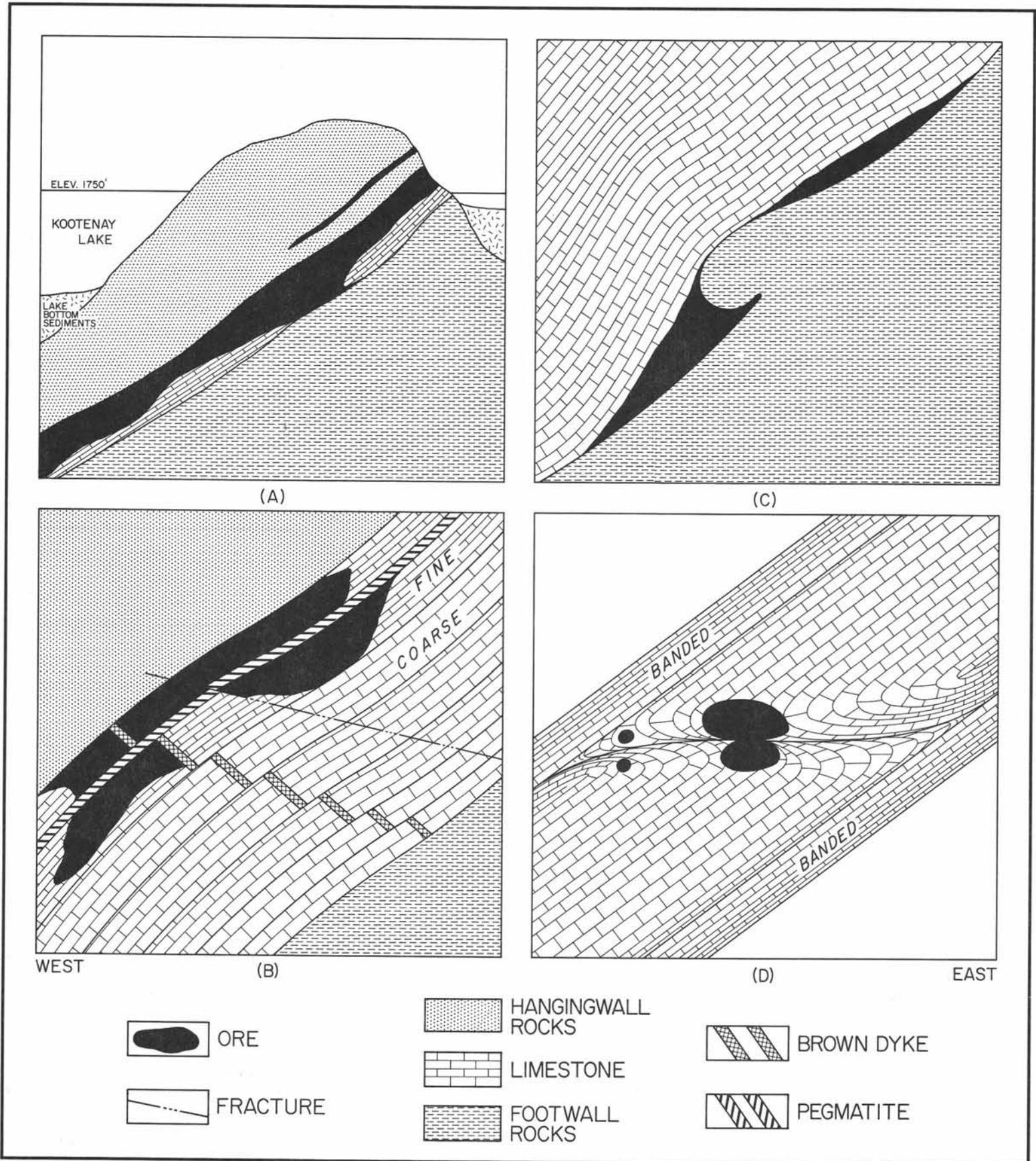


FIGURE X-1.—Diagrammatic cross-sections showing:
 A. A typical structural section in the Kootenay Chief Zone.
 B. Displacement of a brown dyke and the occurrence of ore.
 C. and D. Typical locations of ore bodies.

The above formations, to the casual observer, would seem to be right side up. Projections of the structure from 30 miles to the north near Duncan Lake, however, indicate them to be on the overturned limb of an isoclinal fold. All the rocks strike about N. 5° E. and dip about 35° W.

GENERAL MINE GEOLOGY

MINE AREA INTRUSIVES

"BROWN DYKES"

These earliest intrusives are pre-regional folding. They are fine grained with a brownish-pink groundmass containing some mica and pyroxene, which, upon alteration to chlorite, becomes green. There are very few of these dykes, but they have been noted in widely scattered parts of the mine. They all strike roughly north-south and dip 30° to 60° to the east. This type of dyke has been found only in the limestone.

PEGMATITE SILLS

These were emplaced before or during regional deformation, but later than the brown dykes noted above. They are fine- to coarse-grained, white, quartz-feldspar rocks, which include granodiorite, aplite, and granite pegmatite. Strong foliations are developed in some places, with the same metamorphic minerals as in the surrounding rocks (Crosby, 1968). The dykes are found in all rocks near the Bluebell mine, and vary in thickness from several inches to several hundreds of feet. It is thought they are pre-Cretaceous in age.

"GREENSTONE DYKES"

Numerous dark green-grey dykes, up to 5 feet thick and near andesite in composition, are present in all rocks near the mine. They contain phenocrysts of plagioclase (labradorite to andesine), olivine, pyroxene, hornblende, and abundant biotite, which are often replaced by calcite, epidote, chlorite, and magnetite. The dykes commonly have a bleached margin about 2 inches wide where they are in contact with limestone. They are the youngest intrusives, have not been folded or affected by regional metamorphism, and are pre-ore. The average strike is nearly east-west, and the dip is steep either to the north or to the south. Probably these dykes, which are commonly grouped with lamprophyres occurring widely in the Kootenay Arc, are Cenozoic in age.

MINERALIZATION AND ITS MODE OF OCCURRENCE

Ore, because it cuts the lamprophyre dykes, appears to be Cenozoic. The primary ore minerals, galena and black ferriferous sphalerite (marmatite) occur as coarsely crystalline aggregates, and are found both along cross fractures and as bedded replacements. Silver in small amounts is associated with galena, but no silver minerals have been identified. Pyrrhotite is the most abundant and probably the earliest sulphide mineral. Occasionally it is found in large separate masses in the larger ore bodies, or, more often, intimately associated with the galena and sphalerite. Pyrite is almost exclusively a secondary min-

eral, but minor primary pyrite occurs. Small amounts of arsenopyrite, chalcopyrite, siderite, and rhodochrosite are present. Considerable crystalline pyrrhotite, marmatite, arsenopyrite, minor chalcopyrite, and galena occur with quartz and calcite crystals in vuggy openings.

The gangue consists of limestone with considerable pegmatite, coarsely crystalline quartz, and carbonates. Wall rock alteration is negligible. Knebelite, an iron manganese silicate (olivine group) is found in abundance in cross fractures and in bedding deposits closely associated with them, and is considered to be the earliest replacement mineral (Westervelt, 1960). Some light-brown to yellow sphalerite is found in some places where marmatite is intimately associated with knebelite; also fine-grained magnetite is present where galena is associated with knebelite. Minnesotaite (Irvine, 1957) and dickite (Ohmoto and Rye, 1970) are found in limited quantities as alterations of knebelite.

Quartz lines vuggy openings, mostly close to the mineralizing fractures, and is almost never found in the knebelite-rich areas. Quartz crystals of the latest period of deposition usually contain numerous carbonate and sulphide inclusions and also CO₂ gas and liquid inclusions along the growth zones (Ohmoto and Rye, 1970).

DESCRIPTION OF ORE BODIES

Irvine (1957, p. 98, 100) described the ore bodies as follows:

There are three known centres of mineralization in the mine, spaced at approximately 1,500-foot intervals along the strike of the Bluebell limestone.

These are called the Comfort, Bluebell, and Kootenay Chief zones (see Figs. X-2 and X-3).

In these mineralized centres the lead-zinc ore bodies, which occur as massive sulphide replacements, are localized along steep cross fractures, which extend across the limestone from hanging wall to footwall. The cross fractures themselves are only a portion of an inch in width, the ore shoots being formed by sulphide replacements along beds adjoining the fractures. Replacement of this sort proceeds in irregular fashion for 5 to 10 feet from the fractures, then cuts out abruptly. The shoots thus form tabular bodies, transverse to the bedding and having irregular outlines due to variations in the extent to which the replacement has proceeded along various beds cut by the cross fractures. In places the control of the mineralization will shift from one cross fracture to an adjoining one, or there may be several adjoining cross fractures, each with sulphide mineralization. Where mineralized fractures are closely bunched in this way the ore may coalesce into larger bodies, 30 to 40 feet, or in exceptional cases as much as 100 feet, in width,

and may extend from hanging wall to footwall.

Where beds particularly favourable to sulphide deposition occur, the ore may spread out from the mineralized fractures for as much as 100 feet along the bedding planes. The most consistently favourable beds for replacement of this type are the dense, closely banded limestone beds in the upper part of the Bluebell Limestone, and the ore occurs either just under the hanging wall quartzite or just under the pegmatite sills which lie a few feet stratigraphically below the base of the quartzite.

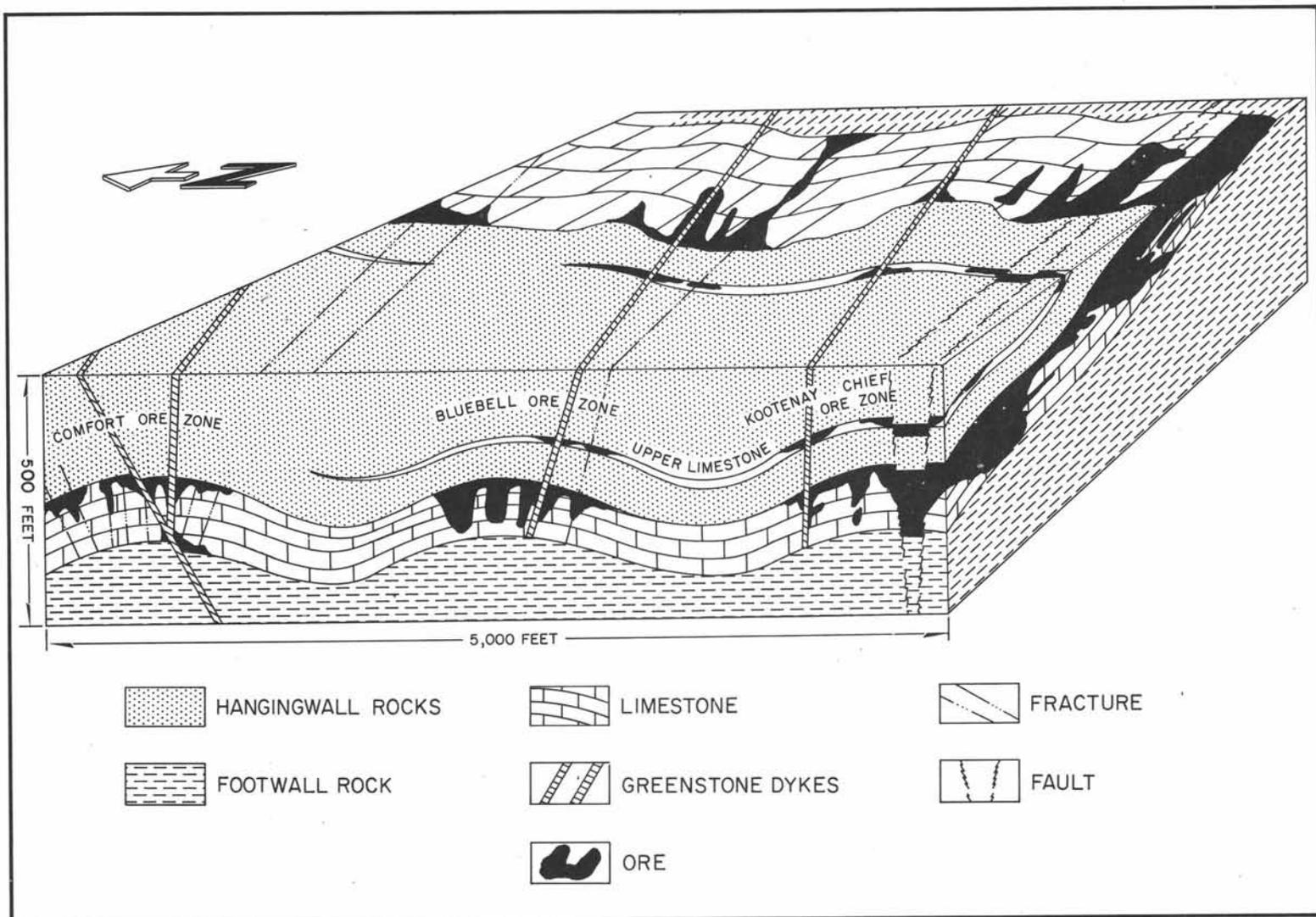


FIGURE X-2.—Block diagram of the Bluebell mine area showing mode of occurrence of the ore.

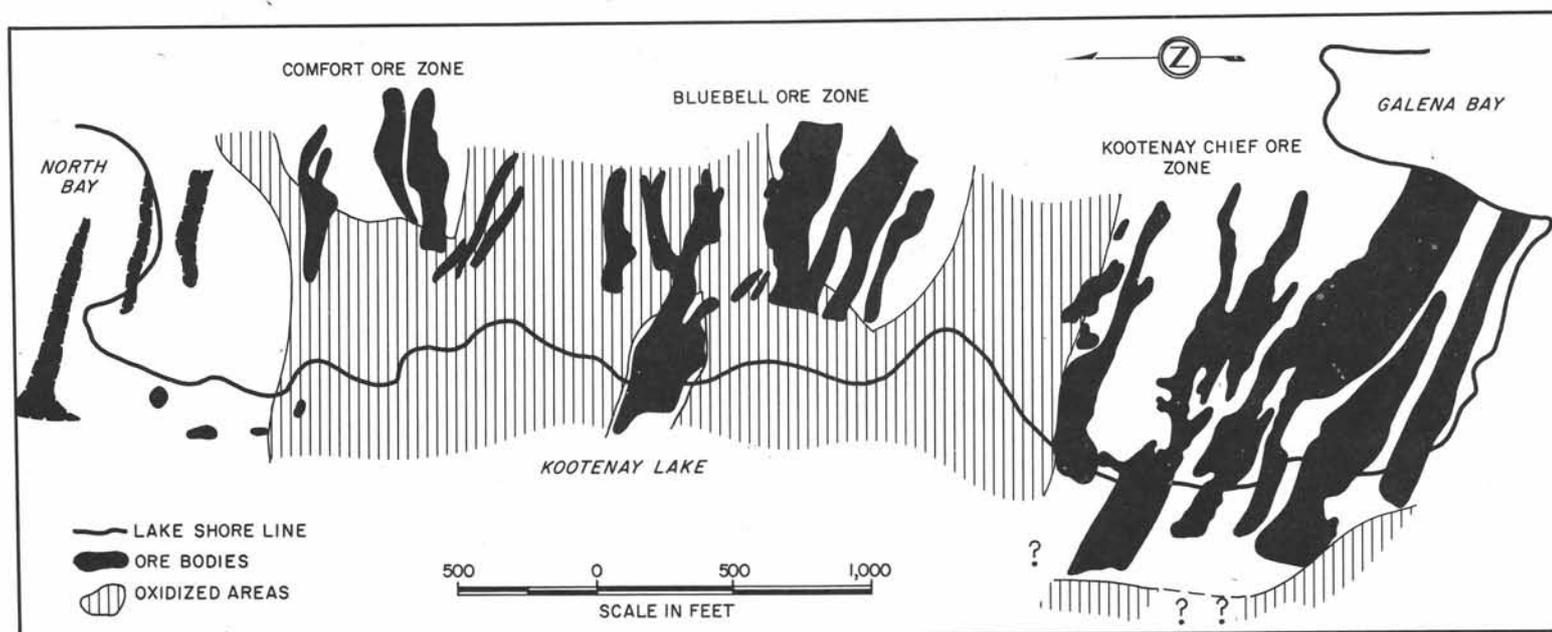
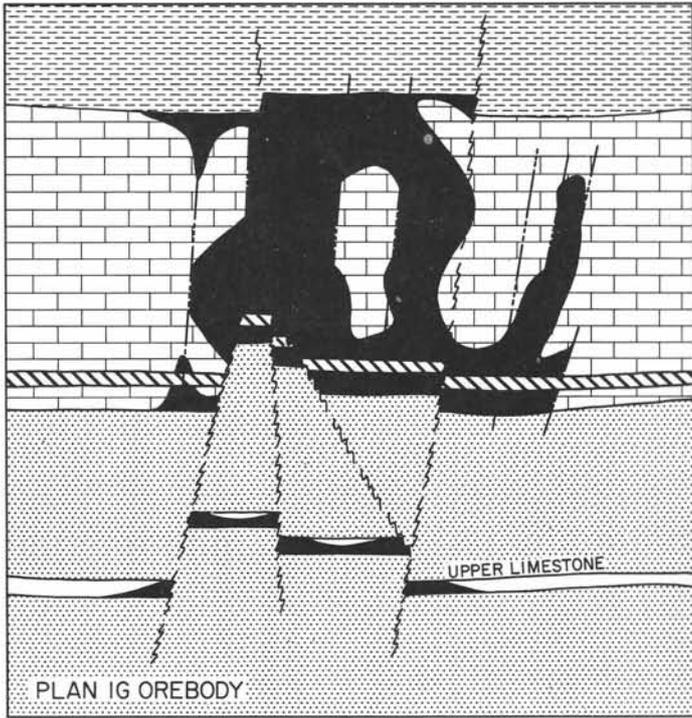
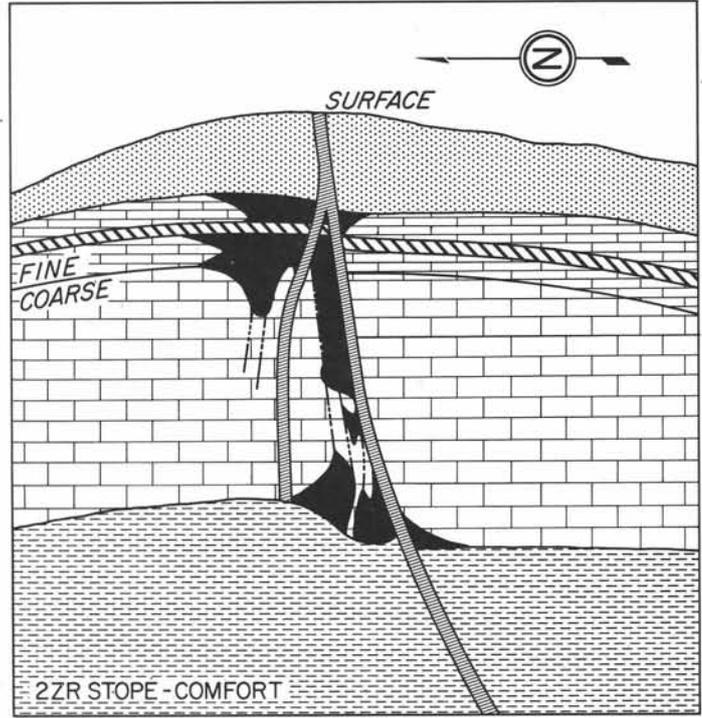


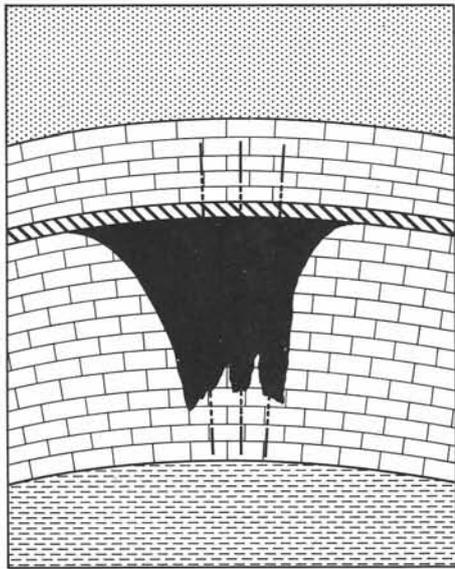
FIGURE X-3.—Plan of the Bluebell mine area showing the main ore zones and areas of oxidation.



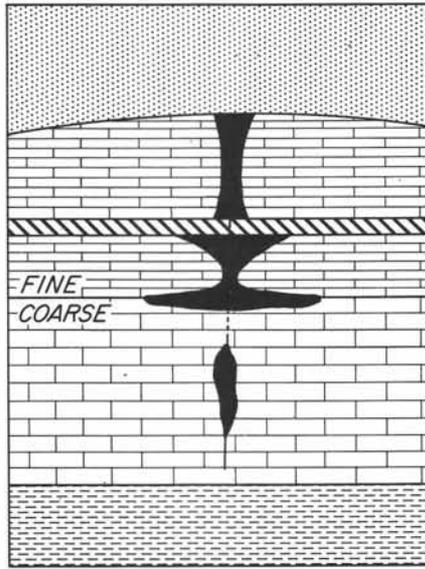
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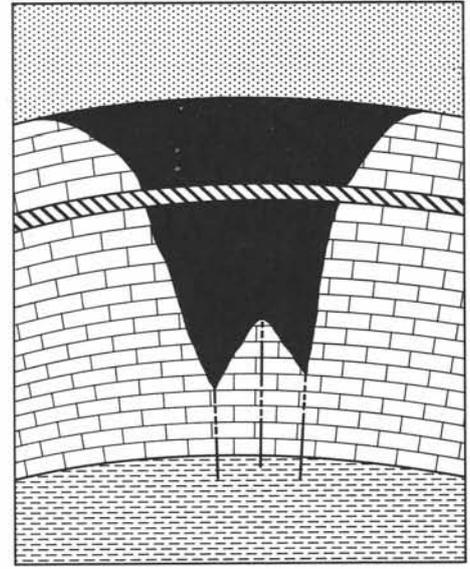
(C)



(B)



(D)



(E)

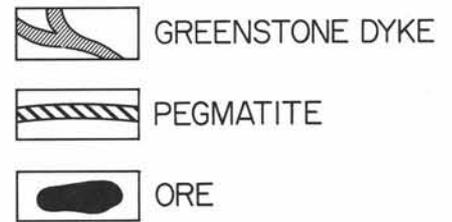
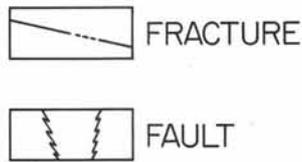


FIGURE X-4.—Diagrammatic plan (A) and longitudinal sections (B, C, D, and E) showing controls of mineralization. North arrow of C applies also to A, B, D, and E.

In the upper limestone there are ore bodies of limited extent. Mineralization is similar to that in the Bluebell Limestone, although there are fewer known structural controls. They extend over parts of the Bluebell and Kootenay Chief zones only (Fig. X-4a).

TEMPERATURES OF DEPOSITION

Paragenetic studies coupled with temperature measurements of fluid inclusions in ore and gangue minerals by Ohmoto and Rye (1970) have indicated that the massive ores were deposited at temperatures between 450° C and 500° C and that the later minerals in vugs, which comprise 10 percent of the hydrothermal mineralization, were deposited at temperatures between 320° C and 450° C. The confining pressure of the ore-forming fluids and the depth of ore deposition are estimated to be in the range of 300 to 800 atmospheres and 6 km respectively.

PARAGENESIS

Gross features of the ore zones suggest three major periods of hydrothermal mineralization:

- (1) Formation of knebelite.
- (2) Deposition of massive sulphides, quartz, and carbonates.
- (3) Development of crystals in vugs.

Periods (2) and (3) are probably different parts of the same period of mineralization (Ohmoto and Rye, 1970).

STRUCTURAL CONTROLS OF MINERAL DISTRIBUTION

Over the long history of the mine, mostly in connection with the smaller ore bodies, a great amount of detail has been noted about the effect of structure on mineralization in limestone. In larger structural situations, only a small part may be visible, making the basic situation much more difficult to recognize. It is hoped that the following descriptions will be of help to geologists working on the early stages of exploration and mining in limestone replacement ore bodies. Several different structural control situations are described, and any combination may exist.

THE MAIN ORE CONTROL

The main ore controls are outlined by Irvine (1957, p. 101, 102) as follows:

The beds in the vicinity of the mine form a gentle secondary synclinal cross fold, with a wave length of about three miles. The mine is situated within this folded structure which plunges down the dip of the beds.

Within the mine, conspicuous features in the Bluebell limestone are numerous steeply dipping cross-fractures, spaced at intervals of a few feet, and occurring continuously along the strike of the formation. Individual fractures have somewhat irregular surfaces, which indicates that they are tension fractures, resulting from warping of the limestone beds. These fractures may be unmineralized, or may be healed by sulphide, quartz, or carbonate. Smooth surfaced and flatter dipping cross-joints are also numerous, and are considered to be shear joints, resulting from the same force which caused the secondary folding. Statistical studies of the attitudes of the tension joints

by the use of point diagrams show that the attitudes of the joints differ significantly in the three ore zones.

	Strike	Dip
Comfort	N. 72° W.	83° N.
Bluebell	N. 75° 30' W.	82° S.
Kootenay Chief	N. 62° 30' W.	84° 30' N.

Separate point diagrams made for each 200 feet of strike length along the limestone formation show that these values do not change gradually along the strike, but that there are sharp discontinuities at points which are taken to mark the boundaries between the segments of the limestone in which conditions caused differences in the attitude of the tension joints. At these boundaries the limestone is more intensely fractured than usual, resulting in breccia-like masses which are up to 50 feet or more in width, measured along the strike of the beds. These brecciated boundaries between the limestone segments can be traced from level to level, and show that the segments trend down the dip of the beds, following the trace of the joints on bedding planes.

In each segment of the limestone formation, as defined above, the ore bodies occupy a central position along the strike, while the barren intervals between ore zones are near the edges of the segments, flanking the boundary breccia zones. This suggests that at the time the ore-forming fluids were circulating, special conditions prevailed which made the central portions of the limestone segments favourable for ore deposition.

In the Kootenay Chief and Bluebell ore zones, there is considerable gravity faulting in the central portions of the various zones, with strike separation of up to 30 feet along steep faults which, from their general attitude, appear to be simply tension joints along which faulting has occurred. . . . The pattern of displacement is strongly suggestive of collapse, following compressional arching of the beds.

In a section titled "Theory of structural control of ore deposition," Irvine (1957, p. 103) further stated:

Two essential facts concerning the Bluebell ore bodies must be taken into account in any theory of structural control. First, the ore shoots, almost without exception, have a rake which agrees exactly with the trace of tension fractures on beds, thus these fractures must have had an influence on ore deposition. Second, although tension fractures are spaced [somewhat] uniformly along the limestone formation, ore shoots occur only in those located [fairly] centrally in the limestone segments; that is, near the crests of the anticlinal arches.

Since only tension fractures near the crests of the arches are mineralized, it is a logical assumption that the mineralizing solutions were active at a time when arching had produced open fractures near these crests, and tighter fractures near the edges of the arches, and that the more open fractures afforded good channels for circulation and sites for ore deposition.

Ore-bearing solutions apparently circulated through this tension fracture system. (Fig. X-2).

The following local but important ore controls have been recognized:

HANGING WALL

Most of the crosscutting tension fractures do not enter the hanging wall, or if they do, they are too tight to be significant as channelways for mineralizing solutions. In the centres of the broad anticlinal arches, where mineralized fractures come in contact with it, the hanging wall

has considerable damming effect on solutions rising within the limestone. The ore bodies are usually richer and extend outward along strike much farther just under the hanging wall (Fig. X-4e).

FOOTWALL

A damming effect by the footwall of the limestone occurs only where mineralized cross fractures intersect the footwall at a synclinal fold, and there only in the immediate area where the mineralized cross fracture crosses the fold. Mineralization does not normally extend very far into the footwall. No ore bodies are found there, but under large ore bodies narrow mineralized fractures may be present (Fig. X-1c).

BROWN DYKES

These narrow (up to 2 feet thick), north-striking, east-dipping dykes are commonly offset by bedding plane faults in the limestone. Although the displacement on individual faults is small and the total amount is not known, it is estimated that, between the footwall and hanging wall of the limestone, the total displacement amounts to between 100 and 200 feet. Where a mineralized fracture intersects a brown dyke, commonly, though not always, the dyke acts as a local dam, causing enrichment on the down-dip side and a barren area on the up-dip side of the dyke. Normally the mineralization will reappear some distance up-dip, apparently because the mineralizing solutions traveled around the "dam" above or below where the dyke is broken by the dip faults (Fig. X-1b).

PEGMATITE SILLS

One nearly continuous sill 2 to 5 feet thick and normally about 12 feet below the hanging wall extends throughout the mine in the limestone. Commonly, mineralized tension fractures through the pegmatite are only $\frac{1}{8}$ to $\frac{1}{2}$ inch wide, with green chloritic alteration up to 2 inches from the fracture. Otherwise the pegmatite is not mineralized. In the limestone, sulphides have spread along strike below the sill. Where the pegmatite has been broken by intense folding or where the fracturing is strong, this damming effect is missing (Fig. X-4e). Beneath strong, unbroken pegmatite, the ore sometimes stops, leaving barren limestone between pegmatite and hanging wall (Fig. X-4b).

FAULTS

Some faults are parallel to the mineralizing fractures; others are at an acute angle to them. Where these faults are in contact with mineralizing fractures, the mineralization may follow up the dip of the fault along the limestone, also through the hanging wall. Several good ore bodies in the upper limestone have resulted from this type of control (Fig. X-4a). Other structures may also effect mineralization along the faults. Ore tends to be richer in galena along the mineralizing fractures and richer in marmatite and pyrrhotite near the edges of mineralization.

"GREENSTONE DYKES"

These dykes are pre-ore and subparallel to the mineralizing fractures (Fig. X-2). Some follow faults. Mineralization occurs along the dykes as it does along the

faults described above. Not all dykes carry mineralization, and those associated with intense fracturing or faulting have better ore bodies along them (Fig. X-4c).

CONTACTS BETWEEN COARSE- AND FINE-GRAINED LIMESTONE

The upper part of the Bluebell Limestone is mostly a fine-grained, fine-banded limestone, whereas the lower part tends to be coarser grained and vaguely or broadly banded. Where a mineralized fracture crosses the contact between these two types, the mineralization may spread out along the bedding (Fig. X-4d). This contact zone appears to have had a relatively high permeability, which probably was increased by movement between the beds and is now obscured by recrystallization.

THIN FRACTURES

Commonly, massive sulphide mineralization stops at a weak fracture of knife-edge thickness. Upon close examination, no apparent reason for the cutoff has been found, and it has been suggested that a damming effect was caused by a very thin film of mud along the fracture that subsequently was removed (Fig. X-4a).

RECUMBENT FOLDS IN THE LIMESTONE

There are many small recumbent folds in the limestone, some of which are confined to a fairly restricted set of beds. They are caused by plastic folding along the bedding planes. The folds may die out as the movement is dissipated along the bedding slips. Where these folds cross an ore body, mineralization follows the crest of the fold for some distance into the limestone outside the normal boundaries of the ore on both sides (Fig. X-1d). The area near the crests of these folds appears to have had a relatively high permeability.

ANTICLINAL FOLDS

These folds are noted both at hanging and footwall. At the hanging wall contact, there appear to be two sets of anticlinal folds, one plunging about 20° to the southwest, the other about 20° to the northwest, and at some places where they intersect they show a pronounced "dimpling" effect of the hanging wall. These "dimpled" areas contain higher grade ore where they intersect an ore body, and ore tends to follow the crests of the folds beyond the normal outlines of the ore body. An ore body along the footwall becomes much weaker or terminates altogether along the crest of such an anticlinal fold.

SUMMARY

The one common feature in all types of ore bodies is the mineralized tension fracture, singly or in groups. It is concluded that, in the Bluebell Limestone, given a deep-seated source of mineralization, any physical feature that increases the porosity or permeability of the rock will exert some control on deposition of the ore.

DEEP OXIDATION

The unique feature of oxidation at the Bluebell mine is that the outcrops and upper levels contain fresh sulphides, with only minor gossan, and oxides encountered

150 feet below surface extend to at least 1,000 feet below surface. Early operators of the mine who worked only the Bluebell, or central ore zone, encountered heavy oxidation on the north side about 150 feet below surface. This oxidation became progressively more intense southward with depth until the whole Bluebell ore zone was oxidized 300 feet below surface, leaving only small sulphide cores. On the level 375 feet below surface, occurrences of CO₂ gas along with water flows exceeded the mine's pumping capacity and forced progressive abandonment of the mine.

When Cominco acquired control of the property, diamond drill holes located good-grade, fresh sulphides in the Kootenay Chief and Comfort zones. The old mine was dewatered, and development was driven north and south at the mine 225-foot level. In both directions oxidized areas were penetrated before good sulphide ore was reached. Later development has shown extensive oxidation that in places has extended below the lowest level, which is 1,040 feet below the surface.

LOCATION

There are four areas of major oxidation in the mine (Fig. X-3). From north to south they are:

NORTH COMFORT

North of the main Comfort zone at surface and raking southwestward through the main Comfort ore zone is an area of oxidation that includes all the main Comfort ore bodies from about 250 feet to more than 825 feet below surface. Mining is active in ore sulphides in the north part of the Comfort zone, north of this oxide zone, between the 375-foot and 825-foot levels. (Level numbers indicate vertical distance to surface.)

NORTH BLUEBELL

Between the Bluebell and Comfort zones at surface and raking southwestward through the Bluebell ore zone between the 225-foot and 375-foot levels is an area of oxidation that includes all ore bodies in the main Bluebell zone to below the 825-foot level. There is one sulphide ore body on the north side of the main Bluebell zone, between the 375-foot and 825-foot levels.

NORTH KOOTENAY CHIEF

Between the Bluebell and Kootenay Chief ore zones at surface and raking almost due west is an oxidized area that extends from north of the Kootenay Chief to below the 825-foot level.

SOUTH KOOTENAY CHIEF

Another area of oxidation extends from near the 675-foot level to below the 825-foot level on the south side of the Kootenay Chief ore zone, and west of the Kootenay Chief to 300 feet below the 825-foot level.

The net result of this situation is that fresh sulphide ore bodies near surface become increasingly oxidized with depth to such an extent that in some places the ore bodies are completely oxidized and are uneconomic.

THE OXIDIZED ORE BODIES

Oxidation of the ore presumably started along the ore boundaries, or prominent fracture zones, where acid meteoric waters came in contact with sulphides. The resulting solutions attacked the limestone, leaving grey mud and generating CO₂ gas. Large crystals of selenite are occasionally found in this mud. The partly oxidized ore bodies show a halo of grey mud and oxidized ore material, while in some cases the central or higher grade portions of the ore remain relatively unoxidized. These have been extracted in some places, and because of bad ground a much higher than normal dilution has been accepted. Some of this "waste" dilution had a fair grade, but contained refractory oxides.

The sequence of oxidation appears to be: First the pyrrhotite shows tiny fractures coated with a spongy appearing pyrite. The pyrite gradually encroaches on the pyrrhotite until all the pyrrhotite has been converted to lacy or spongy pyrite, which may be intimately associated with unaltered galena, marmatite, arsenopyrite, knebelite, or quartz. Limestone and calcite disappear early. Oxidation next converts the pyrite to hematite or limonite, followed by oxidation of arsenopyrite, marmatite, and knebelite in that order. At this stage the ore body has a halo of grey mud, consists of red-brown to yellow, muddy material enclosing nodules of galena and quartz. Some of these nodules have been mined where the original ore bodies contained sizable bodies of galena. The final leaching product consists of only mud, oxides, and quartz. None of the explored oxide areas contains more than 8 percent combined lead and zinc. This fact has led to the conclusion that there has been a considerable metal loss during the leaching process.

EXPLANATION OF DEEP OXIDATION

A Hydrosonde survey of Kootenay Lake to determine the location of the bedrock bottom of the lake shows the lowest bedrock contours west of the Bluebell to be 1,400 feet below sea level, or 3,180 feet below the adit level which is at 1,780 feet above sea level. The eastern bank of the lake tends to follow roughly the prevailing dip slope at about 32°. The west side tends to be quite steep.

Investigation of the valley system shows that, most likely, the Duncan River flowed southward through the present Kootenay Lake Valley as far as Bonners Ferry, Idaho, where it met the Kootenay River, and both flowed southwestward through present Lake Pend Oreille in Idaho, and through Spokane, Washington. In Miocene time this ancestral river system was dammed near Bonners Ferry by a tongue of lava from the Columbia River Plateau (Gilbert, 1949; White, 1959). Above the dam the rivers then filled the valley, which overflowed westward through the valley of the west arm of Kootenay Lake near Nelson. This raised the water table to its present level. On the Bluebell peninsula the outcrops of sulphide ore occur on rocky cliffs along the east side of a small ridge near the shoreline. East of the ridge is a fairly extensive flat, the present townsite. The rocky ridge is cut in two places by narrow valleys leading to the lake shore. The surface outcrop of two of the major oxidized zones

coincides with the low areas in the narrow valleys, and the other two oxide outcrops occur near the north and south shores of the peninsula. It is thought that before the rivers were dammed and Kootenay Lake Valley filled, Bluebell peninsula, then a bench on the mountain-side, served as a catch basin for water, some of which circulated through fractures and later through cavernous openings in the limestone to the valley bottom. These structures have presumably been obliterated by the intensity and widespread action of oxidation (Fig. X-2).

THERMAL WATER

Another unusual geological feature in the Bluebell mine is thermal water. It did not occur at surface, but was discovered in the main drive in the Kootenay Chief zone on the 375-foot level. Later work indicated that the thermal water may have originally risen to a little above the 225-foot level and the upper openings became plugged with mud, an end product of solution of limestone by water. Occurrences of thermal water have become numerous and widespread at depth, suggesting that limestone dissolved in the lower levels may have precipitated as calcite and mud in the upper levels and reduced the flow. At the 825-foot mine level, thermal springs occur at intervals through the complete length of the workings, and probably contribute about half of the 5,200 gallons of water per minute pumped from the mine. Individual springs consist of very minor CO₂ or water flows, whereas others flow up to about 1,700 gallons per minute. Many of the larger springs have been grouted off. The springs are confined mainly to the limestone, although small flows through fractures have been encountered in both footwall and hanging wall rocks. The springs flow in pipelike openings and fractured areas, parallel, or nearly so, with the mineralizing tension fractures.

The thermal springs deposit almost pure CaCO₃ and do not have an oxidizing effect except where they come in contact with oxide zones or air, at which places limonite is deposited.

EFFECT ON MINING AND THE LIFE OF THE MINE

The thermal water emits a large amount of CO₂ gas, initially at high pressure, which poses major problems in ventilation, particularly of primary openings. Of second importance are major flows of water that must be pumped, and the precipitation of hard carbonates on all pumps, pipe lines, etc. The stopes near thermal springs usually have bad ground. These conditions cause heavy financial outlays that vitally affect exploration and mining (Hammond, 1969).

CHEMICAL NATURE AND TEMPERATURES OF THERMAL WATER

In the early stages of dealing with thermal springs in the mine, fairly complete chemical analyses were made. Later determinations were done only for temporary hardness (which compares approximately with total dissolved solids), chlorine, and pH.

Analyses of the thermal spring water show considerably higher chlorine content than nearby surface and lake waters, which are quite low in chlorine. It is assumed that the chlorine is derived from juvenile water, not from meteoric water, and that the origin of the thermal water could be estimated from its chlorine content. The temperature of the water ranges from about 44° F, which temperature suggests a high percentage of admixed lake water, to about 104° F. The hottest water contains the most chlorine. Thermal water is usually slightly acidic, but water from footwall rocks may be slightly alkaline.

Tabulated below are water analyses from the Bluebell mine and one from Ainsworth Hot Springs.

	Bluebell mine								Ainsworth Hot Springs
	Kootenay Chief Area				North Comfort Area				
	525' Lev.	875' Lev.	925' Lev.	1,125' Lev.	825' Lev. in F.W.	825' Lev. in 1s. near F.W.	Surface D. Drill Index Form.	Comfort 825' Lev.	
Temp.	80°F	70°F	90°F	104°F		44°F		86°F	112°F
Analyses in parts per million									
TDS	3,300	2,500	4,700	3,340	630	320	130	3,870	800
SiO ₂	110	60						175	140
Ca	430	410							
SO ₄	280	190						214	55
Mg	280	160						190	7.5
Cl	80	50	106	178	18.6	10.2	23.3	70	45
B	.1	0.3							
F	2.0	1.6							
Al ₂ O ₃	7	6							
Cu	.1	< .1						< .01	< .01
Pb	<2	<2						< .1	< .1
Zn	2	14						.02	< .01
P ₂ O ₅	5	5							
Fe	<1	<1						2.1	0.8
Na	440	290						440	215
Mn								< .05	< .05
Sr								5	1.2
K								60	20
pH			7.2	7.4	6.5	6.5	8.5		

In comparing the Ainsworth Hot Springs analysis with the Comfort 825' Lev. analysis (immediately to the left of the Ainsworth analysis), J. F. Harris, Cominco senior research geologist, stated:

Notable features are the very low concentrations of heavy metals. The most striking differences between the samples are in the contents of the alkalis, alkaline earths, and the combined anions, notably Mg, Na, K, SO₄. The Bluebell water is considerably more saline than the Ainsworth sample, and the differences in compositional ratios such as Ca/Mg and SO₄/Cl, seem at first sight to argue against derivation by differing dilution from a common source.

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XI

THE VAN STONE MINE, STEVENS COUNTY, WASHINGTON

By

S. NORMAN KESTEN



THE VAN STONE MINE, STEVENS COUNTY, WASHINGTON

By S. NORMAN KESTEN^①

The Van Stone mine is in the Colville Mountains at the headwaters of Onion Creek, a tributary of the Columbia River, in section 33, Township 38 North, Range 40 East, Stevens County, Washington (Yates, Fig. III-2, this guidebook). The district was found prior to World War I, and the Van Stone deposit was first explored by Hecla Mining Company in the middle Twenties. Willow Creek Mines did some diamond drilling and drove two short adits through the ore bodies between 1938 and 1942. In 1950 American Smelting and Refining Company acquired control of the property. After geophysical and geochemical work and considerable diamond drilling had been done, that company constructed a flotation concentrator with a capacity of more than 30,000 tons of ore per month for the production of lead concentrates and zinc concentrates. Production began in 1952. Low metal prices caused a shutdown from 1957 to 1964 and another closure from 1967 to 1969. During a total of approximately 8.5 years of operation ending December 31, 1969, about 7.5 million tons of rock has been removed from one small and one large pit, of which 41 percent has been ore. To the end of 1969, *recovered* metal totals about 10,500 tons of lead and 86,000 tons of zinc. Northwest of the southerly part of the North Pit, diamond drilling, which was started in the fall of 1969, has revealed the presence of additional bodies of mineralized rock that may be minable. If it is minable, the new ore will prolong the life of the mine.

The mine area is near the south end of the Kootenay Arc, otherwise known as the Selkirk Mountains Lead-Zinc Belt, extending southward from Revelstoke, British Columbia (Yates, Fig. III-1, this guidebook). The rocks in which the ore deposits occur are Paleozoic marine sediments overlain in part by Mesozoic formations and intruded by late Mesozoic batholiths. The important mineral deposits occur as replacement deposits in carbonate rocks, but vein-type ore bodies are found to a lesser extent in noncalcareous rocks. The bedrock surface is extensively covered by a thick mantle of coarse to fine outwash material from continental glaciers.

At the Van Stone mine, ore is found in the Middle Cambrian Metaline Limestone within several hundred feet of the Spirit pluton (Yates, Figs. III-5 and III-5A, this guidebook). The Spirit pluton is believed to be related to the Kaniksu batholith but appears to be a separate intrusion (Yates, oral communication). As described by Yates and Engels (1968, p. D-243) the Spirit pluton is a composite intrusion elongate in an east-west direction that cuts discordantly across northeast-trending folds. It consists of two phases: (1) a central mass of porphyritic biotite granodiorite locally bordered by (2) a nonporphyritic granodiorite. Both phases contain inclusions and selvage remnants of quartz diorite and hornblende gabbro. Both phases are cut by dikes and by elongate intrusions of quartz monzonite and alaskite. Contacts between the pluton and the intruded rocks are sharp. Within 100

feet of the contact the intruded rocks are metamorphosed to sillimanite-cordierite-, or wollastonite-bearing rocks. Biotite and andalusite-bearing rocks were formed as much as three miles from the contact. Radioactive age-dating of the various rocks that compose the pluton indicate that they range in age from 100.0 ± 2.8 million years to about 91.0 ± 3.0 million years, about the same age as the rocks that form the main mass of the Kaniksu batholith (Yates and Engels, 1968, Table 1, p. D-243).

The ore bodies lie along a broad S-shaped fold between the Spirit pluton on the north and the Rogers Mountain fault to the south (Cox, 1968, Fig. 1). The host for the ore is the Middle Dolomite Member of the Metaline Limestone. The dolomite is dark gray to gray to white-banded. In addition, the mineralized zones are characterized by the development of jasperoid and tremolite. No other alteration minerals are present in the dolomite in any quantity, and none are observed along the granite contact. Mineralization occurs as replacement of the host rock by brown to amber sphalerite with some associated galena in pods, streaks, and stringers. Minor pyrrhotite and pyrite, like the galena, are erratically distributed with respect to the sphalerite. Lamprophyric and diabasic dikes that cut through the host rocks are mineralized locally to a slight extent. Oxidation extends to about 80 feet below the surface. Any evidence of leaching and redeposition above the primary zone has been wiped off by glacial action or destroyed by mining. Studies are in progress to try to determine, among other things, the relation of mineralization to both types of alteration and of both mineralization and alteration to the intrusion.

A visit to the mine will consist principally of a trip into the North pit, because the only geologic feature to be seen beyond the pit rim is the glacial debris, there being no exposed rock nearby. In the pit there will be afforded an opportunity to examine the mineralization and attendant alteration and, to some small extent, possibly near the entrance to the pit, the unaltered and barren dolomite. One of the basic dikes may be accessible at the time of the visit, and it may be possible to see from a distance a major fault crossing a corner of the pit at its east end. Because the rock exposed is mostly altered, a good example of bedding may be hard to find. In the ore zone, bedding may be easily confused with jointing or other sets of parallel fractures.

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