Society of Economic Geologists’ Coeur d’Alene Field Conference Idaho—1977

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Edited by
Rolland R. Reid
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<table>
<thead>
<tr>
<th>Contents</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Preface</td>
<td>v</td>
</tr>
<tr>
<td>Pricard-Ravalli (Belt-Precambrian) Sediments Through Space and Time,</td>
<td></td>
</tr>
<tr>
<td>by Martin Mumma, Gerald Harbour, and Douglas Jayne</td>
<td>1</td>
</tr>
<tr>
<td>A Proposed Set of Exploration Criteria for Delineating Potential Target</td>
<td>3</td>
</tr>
<tr>
<td>Zones Within the Coeur d'Alene Mining Region, by Donald C. Springer</td>
<td></td>
</tr>
<tr>
<td>Ore-Stratigraphic Relationships in the Coeur d’Alene District,</td>
<td>23</td>
</tr>
<tr>
<td>by J. C. Farmin</td>
<td></td>
</tr>
<tr>
<td>The Revett-St. Regis “Transition Zone” Near the Bunker Hill Mine,</td>
<td>25</td>
</tr>
<tr>
<td>Coeur d’Alene District, Idaho, by Brian G. White and Don Winston</td>
<td></td>
</tr>
<tr>
<td>Structure of the Bunker Hill Mine, Kellogg, Idaho, by Dwight S. Juras</td>
<td>31</td>
</tr>
<tr>
<td>Stratiform Mineralization and Origin of Some of the Vein Deposits,</td>
<td>35</td>
</tr>
<tr>
<td>Bunker Hill Mine, Coeur d'Alene District, Idaho, by V. M. Ramalingaswamy</td>
<td></td>
</tr>
<tr>
<td>and E. S. Cheney</td>
<td></td>
</tr>
<tr>
<td>Petrography of Stratiform Lead-Zinc-Silver Deposits at the Bunker Hill</td>
<td>45</td>
</tr>
<tr>
<td>Mine, Idaho, with Some Thoughts on the Timing of Ore Deposition,</td>
<td></td>
</tr>
<tr>
<td>by Rolland R. Reid</td>
<td></td>
</tr>
<tr>
<td>Disseminated Galena in the Lucky Friday Mine, Coeur d’Alene District,</td>
<td>49</td>
</tr>
<tr>
<td>Idaho, by C. E. Hauntz</td>
<td></td>
</tr>
<tr>
<td>Characteristics of Lead-Zinc-Silver Veins Located in Belt Rocks to the</td>
<td>51</td>
</tr>
<tr>
<td>North of the Coeur d’Alene District, by William R. Green</td>
<td></td>
</tr>
<tr>
<td>Metallocogenesis in the Coeur d'Alene Mining District, by Wilfred</td>
<td>55</td>
</tr>
<tr>
<td>Walker</td>
<td></td>
</tr>
<tr>
<td>Age of the Crossport C Sill Near Eastport, Idaho, by R. E. Zartman,</td>
<td>61</td>
</tr>
<tr>
<td>Z. E. Peterman, J. D. Obradovich, M. D. Gallego, and D. T. Bishop</td>
<td></td>
</tr>
<tr>
<td>Reconnaissance Geology of the Blackbird Mountain Quadrangle,</td>
<td>71</td>
</tr>
<tr>
<td>Lemhi County, Idaho, by Earl H. Bennett</td>
<td></td>
</tr>
<tr>
<td>Upper Proterozoic Stratigraphy of Northwestern Canada and Precambrian</td>
<td>73</td>
</tr>
<tr>
<td>History of the North American Cordillera, by G. M. Young, C. W.</td>
<td></td>
</tr>
<tr>
<td>Jefferson, G. D. Delaney, G. M. Yeo, and D. G. F. Long</td>
<td></td>
</tr>
<tr>
<td>Some Proterozoic Sediment-Hosted Metal Occurrences of the Northeastern</td>
<td>97</td>
</tr>
<tr>
<td>Canadian Cordillera, by G. D. Delaney, C. W. Jefferson, G. M. Yeo,</td>
<td></td>
</tr>
<tr>
<td>S. M. McLennan, J. D. Aitken, and R. F. Bell</td>
<td></td>
</tr>
<tr>
<td>Regional Aspects of the Helikian (Precambrian V) Little Dal Group and</td>
<td>117</td>
</tr>
<tr>
<td>Correlatives, Northwestern Canada, by J. D. Aitken and M. A.</td>
<td></td>
</tr>
<tr>
<td>Semikhatchov</td>
<td></td>
</tr>
<tr>
<td>Depositional Environments and Stratigraphic Setting of Rocks of the</td>
<td>119</td>
</tr>
<tr>
<td>Tsetsoite Formation and Katherine Group, Mackenzie Fold Belt, Yukon and</td>
<td></td>
</tr>
<tr>
<td>Northwest Territories, Canada, by Darrel G. F. Long</td>
<td></td>
</tr>
<tr>
<td>Conference Discussion, Rolland R. Reid, moderator</td>
<td>121</td>
</tr>
</tbody>
</table>
Preface

The Society of Economic Geologists' Coeur d'Alene Field Conference, Idaho—1977 was held in Wallace, Idaho, on November 3-5, 1977. Sponsors for the conference were the Society of Economic Geologists, College of Mines and Earth Resources at the University of Idaho, Idaho Bureau of Mines and Geology, The Bunker Hill Company, Sunshine Mining Company, American Smelting and Refining Company, Hecla Mining Company, and Day Mines. Over 120 managers and scientists from the mining industry, state and federal governments, and universities attended this three-day meeting.

Rolland R. Reid
Conference Organizer
Prichard-Ravalli (Belt-Precambrian) Sediments
Through Space and Time

by

Martin Mumma¹, Gerald Harbour¹, and Douglas Jayne¹

ABSTRACT

Lower and middle Belian sedimentary rocks (Prichard, Burke, Revett, St. Regis, and Wallace Formations) in the western portion of the Belt basin in western Idaho and Montana and eastern Washington were deposited mostly in paralic and shallow open-marine environments. There appears to have been a north-south depositional hinge line along a zone on either side of a line between Libby and St. Regis, Montana. East of this zone deposition occurred mainly in open-marine environments, and westward sedimentation was predominately in paralic and very shallow near-shore environments.

The Prichard Formation, particularly the lower Prichard, represents the deepest water sedimentation, being deposited on an open-marine shelf perhaps in water several hundred feet deep. In the upper Prichard, however, noticeably west of the hinge line, sedimentary rocks represent intertonguing near-shore marine and nonmarine fluvial processes. In this horizon, material was deposited as offshore barrier bars, beaches, back bar lagoonal or bay deposits, and perhaps as tidal flat deposits that bordered a low coastal fluvial system.

This sedimentary regime is even more pronounced in the Burke and Revett Formations and indicates a general marine regression that resulted in the development of sizable barrier bars during Revett time. Fluvial-deltaic sedimentation was minimal.

The St. Regis sediments were laid down on supratidal and intertidal mud flats and in shallow bay and shelf environments. Basinwide shoaling occurred at this time accompanied with basin expansion and shallow marine transgression. This trend continued through Wallace time.

Wallace sediments appear to have been deposited in a region made up of shallow carbonate banks and inter-bank basins that may have had restricted circulation where fine muds accumulated. There is also evidence that some of the sequences of irregular intermixed carbonates and sands were deposited on very low supratidal flats. Those sediments containing "molar tooth" structures may have also been deposited this way, originally as "chicken wire" gypsum-anhydrite that was later altered to calcite by chemical processes not now understood.

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A Proposed Set of Exploration Criteria for Delineating Potential Target Zones Within the Coeur d'Alene Mining Region

by

Donald C. Springer*

The Coeur d'Alene mining region ranks first in the annual production of silver currently being mined in the world. The top four producing silver mines in the United States are in the Coeur d'Alenes. Recorded production in the district started in 1884, and through 1975 the region has recovered almost $2.9 billion worth of metals. Thus far the district has produced a total of almost 900 million ounces (25.5 million kg) of silver. This production is surpassed only by that of Potosi, Bolivia, which was discovered in 1544 and has produced about one billion ounces (28.3 million kg), but is now worked out. In addition, the district has produced 7 1/2 million tons (6.8 million metric tons) of lead, 5 million tons (2.7 million metric tons) of zinc, 1/2 million ounces (14,000 kg) of gold, and over 100,000 tons (90,000 metric tons) of copper.

Production continues in the district at the rate of over $120 million worth of recovered metals per year from at least seven producing mines. The annual future silver production for the region is projected to total between 14 and 19 million ounces (400,000 and 540,000 kg) from known reserves for at least the next 10 to 15 years. Mine reserves are traditionally replaced on a 1:1 basis by systematic exploration and development work in each of the producing mines. With the replacement of mined reserves on a continuing basis, the district may continue to produce for several more generations. It is estimated that by 1984, the centennial year, the Coeur d'Alene mining region will have out-produced Potosi, Bolivia.

Most of the early production came from the long, wide, persistent veins found in the Mullan, Burke, and Kellogg areas (Figure 1). These ores were primarily argentiferous galena and sphalerite. A few small prospects up Big Creek and McFarren, Rosebud, and Lake Gulches had produced small quantities of argentiferous galena, native silver, and some gray copper from shallow workings, principally in the Wallace Formation, but the ore shoots were small, irregular, and discontinuous.

Continued work on the prospects up Big Creek led to the development of a very rich silver vein on the 1,700 level of the Sunshine Mine in 1931. By 1937 the Sunshine Mine had become the largest silver producer of any mine in the world.

I am told the Sunshine Mine alone has now produced about four times more silver than the entire Comstock district. Work up McFarren and Rosebud Gulches developed the Coeur d'Alene, Polaris, and Silver Dollar mines on similar tetrahedrite ores. In the late 1940's deeper exploration was started on the prospects up Lake Gulch, and this work led to the development of the second largest silver mine in the United States, the Galena Mine. Major expansion of the Lucky Friday Mine, which lies just a mile east of Mullan, developed this mine into the third largest silver mine in the United States during the late 1950's. More recent work in Shields Gulch has developed the Coeur Mine which may become the nation's fourth largest silver producer.

In addition to the Bunker Hill, Lucky Friday, and Star-Morning mines, production is coming from four mines located along the strike length of the known Silver Belt (Figure 2). The present producing mines, beginning at the western end of the Belt, are the Crescent, Sunshine, Coeur, and the Galena on the east. As seen on the map of the known Silver Belt, each of the principal veins appears to be structurally associated with one or more of the major faults that diagonally link the Osburn and Placer Creek faults. These vein openings are believed to be the result of rhomboid-tension forces which developed along and between the major diagonal faults. In some places there are indications that the host fractures may have been further developed and enhanced by the formation of cymnod loop and multiple cymnod cam or lens-shaped fracturing which developed along or within the rhomboid-tension zones. Although exposures are incomplete, suggestions of the typical or modified cymnod loop patterns have been observed in the Bunker Hill, Sunshine, Coeur, and possibly the Galena mines.

Thus far, exploration and development on the known veins in the Silver Belt have been the most complete and successful within the north limb of the Big Creek anticline. The approximate surface trace of the axial plane of the anticline is illustrated by the dot-dash line on the Silver Belt map (Figure 2).
Figure 2. Coeur d’Alene Silver Belt.
The north limb may or may not be significant as far as the occurrence of veins is concerned in the Silver Belt, but the attitude of the bedding in relation to the attitude of the veins may have a drastic effect on the size of the orebodies (Figure 3). In general, the bedding on the south limb of the anticline is relatively gently dipping, whereas the beds on the north limb are predominately steeply dipping—even steeply overturned in some localities. Most of the productive veins are also steeply dipping, 60 to 90 degrees to the south. The combination of steeply dipping beds and steeply dipping veins, but not parallel in both dip and strike, is conducive to the development of extreme down dip persistence of the ore shoots. The primary reason for this persistence is that vein structures are within the more favorable host rocks for longer distances in this configuration than as a steep vein in flatly dipping beds. If the same geometric relationships were to exist within the unexplored south limb, it is conceivable we may have comparably sized orebodies there also.

Statistical evidence indicates over 70 percent of the ore produced from the Coeur d'Alene mining region has come from ore shoots occurring within the transition rocks between the St. Regis and the Rebeet Formations. This relationship is currently considerably higher in the Silver Belt, because exploration and development have not yet extended into the transition rocks between the Burke and the Richard Formations. Statistically these latter rocks are the second most favorable stratigraphic horizon in the district. Ore shoots occurring within these latter transition rocks account for over 23 percent of the past production from the region.

This leads us to at least three basic criteria for large, persistent ore shoots in the district: (1) structural preparation; (2) veins cutting the beds at relatively small angles; and (3) favorable host rocks.

The vertical longitudinal projection (Figure 4) of the Silver Belt reveals some relatively lean areas between the major ore shoots. Thus far, underground exploration has not found any appreciable quantity of ore within these relatively blank areas, even though all three criteria are present. From this, it would appear there must be at least one other criterion, which when identified, may possibly be used as another aid in delineating favorable areas for future exploration.

During the late 1960's the U. S. Geological Survey started a geochemical sampling program in the district under the direction of Garland B. Gott. This program involved the collection and analysis of 8,195 soil samples and 4,000 rock samples covering an area roughly 38 by 15 miles (52 by 24 km). These samples were collected at 300- to 500-foot (100-150 meters) intervals along the ridges and roads in the district and were analyzed for 36 elements using a variety of analytical techniques—mostly atomic absorption. These analytical data were subsequently put on magnetic tape, and a computer-implemented graphics technique was used to illustrate the geochemical relationships. Many of these graphics have been released and open-filed by the USGS.

Several interesting relationships are depicted by the graphics for the individual elements and the ratio studies. Of particular interest is that over 93 percent of the ore sampled in the region occurs within zones of high manganese combined with high for other key elements. The manganese association with ore is most vividly illustrated in Figure 5.

The high manganese bands more or less form links or "ladder rungs" between the Osburn and Placer Creek faults. Most of the major productive veins are also within one of these high manganese bands. Conversely, the lean zones, as depicted on the longitudinal projection, are between the high manganese bands. Within the Silver Belt, the isopleth maps for most of the other metals show an almost complete, strong, east-west zone of highs along the trend of the known orebodies.

The present pattern of antimony in rocks within the district is shown in Figure 6. Most of the known orebodies are also within or adjacent to the areas containing abnormally high quantities of antimony.

The trend of metal highs is almost at right angles to the manganese highs and therefore crosses the manganese at the ore zones. From these studies it appears that surface geochemistry may be a fourth criterion, which when used with the other three, may possibly be a useful tool for determining areas for future exploration.

As more and more information was developed and accumulated, there were numerous attempts to reconstruct a plan map of the district to the approximate configuration of the veining which may have existed at a time before postore, strike-slip movement on the major faults. Matching various features such as stratigraphy, fold axes, and older faulting have been used for these reconstruction projects.

Gott and Bobolz made a reconstruction of the district using these features, plus the assumption that the Gem stocks were truncated and displaced from the Dago Peak stocks by the Dobson Pass fault, and further refined the realignment using the surface geochemical patterns. Their open-file report uses antimony to illustrate these geochemical relationships; however, several other elements depict essentially the same pattern.

Figure 7 shows the manganese reconstruction and that the bands on the south side of the Osburn fault match those on the north side of the fault.

Now, when the antimony is combined with the manganese, it appears that almost all of the ore-bearing
Figure 3. Idealized Silver Belt ore shoots.
Figure 4. Coeur d'Alene Silver Belt vertical longitudinal projection, looking north.
Figure 5. Coeur d'Alene mining region manganese belts after postore faulting.
Figure 6. Reconstruction Coeur d'Alene mining region manganese belts before postore faulting.
Figure 7. Coeur d'Alene mining region antimony zones after postore faulting.
Figure 8. Reconstruction Coeur d'Alene mining region antimony and manganese before postore faulting.
veins occur within zones of high manganese and high antimony (Figure 8). This same relationship holds for many of the other base metals, and for more detailed reconnaissance sampling it has been found that silver, lead, zinc, and copper are readily amenable to this technique.

It is interesting to note how the zone of orebodies, beginning with the Galena, Coeur, and Coeur d'Alene mines on the south side of the Osburn fault, extends northward across the fault to include most of the mines on the north side of the Osburn fault. This zone is apparently interrupted by the intrusion of the Gem stocks, but continues on the north side of the stocks with the zinc orebodies found in the upper Prichard rocks in the Sunset-Carbon Center area.

The high manganese-antimony area a mile or so immediately southwest of the Sunset-Carbon Center area after reconstruction is believed to be particularly significant in that Revet-St. Regis transition rocks are found at the surface. Several northwestward-trending, throughgoing faults and some small veins containing tetrahedrite, galena, and sphalerite were haphazardly prospecting years ago.

To determine the prospective merit of this area a reconnaissance geochemical soil-sampling program was initiated on a block of over 200 contiguous mining claims (Figure 9).

Soil samples were collected from the top of the "B" soil horizon at 100-foot (31 m) intervals along the claim end lines making the traverse lines essentially 1,500 feet (460 m) apart. In that this was a reconnaissance survey, only the alternate samples were analyzed; the remaining samples were retained for possible future study. The samples were screened; the -80 mesh was digested in hot aqua regia; and atomic absorption analyses were made for manganese, silver, lead, zinc, and copper. The mean and standard deviation were calculated from the assay data for each of the five elements, and isoploth maps were drawn using the mean, the mean plus one standard deviation, and the mean plus two standard deviations as isoploth intervals. Figures 10-14 illustrate the results of this phase of the project.

This mass of information has been reduced further by using the manganese as a prerequisite, so that only those higher metal value areas occurring within a high manganese isoploth are considered to be zones worthy of further investigation. Figure 15 illustrates the composite anomalies resulting from this procedure. The solid areas are considered primary zones and are areas containing the mean plus one standard deviation manganese and the mean plus two standard deviations silver, lead, zinc, or copper. The patterned areas are considered secondary zones that contain mean manganese and mean plus two standard deviations base metals.

To test the effectiveness and reliability of the 200-foot (62 m) sample spacing and 1,500-foot (460 m) line spacing, two key areas were resampled the following year by collecting and analyzing samples at 100-foot (31 m) intervals along lines 400 feet (120 m) apart. This latter sampling confirmed the original work and detected essentially no significant differences in location, size, shape, or magnitude of the anomalies, thereby indicating that closely spaced soils sample lines may not be necessary in the Coeur d'Alene mining region.

By comparing the anomalous areas to the surface geology, several anomalies appear to occur within two relatively distinct belts: (1) along or in close proximity to the contact between the Revett and the St. Regis Formations in the footwall of the Blackcloud fault, and (2) in the Revett and Burke Formations along the footwall of the Carpenter Gulch fault. In addition, most of the anomalies appear to strike N. 60°-70° W. within a structural pattern of N. 20°-50° W. trending throughgoing faults. These relationships are very similar to the more well-known productive portions of the district. Furthermore, the reconstruction maps suggest the Blackcloud fault may be a northwestward continuation of the Star-Morning fault and vein zone.

These comparisons reveal at least three of the basic criteria necessary for a vein-type orebody in the Coeur d'Alene mining region. These are structural preparation, favorable stratigraphy, and composite geochemical anomalies, all within a high manganese-antimony zone.

A cross section drawn through the larger anomalies associated with the Blackcloud fault illustrates some of the possible spatial relationships which may help guide the future exploration of these more promising zones (Figure 16). It is interesting to note that these particular zones lie relatively close to the surface and may be initially prospected with a series of well-designed and inexpensive diamond drill holes.

In summary, most of the large, productive, silver-bearing veins, which have been mined in the Coeur d'Alene mining region to date, occur within the transition rocks between the Revett and St. Regis Formations and the Burke and Prichard Formations. Most of the veins are associated with major, northwestward-trending throughgoing faults. The size of the orebodies is probably determined by the geometric relationships between the strike and dip of the veins and the attitude of the structures. Most of the ore appears to occur within zones containing high manganese and high base metal values in the soils and float rocks overlying the deposits. Through diligent use of these factors, the zones worthy of further exploration may be delineated. I am confident the prudent exploration of many of the areas thus delineated will result in the discovery of additional major orebodies within this most rare of geologic oddities—the Coeur d'Alene mining region.
Figure 9. Topographic map of the 200 contiguous mining claims and the surrounding areas that were sampled, Coeur d'Alene Mining District, Idaho.
Figure 10. Manganese isopleths of the 200 contiguous mining claims, Coeur d'Alene Mining District, Idaho.
Figure 11. Lead isopleths of the 200 contiguous mining claims, Coeur d'Alene Mining District, Idaho.
Figure 12. Silver isopleths of the 200 contiguous mining claims, Coeur d'Alene Mining District, Idaho.
Figure 13. Copper isopleths of the 200 contiguous mining claims, Coeur d'Alene Mining District, Idaho.
Figure 14. Zinc isopleths of the 200 contiguous mining claims, Coeur d’Alene Mining District, Idaho.
Figure 15. Surface geology and composite geochemical anomalies.
Figure 16. Cross section drawn through the larger anomalies associated with the Blackcloud fault.
Ore-Stratigraphic Relationships in the Coeur d’Alene District

by

J. C. Farmin

ABSTRACT

Tentative and broad conclusions concerning ore-stratigraphic relationships in the Coeur d’Alene mining district, north Idaho, are drawn from the research work of White, Juras, Reid, Winston, and Harbour, 1974-77, and from conversations with Crosby, Droste, and others. The conclusions were prompted by observations over a number of years and by some detailed work by Farmin and Meyer.

Qualitatively, in the Bunker Hill Mine, sphalerite preferentially replaces siderite in west-northwest, overturned fold fabric and in fold-related northwest-striking reverse faults. Ores rich in sphalerite, relative to galena, occur almost entirely in the quartzites and sericitic quartzites of the lower 1,200 feet of the upper Revett Formation, as the formation is defined by Winston and White.

Galena preferentially replaces quartz in east-west and northeast-striking veins. Galena ores are hosted by the quartzites and sericitic quartzites of the upper 1,400 feet of upper Revett (and lower St. Regis). Mixed ores occur in the 400-foot stratigraphic interval where sphalerite and galena zones overlap in the 2,000-foot ore-bearing section.

Galena ore rich in silver content appears to occur specifically in a 600-foot stratigraphic interval high in the upper Revett Formation, as defined by Winston and White. Tetrahedrite silver ores appear richest in this stratigraphic interval also, although they have been mined from the entire upper Revett and St. Regis Formations in mines of the Silver Belt.

In general, these relationships hold for all major mines in the district which are in the upper Revett Formation. There is no known ore in middle and lower Revett.

A parallel is drawn between the “mixed clean quartzites and salt and pepper quartzites” of the upper Prichard, as defined by Harbour, and the quartzites of the lower upper Revett. It appears that these are both sphalerite hosts. A further parallel is drawn between the upper upper Revett, with relatively thinner quartzites, more sericitic quartzite and more silite-argillite, on the one hand, with Harbour’s sparsite, thin quartzites, and silites of the lower Burke. These intervals host galena. The important difference is that lower Burke contains very little argillite. It is provocative to note that galena enriched in silver, relative to the well-established district ratio, is rare in Burke-Prichard hosts, whereas galena deficient in silver, relative to the norm, is not uncommon. Also, frozen wall fissure fillings seem to be more abundant in Burke-Prichard rocks than in Revett rocks, where replacement is common.

Several alternative explanations for stratigraphic control are suggested: (1) Minerals are deposited in a classic paragenetic sequence as solutions ascend through progressively younger rocks—quartz, pyrite, siderite, sphalerite, galena. The favorability of quartzites is strictly mechanical. (2) Mineral deposition is somehow controlled by differences in the thermal conductivity among lithologies. Quartzites are conductors, argillites are insulators (Crosby). (3) Wall rocks supply siderite or quartz to the gangue, depending on lithology. Solutions are charged or recharged with the favorable gangue when passing through rocks containing the requisite gangue producer (White, Farmin). (4) Thin quartzites, which are interbedded with silite-argillite, crush and fold more easily than thick quartzites to form better structural locations for ore (Reid, Murray). (5) Diagenetic reduction of feric iron to ferrous iron may in some way form favorable host rocks, as may be the case with anomalous copper in Missoula Group rocks (Winston). (6) Some constituent of specific sericitic quartzites or silite-argillites is somehow catalytic to tetrahedrite deposition (Farmin).

All of these suggestions may be in difficulty. All contain many unknowns. A viable hypothesis for stratigraphic control must be compatible with Juras’s hypothesis for structural control and Reid’s rules for refraction control.

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The Revett-St. Regis "Transition Zone" Near the Bunker Hill Mine, Coeur d'Alene District, Idaho

by

Brian G. White1 and Don Winston2

ABSTRACT

The "transition zone" between the Revett and St. Regis Formations is a major stratigraphic localizer of ore in the Coeur d'Alene mining district. However, the concept of a "transition zone" from Revett quartzite below to St. Regis purple argillite above is fundamentally incorrect, having arisen from the erroneous characterization of the Revett Formation as "quartzite" and the too-liberal extrapolation of observations of "hydrothermal bleaching." Strata of the so-called "transition zone," as well as lower portions of strata that have been locally included within the St. Regis Formation, are logically placed within the Revett Formation.

The Revett Formation near the Bunker Hill Mine contains three major rock types that reflect combinations of grain size, sorting, mineralogy, and sedimentary structures. The rock types are (1) crosbedded and horizontally laminated, medium- to thick-bedded, hard, vitreous, light gray or white quartzite, here named vitreous quartzite; (2) horizontally laminated, medium- to thick-bedded, light to medium gray-green, subvitreous, sericitic quartzite, here named sericitic quartzite; and (3) horizontally and ripple cross-laminated, fine-grained, thin-bedded strata, which contain abundant reworked mud inclusions or "mudskins," here named silty argillite.

Several characteristics distinguish the Revett Formation from the overlying St. Regis Formation: in addition to containing white vitreous quartzite intervals, the Revett Formation exhibits grayish and greenish colors, "mudskins," and rusty weathering stains; whereas the St. Regis Formation is more argillitic and contains abundant purple and brighter green hues, mud chips, and desiccation cracks. The boundary between the two formations is defined by the lowest purplish argillite beds of the St. Regis lithology.

The Revett Formation contains well-defined lower, middle, and upper members. The lower Revett member (more than 260 meters thick) contains thin intervals of vitreous quartzite and sericitic quartzite. The middle Revett member (130 meters) primarily contains silty argillite near the Bunker Hill Mine, but may become more quartzitic several kilometers eastward. The upper Revett member (600 meters thick) contains alternating, thick intervals of quartzite and silty-argillite, some of which correlate for at least 6 kilometers. The upper Revett member essentially equals the so-called "transition zone," although it does not, in fact, represent a lithologic transition.

The Revett Formation represents braided stream and fan-delta sequences, analogous to those of the Missoula Group, but in a more moist sedimentary environment.

INTRODUCTION

Many geologists who have worked in the Coeur d'Alene mining district believe that ore is concentrated in the "transition zone" across the Revett-St. Regis formational boundary. The Revett Formation has been conceived to be predominantly hard, vitreous white quartzite that grades through an interval many tens of meters thick into the overlying St. Regis Formation of "impure quartzite" and purple argillite. Furthermore, extensive "hydrothermal bleaching" is believed to have preceded ore emplacement, changing purple St. Regis colors to green. Many district geologists have believed that the bleaching, combined with the favorable mixture of lithologies within the "transition zone," have somehow promoted ore concentration.

We propose to demonstrate the following in the vicinity of the Bunker Hill Mine: (1) that the Revett Formation is not predominantly quartzite, but, in reality, is composed somewhat equally of interbedded units of vitreous quartzite, sericitic quartzite, and greenish silty-argillite; (2) that the greenish silty-argillite of the Revett Formation differs fundamentally from argillite within the St. Regis Formation; (3) that the silty-argillite lithology of the Revett Formation is not part of
a single "transition zone," but rather extends far down into the Revett Formation as repeated stratigraphic intervals 50 meters or more thick; (4) that therefore the ore-bearing strata lie primarily within the Revett Formation, not necessarily near the upper boundary; (5) and that the greenish color within the Revett Formation, actually resulted from diagenesis, so that "hydrothermal bleaching" in the district has been greatly exaggerated. Finally, we shall propose an environmental model for deposition of the Revett Formation.

This paper is a product of an extensive research program that has been undertaken by the Bunker Hill Company. From 1975 to 1977, the senior author measured and described stratigraphic sections and mapped stratigraphy within the Bunker Hill Mine and on the Bunker Hill property, as a full-time participant in this research. In 1975 Don Winsom, assisted by Phil Jacob, a geology student at the University of Montana, mapped and measured sections on the surface. Our experience with these rocks is primarily limited to the Bunker Hill area (Figure 1), which includes, however, the most extensive surface occurrence of the Revett Formation in the western portion of the Coeur d'Alene district.

ROCK TYPES

Three distinct lithologies or rock types, based on grain size, mineralogy, color, and sedimentary structures, comprise the 1,200 meters of the Revett Formation in the Bunker Hill area. They are (1) the vitreous quartzite, (2) the sericitic quartzite, and (3) the silite-argillite.

VITREOUS QUARTZITE

The vitreous quartzite originated as well-sorted, fine- to very fine-grained sand that is now lithified into hard, light gray or white, commonly vitreous quartzite. Beds average 40 to 100 centimeters thick, and most are horizontally laminated. Planar crossbedded strata 30 to 60 centimeters thick are common and define relatively tabular sedimentation units that pinch and swell somewhat across large outcrops. Some crossbeds have overturned tops, indicating deposition by strong, traction load currents. Paleocurrent directions from crossbeds at Bunker Hill are in accord with the observations of Hrabar (1974) and Bowden (1977), demonstrating a southerly source.

Some vitreous quartzite units contain 0.5 to 1.0 millimeter-sized carbonate grains that weather to limonite, staining outcrops dark brown.

Outcropping white, vitreous, horizontally laminated and crossbedded quartzite characterizes this rock type and distinguishes it from others of the Revett Formation.

SERICITIC QUARTZITE

Sericitic quartzite is pale green, slightly scratchable, and subvitreous. These properties result from a higher sericite content than in vitreous quartzite, probably reflecting more original interstitial clay within very fine-grained sand and coarse silt.

Sericitic quartzite forms beds that are mostly horizontally laminated, averaging 40 to 70 centimeters thick. Large-scale crossbeds are rare. In further contrast to vitreous quartzite, sericitic quartzite displays intervals of climbing ripple crossbeds, occasionally a meter or more thick.

Sericitic quartzite commonly exhibits penetrative limonitic stain with liesegang banding due to weathering of carbonate in surface exposures. Sericitic quartzite is mainly characterized by uniformly laminated, pale greenish, slightly scratchable, thick beds.

SILITE-ARGILLITE

Silite-argillite contains variably sorted thin layers of silt or very fine sand and argillite, but is predominantly silty. Beds are mostly 1 to 10 centimeters thick, primarily exhibiting horizontal and lenticular to wavy laminations.

Silite-argillite is characteristically soft and gray-green or yellowish green. In contrast to the idea that these greenish colors resulted from hydrothermal bleaching, we will later attempt to demonstrate that the greens of the Revett Formation resulted from diagenesis.

Although purplish argillite characterizes the St. Regis Formation, purple coloration is not entirely lacking within the silite-argillite of the Revett Formation. Mine geologists of the Bunker Hill Company have noted many examples of very minor purplish beds in drill holes and crosscuts within the silite-argillite units of the upper Revett member. The authors have seen purple beds in this part of the section in only one location, where the thickness of the beds totaled about 1 meter. A section of about 100 meters, just above the middle of the middle Revett member, contains many thin purple intervals. We have not examined any good exposures of these beds, but rare outcrops, float, and fairly abundant drill hole data within the Bunker Hill Mine indicate that purple beds are present within this interval over several kilometers distance, although apparently not everywhere.
Penetrative iron staining and liesegang banding are common in surface exposures and serve to distinguish these rocks from similar thin-beded, fine-grained lithologies in the St. Regis Formation, which contains brighter green as well as abundant purple layers. Underground, reddish surface stain due to post-mine weathering of disseminated carbonate is sensitive to specific lithology and restricted to the cleaner, coarser layers. This serves in places to emphasize sedimentary structures that are otherwise subtle and may also give this rock type a "tiger stripe" appearance that aids in its identification. Oscillation ripples occur on some bedding surfaces but are not so abundant as within silite-argillites of other Belt formations.

Thin, wavy mud inclusions that exhibit sharp edges in cross section are common within silite-argillite of the Revett Formation. The thin, apparently flexible and ductile original character of these inclusions distinguishes them from the thicker, more brittle mud chips or flat pebbles that are common in other Belt formations. The senior author has termed these features "mudskins," to differentiate them from "mud chips" which are the product of desiccation. "Mudskins" probably resulted from scouring of thin, slightly dewated, somewhat coherent mud layers.

In strong contrast to many other Belt formations, and also to the Revett Formation 40 kilometers to the north (Bowden, 1977), argillite layers of the Revett Formation in the Bunker Hill area contain essentially no desiccation cracks. This, plus the observation that the Revett Formation near Bunker Hill contains mudskins, rather than mud chips, indicates that Revett Formation silite-argillite, and perhaps the entire Revett Formation in this area, is the product of a continually moist depositional history.

In summary, the silite-argillite of the Revett Formation in the Bunker Hill area is characterized by thin bedding, horizontal to lenticular and wavy lamination, softness, greenish color, some oscillation ripples, and abundant mudskins.

**STRATIGRAPHY**

From good exposures along Big Creek (Figure 1), the Revett Formation falls handly into lower, middle, and upper members. Each of these, in turn, has a somewhat distinctive makeup of the three rock types described above.

The appearance of the Revett Formation in the Bunker Hill area is summarized in Figure 2. The upper Revett member of this section comes from the Bunker Hill Mine, where the senior author has examined it in considerable detail. The middle and lower Revett members do not lend themselves to direct measurement or continuous examination in the mine vicinity, although a few crosstabs and numerous drill cores make their general lithologies evident. Consequently, the appearance and thicknesses for these portions of the Revett Formation (Figure 2) are taken from sections that we have measured at Big Creek.

Vitreous quartzite within the Revett Formation forms stratigraphic units 30 to 130 meters thick. These tend to outcrop as resistant cliffs and "ribs," separated...
by mostly covered intervals of sericitic quartzite or silite-argillite that are tens or hundreds of meters thick. In the lower Revett member, vitreous quartzite is interlayered with intervals of sericitic quartzite. In the upper Revett member, on the other hand, vitreous quartzite separates intervals of silite-argillite, and sericitic quartzite is less abundant. The middle Revett member is predominantly silite-argillite, which forms covered slopes.

The middle Revett member is 330 meters thick. Because this considerable thickness of silite-argillite has been regarded as uncharacteristic of the Revett Formation, and because its green colors have been alternatively explained as the product of “hydrothermal bleaching” of purple strata, the middle Revett member has been locally confused with the St. Regis Formation, and possibly also the Burke Formation. Similarly, in different fault blocks in the Bunker Hill Mine area, the Revett-St. Regis formational boundary has been variably placed at the tops of a number of different vitreous quartzite intervals within the upper Revett member.

The lower, middle, and upper Revett members have been mapped throughout the Bunker Hill area, eastward through Big Creek and southward to Silver Hill. Several units in the lower portion of the upper Revett member appear to persist throughout this area, marking them as relatively tabular units.

The highest Revett units are not exposed at Big Creek. However, float as well as exposure in the nearby Crescent Mine indicates that several vitreous quartzite units prominent in this part of the section at the Bunker Hill Mine have changed southeastward to silite-argillite and lesser sericitic quartzite. The highest Revett beds are not well exposed at Silver Hill, but abundant vitreous quartzite float implies that the uppermost Revett member consists in this direction.

The changes described above underscore the problem that arises in attempting to define the bounds of the Revett Formation by its quartzite. In this case, the pinchout of the highest vitreous quartzite units at Big Creek would lower the Revett-St. Regis boundary, decreasing the thickness of the Revett Formation by several hundred meters over only 6 kilometers and correspondingly increasing the thickness of the St. Regis Formation.

Both a partial section of upper Revett strata at Pulaski Peak, 18 kilometers east of the Bunker Hill Mine, and the Revett and “St. Regis” sections in the Twin Crags quadrangle (Campbell and Good, 1963), 16 kilometers southwest, are similar in lithology and thickness to the upper Revett member near Bunker Hill. However, we have neither confidently correlated them nor attempted to correlate to any other sections outside of the Bunker Hill area.

Inadequate exposures and possible structural complexities preclude definite conclusions on possible lateral changes within the middle and lower Revett members in the Bunker Hill area. The lower Revett member appears to contain more sericitic quartzite and less vitreous quartzite in its lower portion at Silver Hill than appears in the Big Creek section. Alternatively, the Silver Hill exposures may correspond to even lower strata than those outcropping at Big Creek. The middle Revett member may be more quartzite at Big Creek than at the Bunker Hill Mine, but the authors are not in agreement with the interpretation which generates this conclusion.

**DIAGENESIS**

There has persisted in the Coeur d’Alene district the notion that widespread pre-ore, hydrothermal bleaching has changed rocks from purple to yellowish green, especially in broad areas surrounding orebodies (Weis, 1964). For example, because the Revett Formation is considered to be “quartzite,” intemtratified greenish “imure quartzite” or argillite has commonly been interpreted to be bleached, originally purple St. Regis lithology.

On the scale of a meter or less, contacts between purple and green rock cut irregularly across bedding. Because the green color clearly replaces the purple and because the orebodies are abundantly associated with green rather than purple strata, the secondary green color was assumed to be hydrothermal in origin.

However, boundaries between purple and green rock in the Coeur d’Alene district compare closely to those in the Missoula Group in western Montana, where there is no evidence of pervasive hydrothermal activity, but where greenish strata have been interpreted to be diagenetically reduced (Winston, 1973). Color contacts, which are irregular on a small scale in the Missoula Group strata, are subparallel to bedding on the scale of several meters or more, as is common in St. Regis strata in the Bunker Hill area.

In the Missoula Group (Winston, 1973), the reddish or purplish hues are due to oxidized, ferric iron in the form of hematite particles that surround terrigenous grains. The red argillic beds have abundant desiccation features such as mud cracks, salt casts, and dried mud chips, which reflect subaerial exposure and contact with free oxygen. Greenish layers in the Missoula Group do not contain hematite. Instead, they are colored by chlorite, which contains reduced, ferrous iron. Greenish strata show less evidence of subaerial ex-
posure, including notably fewer desiccation cracks, mud chips, and salt casts, and conversely more subaqueous synclasis cracks and water structures. From these and other stratigraphic observations, Winston (1973) has concluded that green Missoula Group strata were deposited farther out in the sedimentary basin, where they were more continuously covered with water and where ferric iron in limonitic sediments was reduced to ferrous iron. In short, green beds are the product of a wetter sedimentary environment than purple beds. The minor irregularities of the purple and green boundaries are more logically the result of local interaction between the diagenetic environments of adjacent beds.

The Revett Formation may be compared to the Missoula Group. The absence of desiccation features in the Bunker Hill area but the presence of abundant mud skins indicate a particularly moist environment, where sediments rarely dried out and became exposed directly to the air. Consequently, by analogy to the Missoula Group, the green beds of the Revett Formation are best explained as the product of early diagenetic reduction, rather than of "hydrothermal bleaching."

However, the present greenish color of Revett strata is not the direct result of early diagenesis, for chlorite is not conspicuous in thin sections. The yellow-green color apparently results from iron-magnesium-rich, "phenritic" illite, which is a product of late diagenetic or early metamorphism (Maxwell, 1974). The magnesite and ferrous iron of the phengitic illite were probably derived primarily from the diagenetic chlorite. At the same time, the occurrence of very minor intervals of purple strata within the Revett Formation demonstrates that late diagenesis-early metamorphism has not removed any hematite produced by early diagenesis from original limonite. Rather, the early diagenetic purplish and greenish rock color remained unchanged with increasing diagenesis and metamorphism.

ENVIRONMENT OF REVETT SEDIMENTATION

The tabular sheets of planar crossbeds and of the vitreous quartzite closely resemble the record of shallow, flat-beded, braided channels of humid alluvial fans (Boothroyd and Ashley, 1975). The horizontally laminated beds have the appearance of the plane-laminated beds of the Gum Hollow Fan (McGowan, 1971) and Bijou Creek (McKee and others, 1967; Miall, 1977), where flow is sheetlike and too shallow to form bars and dunes. The vitreous quartzite may therefore represent the distal part of a large alluvial fan, where shallow distributary channels became almost sheetlike before meeting the "Belt Sea," forming a fan delta (Figure 3).

The sericitic quartzite represents an accumulation of finer sand with entrapped clay, perhaps both adjacent to the main alluvial fan lobes or seaward from them (Figure 3).

The sediment of the silite-argillite accumulated in water that was occasionally placid but frequently subjected to waves and currents, perhaps in the shallow waters of the "Belt Sea" margin. Waves and currents repeatedly reworked the sediments into ripples and lenticular and wavy beds. Much of the clay component, so common in most Belt thin-bedded rocks, appears to have been winnowed from the silite-argillite beds by wave action or currents and transported elsewhere.

Vertical sequences of vitreous quartzite up to sericitic quartzite to silite-argillite are common in the Revett Formation, and they reflect transgression of the "Belt Sea" with decreasing current velocities over distributary sand lobes. Conversely, vertical sequences of silite-argillite up to sericitic quartzite to vitreous quartzite are equally common and reflect progradation of the fan delta lobes with increasing current velocities. Therefore, the Revett Formation near Bunker Hill represents repeated progradation and retreat of the alluvial fan, fan delta, and sea margin sediments. The sediments of the alluvial fan channels and sheetwash surfaces (vitreous quartzite) became interstratified with the fan delta and lobe-margin sediments (sericitic quartzite), which in turn became interstratified with the shallow "Belt Sea" sediments (silite-argillite). Tom Bowden (1977) has developed a similar interpretation for the Revett Formation from stratigraphic sections 50 kilometers north and northeast of the Bunker Hill area, although he prefers to interpret the equivalent silite-argillite lithology as the product of flood plain deposition.

Figure 3. Schematic block diagram showing facies patterns of Revett rock types.
The facies configuration of the Revett Formation closely resembles facies patterns in the middle part of the Missoula Group, for which there is a regional stratigraphic framework and for which a fan delta interpretation has been worked out in some detail (Winston, 1973; Winston and Jacob, 1977). Notably absent in the Revett Formation are the low angle foreshore crossbeds characteristic of large beaches and the channel deposits with herringbone crossbeds characteristic of tidal flats.

SUMMARY

Several conclusions of this paper ought to be emphasized. In the Bunker Hill area, strata of the so called "transition zone" of the Revett-St. Regis boundary contain intervals of vitreous quartzite, siliceous-argillite, and lesser sericitic quartzite that are representative lithologies of the Revett Formation as a whole. Greenish color within the Revett Formation is a characteristic of the finer grained Revett lithologies in this area and largely the result of diagenesis, not of hydrothermal bleaching. Consequently, the broad areas of hydrothermal bleaching that have been thought to surround most orebodies in the Coeur d'Alene district are instead areas of diagenetically reduced Revett Formation. In short, lead-zinc-silver orebodies of the district have preference for the Revett Formation proper.

Finally, the Revett Formation was deposited in a humid, predominantly fluvial environment and is unique among Belt rocks in this respect. Its unique depositional and diagenetic history are probably responsible for its affinity for ore.

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Structure of the Bunker Hill Mine, Kellogg, Idaho

by

Dwight S. Juras

After the deposition of the Belt sediments, which was probably completed by 850 m.y. ago (Obradovich and Peterman, 1968; Harrison, 1972), large-scale folds were produced along north-northeast trends (D1). Harrison (1972) correlated these folds from northern Idaho into British Columbia. In British Columbia, White (1959) and Leech (1962) dated a north-northeast-trending fold deformation at 700 to 800 m.y. ago. Representative examples of this deformation in the district include the Burke anticline, Mill Creek syncline, Montgomery and Moon Gulch anticlinorium, and possibly Pine Creek anticline (Figure 1). Although the first deformation appears more pronounced north of the fault, this deformation also occurs south of the Osburn fault, but it is overprinted by the structures that were produced in the second deformation. The second deformation produced large-scale folds along west-northwest trends (D2). Both sets of folds may be seen at the same locality. The north-northeast-trending folds have not been observed to curve into west-northwest-trending folds. The location of north-northeast-trending folds south of the Osburn fault can be determined by studying the plunge direction of west-northwest-trending folds. For example, west-northeast-trending folds in the Crescent and Sunshine mines plunge east but in the Bunker Hill Mine plunge west. A north-northeast-trending anticline is postulated between them. The age of the west-northeast-trending folds is uncertain, but researchers (Rolf Aadland, personal communication) in the Pend Oreille area do not indicate conspicuous west-northeast-trending folds in the Cambrian rocks.

In the district, west-northeast-trending folds are represented by the Big Creek anticline, East Fork anticline, Gold Creek anticline, Deadman syncline, and a newly named structure, the Tyler Ridge flexure (anticline). One of the exploration philosophies of the district geologists has been to look for ore in the north overturned limbs of folds, especially the Big Creek anticline.

The Bunker Hill Mine lies at and just north of the Tyler Ridge flexure which plunges 20 to 40 degrees west. This flexure appears as a large parasitic fold on a larger anticline to the south. This larger anticline may be an extension of the East Fork anticline. The axial trace of the Tyler Ridge flexure subparallels the Cate fault. Limonite-filled crinkle breccias occur at the hinge line in outcrop. Stereonet projections of the structure are displayed in Figures 2 and 3.

Figure 1. Coeur d’Alene district folds (incomplete). Modified from Hobbs and others, 1965.
Figure 2. Stereonet of fabric at the Tyler discovery. Dots—bedding, small circles—limonite-filled fractures, small x's—intersections, small triangle—measured axis, large X—girdle axis.

Figure 3. Stereonet of fabric in hanging wall of Cate fault (surface). Dots—bedding, small circles—limonite-filled fractures, small x's—intersections, small triangle—measured axis, large X—girdle axis.

Figure 4. Diagrammatic cross-section illustrates the fold fabric that localizes mineralization. Drawing by J. Farmin and A. Poeiz.

Many of the orebodies in the mine occur near or on the Cate fault. The March ore shoot was the largest and richest orebody in the mine and consisted of high grade galena with sphalerite, siderite, quartz, and pyrite. The true character of the orebody is obscure because most of the workings are inaccessible. The plunge of the March orebody is about 35° N. 76° W., which is subparallel to the axis of the Tyler Ridge flexure.

In Figure 4, a diagrammatic cross-section of the Tyler Ridge flexure displays the types of fold-related fabric along which mineralization occurs in the mine. Ore zones can consist of (a) reverse to right lateral faults (and brecias), which may or may not accompany smaller parasitic folds, (b) fracture fillings, which occur as fracture cleavage, extension fractures, and flexural cracks, and (c) disseminated zones in limb or hinge of fold.

Generally, however, the orebodies and mineralized fractures can be divided into two groups. The first group trends northwest, but due west to due north trends also occur (Figure 5). The apparent displacement on mineralized faults was different for each trend. Structures with north strikes have apparent right lateral offsets. Those with northwest strikes have apparent right lateral reverse offsets, and others with west-northwest strikes have apparent reverse offsets. Every gradational type of offset may have occurred. A compressive stress directed from the south-southwest is implied. Although fault slickensides are difficult to
observe on these structures, the slickensides do support the apparent drag directions of bedding. The intersection of flow and fracture cleavage with bedding defines the orientation of the axis of folding. The intersection of this first group of mineralized structures with bedding defines an area overlapping the axis of folding on the stereonet. The faults, which were previously described, generated on fold-related fractures. The same orientation of stresses that produced the west-northwest folds produced the ore-bearing faults.

The second group of faults and fractures strikes east-northeast to north-northeast and dips moderately southeast. Mine geologists refer to this group as "link veins," because these veins appear to extend between and be structurally related to two large faults. The origin of "link vein" fractures is uncertain, but their orientations are close to the joints of the folds. The apparent displacement on "link-type" structures is left-lateral reverse (Figure 6). The movement on these structures also appears related to the same stresses that produced the west-northwest folding.

These structural groups are characterized by contrasting mineralogy. The first mineralized group of northwesterly trend is conspicuously richer in siderite, pyrite, and sphalerite. Large orebodies, however, exist where galena replaces northwest-trending siderite-filled fractures, as in the March orebody. Galena, however, predominates in this group, in west-northwest-to east-west-trending reverse faults. The second group (link veins) is conspicuously richer in quartz and galena.

By using the method of cross-cutting fractures, a generalized sequence of mineralization is quartz, siderite, pyrite, sphalerite, galena, and quartz. This sequence indicates that the northwest trends were mineralized before the east-west and northeast trends, but exceptions and overlap of the sequence may be seen.

Both groups of mineralized fractures and faults are offset by bedding movements that are consistent with the fold form. In the mine, most bedding movements are south side down, because most of the mine lies in the north overturned limb of the fold. The folding outlasted mineralization, because mineralization rarely extends along bedding planes between offset mineralized fractures.

The stratigraphic compilation by Brian White in the mine indicates that over 1,300 feet of apparent right lateral offset is present along the Cate fault. The dip slip offset reconstruction is uncertain, but the 15 and 17 levels have been matched in one attempt. With this dip reconstruction, the apparent right lateral component is 1,600 feet and the dip component is 400 feet reverse. The reconstruction brings orebodies much closer together. This situation implies a postore age for the Cate fault, but its timing is still uncertain.

A great debate has existed in the district between hydrothermalists and syngeneticists. Orebodies of cer-
tain metal content appear to relate to certain formations of units, although most are controlled by structure. Evidence shows that the orebodies relate to the second folding. If the conclusion is correct, it is implausible to suggest that metamorphism during the second folding mobilized the metals when the metamorphic grade of both deformations is similar, that is, lower greenschist. The most noticeable difference in deformations is the production of wide, deep, and more intense fracturing and shearing during the second deformation. It seems more logical to suggest that the deep fissures and faults of the second deformation allowed the entrance of metal-bearing hydrothermal solutions.

The relationship between ore, structure, and stress, which were operating during the folding, are incompatible with a right lateral or left lateral Osburn fault. The right lateral movement is postore. If the Osburn fault existed during the west-northwest folding, its movement must have been reverse.

In summary, the orebodies in the Bunker Hill Mine are structurally controlled by the fold fabric of the overturned, west-northwest-trending D3-folds, specifically the Tyler Ridge flexure. The plunge of the west-northwest-trending folds is controlled by its position on the older north-northwest-trending D2-folds. Therefore, the metal-bearing fluids were introduced through the intense fractured zones during the D3-folding.

ACKNOWLEDGMENTS

This report is a result of research from 1975 to 1977 in the Bunker Hill Mine and on adjoining ground. Additional supporting evidence was also obtained from the Crescent Mine, the Dayrock Mine, and some reconnaissance throughout the district. The geology staff at the Bunker Hill Mine participated in this research, but special mention must be given to my coworker, Brian White, who compiled the stratigraphy in the mine and to Jerry Farmin, who initiated the research project. Special thanks must be extended to the Bunker Hill Company for allowing this presentation.

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Stratiform Mineralization and Origin of Some of the Vein Deposits, Bunker Hill Mine, Coeur d’Alene District, Idaho

by

V. M. Ramalingaswamy and E. S. Cheney

ABSTRACT

The Coeur d’Alene district, Idaho, is well-known for its vein deposits. However, several years ago mine geologists of the Bunker Hill Company noted stratiform deposits of sphalerite and argentiferous galena in thin-bedded, impure quartzites of the St. Regis-Revett transition zone. The sulfides occur in cross bedding, current ripple laminations, Bouma sequences, and structures produced by soft sediment deformation.

Stratiform mineralization is impoverished adjacent to veins in hand specimens. On a larger scale, impoverished zones that border veins in the “J” area of the mine are approximately twice as wide as the vein ore. Vein orebodies occur only where faults cut stratiform mineralization. Lead-zinc ratios of samples from veins vary from 5 to 3,000, but ratios of stratiform sulfides vary from 5 to 40. Lead-silver ratios of vein and stratiform samples are virtually the same. Evidently, stratiform lead and silver were preferentially mobilized into the veins. Vein ores have Precambrian lead-lead ages (Long, Silverman, and Kulp, 1959, 1960).

The stratiform Newgard ore is richer in zinc than the stratiform “J” mineralization. In the Newgard, the lead-zinc ratio increases stratigraphically upwards. The stratiform orebody that is currently being mined is about 1,100 feet long and approximately 200 feet thick, and it extends about 1,400 feet down dip.

The stratiform sulfide units can be used, at least locally, within the mine to delineate stratigraphy and structure, thereby enhancing the discovery of new vein and stratiform bodies.

INTRODUCTION

The Bunker Hill Mine is near the town of Kellogg in the Coeur d’Alene mining district of northern Idaho.

Virtually all of the geological literature on the district has been on the well-known vein deposits. Most authors have concluded that the veins are magmatic hydrothermal, but Long, Silverman, and Kulp (1959, 1960) have shown from lead isotopic studies that the vein lead is Precambrian. Stratiform lead-zinc deposits were recognized in the “J” area of the Bunker Hill Mine by the mine geologists as early as 1970 (Meyer, personal communication, 1973).

The purpose of the present study is (1) to describe these stratiform lead-zinc deposits and their host rocks and (2) to determine the stratigraphic, structural, and geochemical relationships of the stratiform ores to vein ores, because these may help in exploring for new stratiform and vein orebodies.

STRATIFORM MINERALIZATION AND SEDIMENTARY STRUCTURES

The stratiform sulfides (Figure 1; and Ramalingaswamy, 1975, Figures 12 and 13) occur in thin-bedded argillaceous or impure quartzites. Dark layers are mainly argentiferous galena; brown layers are mainly sphalerite. Mineralization participated in a variety of sedimentary structures (Figure 1). The thick-bedded pure quartzites contain very few sulfides. Sphalerite-rich stratiform ores tend to occur in thicker bedded impure quartzites, whereas galena-rich stratiform deposits of the “J” area occur in thinner bedded impure quartzites (Ramalingaswamy, 1975, Figures 10, 11, 12, and 13).

The host rocks consist mainly of quartz and highly birefringent, microscopic phyllosilicate, which is assumed to be sericite. In thin section the stratiform sulfides appear to be associated with finer grained and more sericitic laminae. Individual sulfide blebs are not parallel to bedding and do not have any recognizable sedimentary structures; the lack of any sedimentary features of individual grains probably is due to

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Figure 1. Galena and sphalerite occurrences:

(A) galena in parallel lamination. 23.5 level, "J" area, 23 stope;
(B) galena in cross bedding. 23.5 level, "J" area;
(C) galena and sphalerite in current ripple lamination. 9.1 level, Newgard area, 27 stope;
(D) galena and sphalerite in Tadde division of Bouma sequence. (a) a lower graded layer, (b) a lower zone of parallel lamination, (c) a zone ripple bedding, (d) an upper zone of parallel lamination, and (e) an uppermost clay layer. 9.1 level, Newgard area, 27 stope;
(E) galena in soft sediment deformation. 23 level, "J" area;
(F) sphalerite and galena in soft sediment deformation. 9.1 level, Newgard area, 25 stope.
recrystallization during metamorphism. The sulfides do not appear to replace sericite or quartz.

STRATIFORM MINERALIZATION AND ITS RELATIONSHIP TO VEIN-TYPE ORES

A zone of impoverishment of stratiform mineralization commonly borders the fractures (Figures 2 and 3). In some hand specimens, sulfides in fractures occur at the intersections of bedding planes containing sulfides with the fractures (Figure 3). The vein sulfides are considerably coarser grained than the stratiform sulfides (Figure 2).

The same characteristics occur on an orebody scale. In the "J" area, stratiform sulfides in the country rock become discontinuous, spotty, and absent adjacent to a fracture or a vein (Figure 4). If mining operations were not restricted to the "J" vein, more continuous stratiform sulfides probably would be found. Most veins containing sulfides and quartz occur on the same side of a fault as the stratiform sulfides (Figures 4 and 6). The impoverished zones are considerably richer in sericite than the original host rock. This suggests that the quartz originally present in the depleted zones may have migrated to the veins.

Of course, historically, the sericitic zones with low values of metal have been considered hydrothermal alteration envelopes that grew from the veins outward and introduced sulfide minerals into the wall rocks. The presence of relict stratiform sulfides and the forementioned smaller scale examples in hand specimens show that this is not the case.

Figure 3. Sulfides occurring at the intersections of bedding planes with the fracture. 23.5 level, "J" area, 23 stope, floor 4.

In contrast to the "J" area, very well-bedded, richer stratiform sphalerite deposits occur in the Newgard area (Ramalingaswamy, 1975, Figures 13, 14, and 15). The tabular orebody is approximately 1,100 feet long and has a stratigraphic width of about 200 feet (Figure 5). The orebody extends approximately 1,400 feet down dip. Even in large fault zones, sphalerite-bearing veinlets are more common than mineable veins.

METAL RATIOS OF THE OREBODES

The lead-zinc ratio of individual samples of the stratiform deposits of the "J" area vary from 5 to 40. The impoverished and semi-impoveryed zones have

Figure 2. (a & b) Impoverishment of stratiform sulfides bordering vein. 21 level, 21 stope, and 23.5 level, 23 stope. "J" area.
variable lead-zinc ratios, generally from 5 to 200 (Tables 1 and 2; and Ramalingaswamy, 1975, Tables 6 and 7). The ratios of percent lead to ounces per ton of silver in both the stratiform and impoverished zones usually vary from 1:1 to 2:1 (Tables 1 and 2).

Thus, the main difference between stratiform mineralization and "J" vein ores is not mineralogical but the absolute and relative amounts of sulfides present. The vein ores have lead to zinc ratios from about 5 to greater than 3,000, whereas the lead to silver ratios are virtually the same as those of stratiform sulfides (Tables 3 and 4; and Ramalingaswamy, 1975, Tables 8 and 9).

In the "J" area the amount of lead, silver, and in some places copper decreases close to a vein (Figure 4; Tables 1 and 2; and Ramalingaswamy, 1975, Figure 6, Tables 4 and 5). The higher ratios of lead-zinc, silver-zinc, and copper-zinc evidently are due to preferential migration of lead, zinc, and copper into fractures and faults.

The stratiform samples of the Newgard orebody are considerably richer in zinc than those of the "J" area. The lead-zinc ratios of bulk samples generally are less than 1 (Figure 5), and the ratios of percent lead to ounces per ton of silver vary from 1:1 to 2:1. The Newgard orebody tends to have higher lead to zinc ratios in its stratigraphically higher portions (Figure 5). Where cut by faults, the orebody tends to have lead to zinc ratios greater than 2 (Figure 5; and Ramalingaswamy, 1975, Figures 16 and 18): perhaps lead preferentially migrated into faults in the Newgard orebody, as it did in the "J" area.

GENESIS OF COEUR D'ALENE ORE DEPOSITS

Because stratiform sulfides participated in the deformation of soft sediment, they must have been introduced quite early in the history of the strata. The characteristics of the stratiform ores in the Bunker Hill Mine are similar to three of the four characteristics described by Renfro (1974) for Sabbath ores: (1) associated with sandstones interbedded with muds and some carbonates (these carbonate beds contain stromatolites); (2) laterally and vertically zoned with respect to metal content; (3) underlain by clastic sediments; and (4) overlain by strata that contain evaporite sequences of dolomite, gypsum, anhydrite, and halite. Furthermore, Rezak (1957) and Harrison (1972) mention the occurrence of stromatolites in the less metamorphosed part of the Raveli Group in Montana. So far, no evaporites have been recognized in the St. Regis-Revett transition zone; if they were ever present, they must have been destroyed prior to or during greenschist facies metamorphism. Incidentally, Clarke (1971), Harrison (1972), Trammel (1975), and Lutz (1977) have described

DELINEATION OF FAVORABLE STRATIFORM UNITS

The stratiform units of the "J" area in the 21, 23, 25, and 27 levels can be used to calculate the displacements on the Knugor fault and probably the Dull fault. On the 21 and 23 levels, the stratiform units occur in the footwall of the southwest-dipping Knugor fault (compare Figures 6 and 7). On the 25 level, the units occur in the hanging wall as well as in the footwall, and on the 27 level they are on the hanging wall. A southward-plunging antiform containing stratiform units occurs in the footwall of the Knugor fault on the 25 level at W5200:N1300 and in the hanging wall on the 27 level at approximately W4100:S2000 (Figure 7). From these relationships, one can calculate approximately 600 feet of vertical displacement and about 1,200 feet of left lateral displacement. The Newgard orebody is faulted exactly like unmineralized strata.

Figure 4. Pockets of stratiform sulfides in impoverished zones bordering vein. The two profiles show the decrease in the content of lead towards the vein.

23.5 level, 23 stope, "J" area.
Table 1. Stratiform samples collected along strike towards vein. 21.5 level, ‘J’ area, 22 stope.

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Distance From Vein in Feet</th>
<th>Pb %</th>
<th>Zn %</th>
<th>Ag oz/ton</th>
<th>Cu %</th>
<th>Pb:Zn</th>
<th>Pb:Ag</th>
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</thead>
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Table 2. Stratiform samples collected along strike towards vein. 21.5 level, ‘J’ area, 22 stope.

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Distance From Vein in Feet</th>
<th>Pb %</th>
<th>Zn %</th>
<th>Ag oz/ton</th>
<th>Cu %</th>
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Table 3. Samples collected in a vein. 21.5 level, "J" area, 22 stope.

<table>
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<th>Pb %</th>
<th>Zn %</th>
<th>Ag oz/ton</th>
<th>Cu %</th>
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LEAD TO ZINC RATIOS

- less than .5
- 5 to 1.0
- 1.0 to 1.5
- 1.5 to 2.0

MINE DRIFTS
BULK SAMPLE LOCATIONS
FAULTS
LIMITS OF THE ORE BODIES

0 100 FEET

Figure 5. Newgard orebody showing geology and zonation of lead-zinc. 9.1 level, Newgard area.
Table 4. Samples collected in a vein, 21.5 level, "J" area, 22 stope.

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Pb %</th>
<th>Zn %</th>
<th>Ag oz/ton</th>
<th>Cu %</th>
<th>Pb:Zn</th>
<th>Pb:Ag</th>
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</table>

Figure 6. Geology of the 21 level, "J" area.
stratiform copper deposits in the Revett and Empire Formations of the Belt Series of northern Idaho and western Montana.

The impoverished zones, the coincidence of vein ores within or adjacent to stratiform deposits in the "J" area, and the lead isotopic data (Long, Silverman, and Kulp, 1959, 1960; Zartman and Stacey, 1971) suggest that at least some of the vein deposits of the Coeur d'Alene district were derived from the Beltian stratiform deposits.

We have seen examples of stratiform sulfides in the Lucky Friday, Sunshine, Galena, Star-Morning, and Crescent mines in the Silver Belt of the eastern portion of the district. Although the ores of these mines are richer in silver and copper than the Bunker Hill, the host rocks are the same St. Regis-Revett transition zone. It is tempting to speculate that the vein ores of many of the Coeur d'Alene mines might have been derived from a single, laterally zoned, sulfide-bearing stratigraphic interval that raveled in tonnage but not in grade the Aldrich Formation (equivalent to the Prichard Forma-

tion of the United States) of the Sullivan Mine at Kimberley, British Columbia, Canada.

ACKNOWLEDGMENTS

The authors are indebted to the Bunker Hill Company for supporting the senior author's field work. We thank the staff geologists, especially Mr. R. L. Meyer and Mr. J. L. Farmin for their assistance, discussions, and criticism. The Society of Sigma Xi and North Pacific Section of A.I.M.E. provided additional financial support.

REFERENCES


Petrography of Stratiform Lead-Zinc-Silver Deposits at the Bunker Hill Mine, Idaho, with Some Thoughts on the Timing of Ore Deposition

by

Rolland R. Reid

Bedding-controlled sulfides occur adjacent to a number of veins in the Bunker Hill Mine. Sulfide distribution appears to be controlled by bedding layers, cross-bedding, penecontemporaneous slump contortions, and intratamformational conglomerates or breccias. These rocks were subjected to petrographic study.

Quartz is a major constituent, in 0.3-0.5 millimeter granoblastic grains; all grains display strong undulatory extinction. Many show deformation lamellae. Most show microstratification along intergrain contacts and particularly strong undulatory extinction at and near intergrain contacts; this does not occur or is less pronounced where sericite is in the quartz-quartz contacts. In places on quartz-quartz boundaries, 0.01 millimeter subgrains of quartz develop and are very intensely strained. Some quartz-quartz boundaries are strongly in block-form sutures, and the adjacent "grains" are mosaics of strongly strained subgrains. Tiny, fresh carbonate rhombs also grow in the quartz-quartz boundaries. No vestiges of original sedimentary texture remain; the quartz grains are tightly intergrown and interlocking granoblastic mosaics. No remnants of original pore space were noted. In places, irregular zones of cataclastic microbreccia extend among the quartz grains in an irregular network of intergranular cataclastic crushing. Quartz is about 75 percent of the rock.

Siderite constitutes about 10 percent of the rock. It occurs in irregular granoblastic grains 0.5-1.0 millimeter in diameter. Sericite constitutes about 10 percent of the rock in an irregular network of grains showing no preferred orientation. Clay is minor in amount. Aggregate about 0.5 millimeter in diameter are made of 0.02 millimeter grains. These aggregates are about 10 percent altered to sericite. A few 0.03 millimeter grains of galena occur in the clay aggregates. Thorite and zircon are minor accessories.

Galena occurs in 0.02 millimeter grains which lie in quartz-quartz boundaries, suggesting that they nucleated there. Some irregular galena aggregates partly replace and enclose sericite flakes and tiny quartz grains and penetrate sericite along its cleavage, very strongly in places. Galena also penetrates quartz grains along cataclastic fractures. Larger galena grains (0.3 mm by 0.02 mm) extend in irregular formations along the quartz-quartz boundaries and display very irregular contacts against the quartz grains. Galena shows irregular replacement contacts against detrital zircon. Local "pseudopods" of galena extend from larger grains out along intergrain boundaries. Galena extends replacement "pseudopodia" into zones of deformation lamellae in quartz grains and is therefore later than the deformation lamellae. The deformation lamellae are associated with zones of cataclastic shearing. Galena occurs in zones of cataclastic microbrecciation and yet displays very thin, delicate, uncrushed filaments enclosing patches of crushed, cataclastic quartzite. Galena is about 5 percent of the rock.

Crossing extension fractures are up to 1 millimeter thick. Quartz grains grow within the extension fractures with their long axes perpendicular to the fracture walls. Microscopically, the fracture walls are very irregular. Galena (no sphalerite) grows in fairly regular grains among the quartz grains. Galena is markedly depleted in the wall rock of the fracture over a zone 2 to 3 millimeters wide enclosing the fracture. Some galena grains are irregular but still show serrated contacts against quartz consisting of the 90-degree angles of galena crystal faces. Grains are much larger, up to 0.75 millimeter, than those in the matrix.

Sulfides occur in several other ways in these rocks. Schistosity cuts the bedding at all angles. The rocks are pervasively cataclastic, and cataclastic shearing paralleled the preexisting schistosity. Megascopically, the appearance in most of the specimens is that of layer-concentrated sulfides parallel to bedding.

Viewed microscopically sphalerite occurs as elongate grains with the long axes parallel to the schistosity; blebs of galena are similarly arrayed. Sphalerite encloses metamorphic sericite. Coarse lenticular sphalerite and coarse pyrite blebs are oriented with their long axes parallel to the trace of the schistosity. Both minerals enclose flakes and patches of cataclastically deformed

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quartz. Galena is finer grained than the sphalerite and subordinate to it, but galena is also in elongate blebs parallel to the schistosity and strung out like beads along it. Galena encloses metamorphic sericite and appears partly to replace sphalerite. In places, sulfides intricately invade and replace cataclastic quartz in quartz-filled extension fractures. Sulfide grains embay, replace, and enclose cataclastic quartz grains. In one rock, sulfide grains appear megascopically to be concentrated along ripple-marked bedding. Schistosity cuts the ripple-marked structures at about 60 degrees, and the sulfide grains microscopically are seen to be lenticular and parallel to the schistosity and strung out along it. Heavy concentrations are confined to certain beds. In the intraformational conglomerates, "clay" fragments occur up to 4 millimeters long. Differential sulfide concentrations exh out the various complex sedimentary features, but only tiny bits of sulfides are deposited within clay grains. Schistosity paralleled by cataclastic shears cuts across these features, and sulfides are lenticular, parallel to the schistosity and strung out along it. Cross-bedding ech out by sulfides displays similar microscopic relations.

Quartz-filled extension fractures are cataclastically folded, and sulfides were deposited along the cataclastic schistosity, axial planar to the cataclastic shear folds. The sulfides replace and enclose cataclastic grains of quartz. Dark layers are dark principally because of sulfide enrichment, but in some of them heavy detrital minerals also occur. One example studied by electron microprobe showed rutile, zircon, tourmaline, and monazite.

The rocks have been metamorphosed in the green-schist facies transition zone, recrystallizing quartz to a granoblastic texture and producing metamorphic sericite. Later came bedding faulting and associated strong cataclastic crushing and microbrecciation of the quartzite. Sulfides penetrated the schistosity-parallel cataclastic shear fractures, depositing quartz, carbonate mineral, galena, and sphalerite. As galena hardly grew in the clay grains, it appears that galena nucleates poorly on clay. Galena was remobilized and moved 1 to 2 millimeters to be redeposited in opening extension fractures. This resulted in a "depleted" zone in the walls adjacent to the fractures. These fractures are perhaps due to tectonic relaxation following the cessation of cataclastic deformation.

In all the rocks studied, schistosity defined by fine-grained metamorphic sericite cuts across the primary sedimentary features such as bedding, ripple mark, intraformational conglomerate, and cross bedding. Furthermore, penetrative cataclastic texture is present, and it appears that penetrative cataclastic shearing has occurred parallel to the preexisting schistosity in most if not all the specimens. This can be explained in several ways. First, deformation outlasted metamorphism in a single orogenetic event so that shearing persisted and became cataclastic after recrystallization had ceased. Second, the rocks were subjected to a later deformation after metamorphism and perhaps much later in time, and the forces of the later deformation were resolved in such a way that cataclastic shearing occurred parallel to the weak direction in the rocks created by the existing schistosity. Third, the relations are accidental in the rocks studied. Fourth, tectonic forces that produced the cataclastic schistosity were oriented in the same direction as those deduced from the other indicators. Since all the features are produced in structures that are more or less parallel to the Osburn fault, a good case can be made for the fourth hypothesis. Normal stresses on the Osburn fault will always have the same orientation whether the fault is sheared left or right laterally or is subjected to the same sense of shear at different times. The normal stress is likely responsible for the generation of folds whose axes and axial planes are subparallel to the fault. Axial-plane shear in such folds will be partly of strike-slip character. Thus, an axial-plane shear generated a sericite schistosity during Precambrian metamorphism involving either one or two events (Reid, R. R., W. R. Greenwood, and G. L. Nord, Metamorphic petrology and structure of the St. Joe area, Idaho, manuscript in preparation). Later, in Laramide slip, the axial-plane shear direction was reactivated to produce cataclastic shearing parallel to the older sericite schistosity.

Deposition of sulfides by replacement occurred along the schistosity, as it was the most pervious direction. The schistosity is cataclastic; that is, cataclastic shearing occurred parallel to the metamorphic schistosity at a later time. Sulfides diffused along the pervasive cataclastic shears to deposit sphalerite and galena in different proportions in different rocks. The appearance is one of a pervasive "soaking" of the rock by the sulfides moving along the closely spaced (0.05 mm) cataclastic shear planes. Some bedding layers favored sulfide deposition more than others for unknown reasons, and they precipitated more sulfides than did others. The result is an apparently differential concentration of sulfides along bedding layers, closely resembling sedimentary deposits when viewed megascopically. In a similar way, features of ripple mark and cross bedding were also accentuated by layer-parallel sulfide concentrations.

The question of age relation of sulfide introduction to cataclastic shearing remains unanswered. It may be that mineralization was contemporaneous with shearing or that it was entirely later. Because the rock will have been most pervious to solution flow along the cataclastic shear planes during the time that shearing was
active, the time of sulfide introduction most likely coincides with the time of shearing. This leads to the hypothesis that older vein sulfides were reactivated by stress and moved by diffusion out of the veins into the wall rock along the cataclastic shears. The fact that sulfides replace and enclose grains of cataclastic quartz places an upper limit on the temperature of the mineralizing solutions—that of the recrystallization of quartz under the P-T stress conditions existing in the rock at the time of mineralization.

Since the cataclasis apparently governs the age of the stratiform mineralization, the age of the cataclasis needs to be considered. This is not certainly known for the Coeur d'Alene region, but from a regional point of view it might be the same age as cataclasis in the St. Joe area 70 miles to the south. The St. Joe fault is a strike slip fault parallel to the Osburn fault and has had perhaps much the same history as the Osburn fault. According to Reid and others (in preparation), cataclasis at the St. Joe fault postdates an intrusion of Idaho batholith character and predates andesite and lamprophyre dikes of early Tertiary age. Thus, the cataclasis is assigned to the Laramide right-lateral movements on the St. Joe fault. The Osburn fault has also undergone right-lateral shear in the Laramide, which is the age assigned to the cataclastic event in the rocks adjacent to the Osburn fault.

If the veins are to be pre-Laramide, then one may speculate about that age. I have seen a contact between one of the Gem stocks and a mainstage vein in the Hercules Mine. The vein was folded near the contact in a manner consistent with lateral spreading of magma at the time of intrusion. This seems to show clearly that the veins are pre-Gem stocks, which are themselves pre-Laramide. Now, the question becomes: where in pre-Gem stock time were the veins emplaced?

Reid and others (in preparation) show a major bleaching event in the St. Joe area at 300 m.y. (K-Ar date). This consisted of extensive fine-grained feldsparization of dark argillites to dull white feltes along many northwest-trending faults. Cutting the bleached rocks and parallel to the faults are scattered quartz-siderite veins with trace galena and sphalerite. Filled extension fractures of the same age show left-lateral strike-slip movement during mineralization, comparable to that recognized in the Bunker Hill Mine (Footwall Francis) and the Galena Mine (290 Vein). This mineralization is probably similar in age to the hydrothermal bleaching and also about 300 m.y. old.

Therefore, the Coeur d'Alene mineralization may also have occurred at 300 m.y. Possibly supporting this view is a potassium-argon age of 477 m.y. ± 18 m.y. on argillite in the wall rock of veins in the Bunker Hill Mine. This age may represent an incomplete resetting of a Precambrian metamorphic age during the 300 m.y. event by hot hydrothermal solutions that introduced the sulfides into the vein.

Consequently, to understand the genesis of the stratiform sulfides in the Bunker Hill Mine, the following hypothetical history has been formulated:

1. Precambrian metamorphism and folding. Normal stresses on the ancestral Osburn fault created folds parallel to the Osburn fault; concurrent metamorphism generated an axial plane cleavage or schistosity in these folds. Two Precambrian fold or metamorphic events are possible. If a second one occurred, the normal stresses on the Osburn fault reactivated the existing folds that lie parallel to the fault and deformed them further along the same structural paths as previously.

2. Paleozoic left-lateral movements along the ancestral Osburn fault zone. Normal stresses in this event reactivated the folds adjacent to the fault that were parallel to it, and hydrothermal sulfides were deposited, controlled both by fault-related fractures and by various fold elements (cf. Juras, this volume). Again, movement in the folds paralleled those of the earlier events because of the parallelism of the driving (normal) stress. One may speculate, though, that the folds reacted in a more brittle way than previously, because no metamorphism was operative at this time.

3. Right-lateral Laramide movements, partly cataclastic in character, in the operation both of normal stress and shear stress in the Osburn fault zone. The normal stress, of course, had the same orientation as that which occurred in the previous faulting events, one or more of which may have been left lateral. Because the axial planes of the fault-parallel folds are more or less parallel to the fault, the reactivated folds underwent cataclasis more or less parallel to their axial planes and thus parallel to their metamorphic axial-plane schistosity. Certain of the sulfides in the veins (principalally galena and sphalerite) were partly activated by stress and thus caused to move by surface diffusion into the rocks, along the active cataclastic shears. Preferential precipitation by certain bedding layers resulted in stratiform sulfides in the wall rock. The stratiform sulfides, however, are derived from the veins, not the other way around.
Disseminated Galena in the Lucky Friday Mine, Coeur d'Alene District, Idaho

by

C. E. Hauntz

ABSTRACT

Disseminated galena is found up to 245 meters away from the lead-zinc-silver vein of the Lucky Friday Mine, Mullan, Idaho. The galena occurs in several discrete zones in quartzite of the Beverett and St. Regis (9) Formations. These zones occur from the 2,150-foot level to the 4,450-foot level.

The contact relationships between galena-bearing quartzites and other rocks on the 4,250-foot level depend on the lithologies involved. In general, there is a sharp contact with argillaceous rocks and a gradational contact with other quartzites. These mineralized zones appear stratabound and, in some places, stratiform at a scale of several meters, but these relationships do not appear to hold for the entire sequence.

The mineralized quartzites consist of quartz (86-93 percent), sericite (3-11 percent), ankerite (1-2 percent), and trace amounts of clay minerals. Assays of the galena-bearing quartzite average 1 ounce per ton of silver. The silver probably occurs in the galena.

Cataclastic shear planes spaced about 0.1 millimeter apart within the mineralized quartzites are most readily identified by the alignment of intragranular, metamorphic sericite grains, many of which are bent. Additional evidence for the presence of these shear planes is supplied by the galena, which penetrates and replaces the quartz grains along these planes.

Three types of quartz veinlets crosscut the galena zones. The first type consists only of quartz. The second type contains quartz and galena with a 2 to 5 millimeter selvage of quartzite leached of galena on both sides. The third type contains quartz, galena, and siderite and does not affect the disseminated galena.

The absence in the quartzite of sphalerite, chalcopyrite, tetrahedrite, pyrrhotite, pyrite, and arsenopyrite is significant because of their presence in the nearby vein. Their absence makes it seem unlikely that the vein was derived by lateral secretion of the intermediate wall rock. For the same reason, hydrothermal leaching of the vein and subsequent precipitation of disseminated galena in the quartzite is also unlikely.

Mechanical remobilization of galena from the Lucky Friday vein along cataclastic shear planes may be the process of emplacement of the disseminated galena in the quartzite.

University of Idaho, Moscow, Idaho 83843.
Characteristics of Lead-Zinc-Silver Veins Located in Belt Rocks to the North of the Coeur d’Alene District

by

William R. Green

INTRODUCTION

Because of its significance, the Coeur d’Alene district has long been studied by economic geologists as is evidenced by the papers presented at this meeting. The studies have added greatly to the body of knowledge concerning deposits of this type. However, the magnitude of the district, its structural complexities, and the possibility of multiple stages of deposition have made the age and origin of the deposits controversial.

The western portion of the Belt basin contains numerous occurrences of lead-zinc-silver fissure veins which in many ways are strikingly similar to those of the Coeur d’Alene district. This paper is an attempt to identify the characteristics of these less complex occurrences in the hope that the information may be useful in deciphering some of the mysteries of the Coeur d’Alene district. Thus, thirty-eight of the more important occurrences in northern Idaho and northwestern Montana were examined. Data used in preparation of this paper were taken largely from published sources, but have also been augmented by the author’s field experience in the region.

GEOLOGIC SETTING

Most of northern Idaho and adjacent portions of northwestern Montana are underlain by the Belt Supergroup. These predominantly clastic sediments exhibit a low-grade regional metamorphism which increases slightly with stratigraphic depth. Very little Phanerozoic sedimentation is found in the geologic record of the region. However, a few isolated occurrences of Middle Cambrian quartzite and limestone have been preserved by block faulting.

Intrusive rocks including dikes, sills, stocks, and batholiths occur throughout the Belt basin. Sills and dikes, having generally a dioritic composition, were intruded into the Belt Supergroup at several times during the Precambrian. Stocks and small batholiths are widely scattered throughout the region but become much more prevalent along the Kootenay Arc which borders the study area on the west. These plutons consist predominantly of granodiorite, quartz monzonite, and syenite and are considered to have been emplaced in late Mesozoic as well as Tertiary time.

The Belt rocks that lie to the north of the Coeur d’Alene district are characterized by broad, gently north-trending folds that probably formed during the late Precambrian East Kootenay orogeny (Harrison and others, 1974). However, the most striking structural characteristic of the study area is a complex fracture pattern which in part originated in the Precambrian and has been reactivated by succeeding tectonic adjustments.

Structure within the Belt basin is dominated by the west-northwest-trending system of tear faults which were collectively referred to as the Lewis and Clark Line by Billingsley and Locke (1935) (Figure 1). The Osburn fault, a major element of this system, cuts the Coeur d’Alene district and thus forms the approximate southern border of the study area.

The Hope fault is similar to the major tear faults that make up the Lewis and Clark Line. However, near the center of the basin it splits from the main trend and extends for more than 200 kilometers in a northwesterly direction across the study area. The Hope fault most probably originated in late Precambrian and has had a long history of recurrent movement resulting in a net right-lateral displacement of 10 to 15 kilometers. Harrison and others (1974) have conclusively shown pre-late Mesozoic and Cenozoic right-lateral offsets as well as substantial Cenozoic dip-slip displacement along the Hope fault. Mapping by the author in the Pend Oreille Lake area to the south of the Hope fault has identified several west-northwest-trending fractures that exhibit displacement of similar age and sense (Green, 1976). However, two faults shown to have formed in the late Mesozoic display a substantial left-
LEAD-ZINC-SILVER DEPOSITS

GENERAL DESCRIPTION

The Belt basin north of the Lewis and Clark Line contains many scattered mining districts noted for their lead-zinc-silver veins. The districts generally contain one or two larger deposits surrounded by a number of much smaller occurrences. The maximum recorded production from any single deposit has been approximately 650,000 tons, and only a few deposits have yielded more than 100,000 tons. Although the tonnage yields have not been great, silver values have been sufficient to make the production from the larger deposits economically significant.

The deposits have many similarities with each other and with those of the Coeur d'Alene district. The veins occur as shoots and lenses along individual fractures or in wider shear zones. In the more important deposits, the shoots have been traced along strike as well as down dip for more than 500 meters. Maximum widths range up to 10 meters, but generally are less than 2 meters. Both megascopic and microscopic vein textures reveal that open space filling is the primary method of mineral deposition. This conclusion is supported by the localization of ore shoots in the open spaces created by movement along faults having a variable attitude due to refraction as well as in the open spaces created by cross fracturing. Aside from the importance of lithologic character on the attitude and degree of fracturing, the formations that comprise the Belt Supergroup appear to have little effect on mineral deposition. Of the deposits studied, no preference was shown for particular stratigraphic horizons.

The lead-zinc-silver veins are all characterized by relatively simple mineralogy, which includes metal sulfides as well as nonsulfide gangue material. Principal vein constituents include quartz, siderite, galena, sphalerite, pyrite, chalcopyrite, and tetrahedrite. Generally these minerals were deposited in stages which represent a change in the character of mineral-bearing fluids with time. The banding of vein filling as well as the texture of polished sections show that repeated fracturing created open spaces for the deposition of succeeding stages. Because the recurrent movement along vein fractures was not uniformly distributed, all stages are not present in every deposit. Also, detailed paragenetic information is not available for many deposits that show multiple stages of deposition. However, existing data fit a paragenetic sequence for the main vein-forming minerals as follows: quartz, siderite, pyrite, arsenopyrite, sphalerite, tetrahedrite, galena, and chalcopyrite.
A typical paragenetic example might be the paragenesis for the ones of the Talache Mine in northern Idaho, shown in Figure 2. (Note that quartz is deposited in several stages, which is another characteristic of the lead-zinc-silver veins.)

Hydrothermal alteration is common but not extensive about the mineral deposits in the study area. Alteration may occur for as much as 20 meters into adjoining wall rock but in most places extends for less than 2 meters. These effects consist of a pronounced lightening of the preexisting color of the host rock due to loss of pigmentation minerals and some increase in sericite, siderite, and quartz content.

<table>
<thead>
<tr>
<th></th>
<th>Stage I</th>
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<td>POLYBASITE</td>
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Figure 2. Generalized paragenetic sequence of hypogene mineralization.

DIFFERENCE IN DEPOSITS

Although deposits in the study area have many similarities with each other, they differ in one important respect. Zartman and Stacey (1971) determined that the lead in the veins was either Precambrian or Mesozoic-Cenozoic in age. These varieties will be referred to as old lead and young lead for the remainder of this paper. Zartman and Stacey found that in most of the Coeur d'Alene district, as well as in a portion of the occurrences tested in the study area, the lead was of the old variety, while the remainder of the occurrences in the study area contained young lead. A careful examination of the lead-zinc-silver veins shows that both types of occurrences have certain distinguishing characteristics, such as differences in metal abundance and the structural setting in which they are found.

RELATIVE METAL ABUNDANCE

Although the mineralogy and paragenesis of the major vein-forming minerals is similar across the study area, the relative metal abundance, reflected in metal ratios, varies substantially. Of special significance are the silver-to-lead-zinc and the lead-to-zinc ratios as expressed in ounces and percentage. The metal ratios can be approximated from past production or sampling records. However, care must be taken that the figures are truly representative and not biased by such factors as mineral zoning or incomplete reporting of subeconomic values.

Although some overlap occurs in ranges of values, the data show that young lead deposits generally have much different metal ratios than those containing old lead. In most places, the deposits containing young lead have a silver to lead-zinc ratio much greater than one, and a lead-to-zinc ratio of about one. For deposits containing old lead, the silver to lead-zinc ratio is much less than one, and lead exceeds zinc by a wide margin.

These relationships are shown in Table 1 which compares the average metal ratios for several districts in the study area and those of the Coeur d'Alene district. The "other" category includes the averaged ratio for a number of districts in northern Idaho and northwestern Montana that contain old and young lead. It should be noted that the figures for the Coeur d'Alene district are based on all-time production records and do not reflect higher silver to lead-zinc ratios which are found along the Silver Belt.

In addition to their much higher proportion of lead to silver or zinc, many veins with old lead contain one or more accessory minerals that are diagnostic. These minerals include pyrrhotite, marcasite, tremolite, and

<table>
<thead>
<tr>
<th>Age of Lead</th>
<th>Metal Ratios</th>
<th>Zn: Pb</th>
<th>Primary Controls</th>
<th>Vein Fracture</th>
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<tr>
<td>Mesozoic-Cenozoic</td>
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<tr>
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<td>8:1</td>
<td>2:1</td>
<td>B</td>
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<tr>
<td>Lakeview</td>
<td>6 1/2:1</td>
<td>3:1</td>
<td>S</td>
<td></td>
</tr>
<tr>
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<td>1:1</td>
<td>1:1</td>
<td>B-SS</td>
<td></td>
</tr>
<tr>
<td>Others</td>
<td>2:1</td>
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<td></td>
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<tr>
<td>Coeur d'Alene</td>
<td>1:1</td>
<td>1:1</td>
<td>SS</td>
<td>F</td>
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<tr>
<td>Others</td>
<td>1:4</td>
<td>1:2</td>
<td>SS</td>
<td>F-S</td>
</tr>
</tbody>
</table>

1Letters refer to block faults (B), strike slip faults (SS), fissures (F), and shear zones (S).
2Metal ratios for the Coeur d'Alene district represent an average based upon total recorded production.
Tourmaline. Minor pyrrhotite has been reported in a few deposits that contain young lead and cut pyrrhotite-rich portions of the Prichard Formation. However, these accessory minerals are not reported in any other deposits containing young lead.

STRUCTURAL SETTING

Most lead-zinc-silver veins are found in fractures and shear zones which are subsidiary to major faults. Furthermore, these veins are associated with particular types of major faults depending upon the age of the lead.

Practically all deposits with young lead are associated with block faults that are genetically related to, and are considered to have guided, the emplacement of late Mesozoic plutons. In a few places, these deposits occur near block faults or tear faults which may have been formed in the Precambrian.

In contrast, the deposits containing old lead are found associated with tear faults of probable Precambrian age. A few of these deposits occur along the crests of folds near Precambrian block faults or in fractures within or along the contacts of Precambrian sills and dikes.

Table 1 shows the association with primary structural features as well as the prevailing type of vein fractures for the selected districts.

CONCLUSIONS

The observed characteristics established a generalized model for these deposits that may be useful in determining their genesis as well as in analyzing their economic potential.

Those veins containing old lead are associated with major fractures considered to be possibly Precambrian and certainly pre-Mesozoic in age. Moreover, these deposits contain a preponderance of lead to either silver or zinc, and they must be economically evaluated accordingly. In contrast, veins containing young lead are associated with pre-Cenozoic fractures, particularly those related to late Mesozoic intrusive action. In these deposits silver plays a much more important economic role than the base metals.

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Metallogenesis in the Coeur d’Alene Mining District

by

Wilfred Walker

The Coeur d’Alene metallogenesis is exceptionally well-developed, as production figures testify, but other aspects, at times considered exceptional, are comparable with those of other places, and the purpose of this note is to provide references for comparison and to summarize the metallogenesis as it appears to me.

The nature of the Belt host rocks as described at this conference, particularly by Mumma, Harbour, and Jayne (1977) and by White and Winston (1977), is clearly platformal. Though turbidites are known in the area, their relationship to the fluvialite quartzites and reef limestones was not elucidated; and turbidites were not described in relation to ore.

The time and duration of deposition of the Belt Supergroup is controversial. At the conference, Zartman (1977) provided data on the Purcell “C” sill, intruding the Fichardt Formation: remarkably concordant uranium-lead ages of 1,430 m.y. were reported. This sill may have been intruded during sedimentation of the Belt Supergroup or during later orogeny.

Though there is no evident reason for platformal deposition to be bound by temporal restriction, nevertheless sequences of fluvialite quartzites and reef limestones, commonly accompanied by glacial products and evaporites, with stratabound gold, uranium, copper, and lead-zinc-silver mineralization do commonly appear in a post-tectonic setting (Table 1). Thus, though deposition of the Belt was probably more temporally restricted than was evident, for example, in the review by Young, Jefferson, Long, Delaney, and Yeo (1977), there are several possibilities for its temporal position. Because of its undoubted post-Hudsonian time and Zartman’s 1,430 m.y. date, the position I prefer is post-tectonic to the Early Proterozoic. Zartman has already given reason for questioning previously published dates, and I do not propose to dwell here on the reliability of radiometric dates.

The next question is the origin of the lead-zinc. From the chair, Farmin, who had already described ore-stratigraphic relationships in the Coeur d’Alene district (Farmin, 1977), reverted to a question on origin—down below. Though the response generated much noise from the audience, all ores, as with other parts of the earth’s crust, originally came from the mantle by igneous activity. The mobilization of metal during orogeny, described for the Coeur d’Alene by Juras (1977), will be considered more widely below; meanwhile, one must consider whether the mobilized metal came directly from the mantle, as Juras seemed to believe during the discussion, or from the host sediments, for which Farmin (1977) and Ramalingaswamy (1977) provided evidence. The various ore types are emplaced in quartzites of various ages, and the control of ore type does not therefore appear to be strictly lithological but rather on some broader basis. Bilbins (1968), tabulating data from many deposits (largely Phanerozoic and in the USSR), ascribed the types of gold, uranium, copper, and lead-zinc stratabound deposits to the late and terminal stages of the tectonic cycle. In the initial part of the tectonic cycle, the relationship of mineralization to volcanism is clear; in subsequent stages, concentration by plutonism is evident. The origins of late stage ores are less clear, but the separation of ore types in time indicates that some ore types may be generated from the mantle at different phases of the tectonic cycle. Furthermore, whereas gold, uranium, and copper all appear in earlier parts of the tectonic cycle, the same is not so for lead, zinc, and silver. I consider lead-zinc-silver to evolve from the mantle during late stages of mobile belt development, rather than after each orogeny within the mobile belt. Some mode of transport from the mantle is needed. During the discussion, Reid noted that the Lewis and Clark lineament was already extant no later than the Early Proterozoic and that the control of mineralization by such major structures has long been known (Billingsley and Locke, 1941). What is unclear is whether such structures are the restricted sites of deposition as well as the sites of remobilization to ore grade. From a comment from the floor about the lack of mineralization in the sediments distant from the Lewis and Clark lineament, the lineament would appear to be the primary channelway.

The platformal sediments, suggested therefore as post-tectonic, undoubtedly suffered orogeny. Reid, Morrison, and Greenwood (1973) have described the plutonism which accompanied deformation and metamorphism in the Clearwater orogeny zone, south of the Coeur d’Alene belt in northern Idaho. In the Coeur d’Alene area, the minor amount of plutonic activity does not allow for such knowledge of the tectonic cycle.
Tectonic cycles are components of mobile belts (which comprise one or more tectonic cycles) and the question of the age of the mobile belt arises. Gastil (1960) provided evidence that mobile belts form during discrete periods and are worldwide, and I have discussed this topic elsewhere (Walker, 1976a; 1976b). Consideration must be given to the orogenic activity which affected the Belt of northern Idaho being either Middle or Late Proterozoic; there is no evidence of two deformational and metamorphic histories, and the preponderance of dates is Mid-Proterozoic (Reid and others, 1973). Deformational and metamorphic histories tend to conform to a pattern (Fyson, 1971), much like that described by Reid and others (1973) for the Clearwater orogenic zone. In the Caledonian orogeny of the Scottish Highlands, D1, D2, and D3, the phases of penetrative deformation were completed in the 50 m.y. between 500 and 450 m.y. ago. Plots of the Archean Rb-Sr dates against $\text{Sr}^{87}/\text{Sr}^{86}$ ratios (I.R.s) indicate that the common development time for orogeny is less than 100 m.y.

The deformation of lead-zinc orebodies has been described at Balmat-Edwards N.Y. (Brown, 1969), but the only previous study where deformation and metamorphism of the rocks hosting lead-zinc-silver ores has been related to ore paragenesis, as Jurus depicted for Bunker Hill at this conference, appears to be that by Sirola (1977) on the Grum deposit in Cambrian sedimentary and volcanic rocks in the Yukon.

By way of comparison, however, there are several works relating ore properties to deformation and metamorphism to mineralization which first appeared during initial stages of tectonic cycles.

For gold, in a classic Geological Survey of Canada memoir, Boyle (1961) showed, at Yellowknife, Northwest Territories, late Archean, gold sulfide ores related to early metamorphism and thrusting, later gold quartz ores to subsequent periods of metamorphism and lesser fracturing. Ramsey (1963) at Barberton, South Africa, showed the Middle-Archean deformational history, and subsequent discussion brought out that gold sulfides were emplaced in early D2, gold quartz in later D2 and subsequent periods of metamorphism and deforma-

<table>
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<tr>
<th>Orogenies</th>
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<td>Dwyka tillites</td>
<td>Europe and Mississippi Valley: lead-zinc</td>
</tr>
<tr>
<td></td>
<td>Mississippi cyclothemys</td>
<td>Europe: copper</td>
</tr>
<tr>
<td>Late Proterozoic</td>
<td>Cambrian limestones</td>
<td>Scotland: minor lead-zinc</td>
</tr>
<tr>
<td></td>
<td>Varangian tillites</td>
<td>Copperbelt: copper</td>
</tr>
<tr>
<td>Middle Proterozoic</td>
<td>Roan and Haut Shiloango including cyclothemys and tillites</td>
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</tr>
<tr>
<td>Early Proterozoic</td>
<td>Wollaston Group of Saskatchewan including terrestrial quartzites, platform carbonates, and evaporites? Menominee and Marquette: iron</td>
<td></td>
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<tr>
<td></td>
<td>Aninikie of central North America including shallow water quartzites and chemical sediments</td>
<td></td>
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<tr>
<td>Late Archean</td>
<td>Huronian cyclothemys and tillites</td>
<td>Elliot Lake: uranium</td>
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<td>Hamersley basin shelf and basin sediments</td>
<td>Mt. Tom Price and Whaleback: iron</td>
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<td></td>
<td>Witwatersand cyclothemys and tillites</td>
<td>Rand: gold-uranium</td>
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tion. At Rice Lake, Manitoba, Stephenson (1971), in describing late Archean gold deposits, noted that the observed depletion of silver, aluminum, calcium, and sodium in altered wall rocks and their prevalence in the veins leads to the conclusion that lateral migration and concentration of these elements were important in the formation of many of the veins in the same volume. McRitchie, Weber, and Soothes (1971) tabulated ore paragenesis in relation to metamorphism and deformation. At the Homestake, Sawkins and Rye (1974) noted the migration of gold from gold-rich host sediments into structural traps.

For base metals, Forsythe (1971: 1972) showed ore paragenesis related to deformation at the Anglo-Rouyn Mine during the Early Proterozoic development of the Saskatchewan section of Churchill Province. The most detailed paper on the theme is by Wakefield (1976) on the Mid-Archean nickel ores of Pikw-Selebi, Botswana. The sedimentation, deformation, and metamorphism at Broken Hill, New South Wales, has been described in a series of papers by the Adelaide and Monash Schools (Both and Rutland, 1976; Glen and Laing, 1975; Glen, Laing, Parker, and Rutland, 1977; Laing, Monjoranks, and Rutland, 1976; Phillips, Lee, and Wall, 1976; Rutland and Etheridge, 1975), and these must be related to ore paragenesis described by Lawrence (1975). The studies at O’okiep in the northern Cape Province of South Africa are at a similar stage (Joubert, 1974; Benedict, Wilt, Cornellissen, and staff, 1974). Lastly, the proceedings of the conference on ore genesis at Lulea, Sweden, resulted in two papers published in English (by Juve 1977) on Stekenjokk and Gadd (1977) on Outokumpu, in which ore paragenesis and deformation are related.

The patterns of deformation in all these areas are similar: relationships of ore paragenesis also merit comparison.

In summary, I suggest that the Belt Supergroup of platformal sediments was deposited, followed by Early Proterozoic mobile belt formation, in the period 1.7 to 1.5 b.y. ago. Lead-zinc-silver from the mantle probably reached the surface via faults wedged by the Lewis and Clark lineament and were either syngenic or diagenetic as described by Ramalingasmwamy (1977). During the Mid-Proterozoic mobile belt formation (1.5 b.y. to 960 m.y. ago) mineralization was mobilized in several phases giving an ore paragenesis related to deformation and metamorphism as described by Juras (1977) and by Reid (1977). Within the rather lengthy Mid-Proterozoic, the history of penetrative deformation in any particular area such as Coeur d’Alene probably lasted no more than 100 m.y.

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Age of the Crossport C Sill Near Eastport, Idaho

by

R. E. Zartman1, Z. E. Peterman1, J. D. Obrodovich1, M. D. Gallego2, and D. T. Bishop3

ABSTRACT

A new approach in establishing geochronologic control on the development of the Belt sedimentary basin involves uranium-thorium-lead dating on zircon of granophytic differentiates of mafic igneous rocks. The Crossport C sill of Bishop (1973), which intrudes the lower to middle parts of the Prichard Formation near Eastport, Boundary County, Idaho, was chosen for preliminary investigation. Its position near the base of the exposed Belt Super group makes the sill useful in placing a minimum age on the beginning stages of basin formation.

Two types of granophyre are present in the C sill—a more or less continuous, differentiated medial zone and a discrete, discordant body related to a local deflection in the sill. Pink, euhedral to anhedral zircon containing microscopic inclusions of chloride (?) and a peculiar internal fracturing was recovered from samples representing both types of granophyre occurrence. Hand-picked size and magnetic fractions of zircon were analyzed and found to have concordant uranium-lead isotopic ages of 1,433 ± 10 m.y. By contrast, the thorium-lead ages show a somewhat disturbed pattern, which suggests that the uranium and thorium may not be located in the same lattice sites within the zircon. The close approach of the uranium-lead isotope system to closed-system behavior leads us to accept the 1,433 ± 10 m.y. age as the time of primary crystallization of the sill.

The application of other dating methods to the C sill gives ambiguous results that reflect the complex geologic conditions to which the rocks have been subjected. A rubidium-strontium whole-rock isochron plot shows substantial scatter and requires assumptions about original inhomogeneity in strontium isotopic composition or subsequent open-system conditions in order to be interpreted. Potassium-argon analyses on amphiboles do not record the primary crystallization age, but attest to the effect of later metamorphism.

Further study is required to determine if the uranium-lead analyses on zircon isotopic systematics will be so well behaved in sills occupying other stratigraphic levels of the Belt Super group. The time of sill emplacement is a minimum stratigraphic age, and certain assumptions must be made about depth control of concordant intrusions and sedimentation rates to arrive at the age of the host rock. Nevertheless, this initial work already revises upward (i.e., older) from some earlier published results the probable age of the Prichard Formation.

INTRODUCTION

Attempts to date the sedimentary units of the Belt Super group have produced equivocal results, and associated igneous rocks are neither abundant nor do they necessarily give stratigraphic control. A low-grade metamorphic overprint, especially within the lower units of the Belt, has also cast doubt on the reliability of published ages for this part of the section. Despite these difficulties, Obrodovich and Peterman (1968) carried out a detailed geochronologic study, which for a decade has remained the basis for assigning depositional ages to the Belt Super group. Some data, however, reveal that the beginning stages of basin formation may exceed by as much as 200 m.y. the oldest age determined in that study, which was about 1,300 m.y. on the lowest strata of the Big Belt (Greyson Shale and Newland Limestone) and the Little Belt Mountains (Chamberlain Shale and Neihart Quartzite). However, no correlation has been established as yet between these rocks and the Prichard Formation to the northwest, for which the evidence of an older age is strongest.

In searching for new approaches in refining the Belt Super group time scale, our attention has turned to the locally abundant gabbric sills, which intrude the Prichard Formation along the western flank of the Purell anticlinorium in northern Idaho, northwestern Montana, and adjacent British Columbia (Figure 1). The sills were emplaced in this formation prior to

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regional deformation, and they hold promise of providing an important time constraint to both the depositional and structural history of the Belt rocks. We report here on a preliminary application of several dating methods to the Crossport C sill near Eastport, Boundary County, Idaho (Bishop, 1973). Although yielding only a minimum age of deposition for the intruded horizon, our results significantly increase the probable age of the lower to middle parts of the Prichard Formation. Particularly encouraging are the almost concordant uranium-lead ages on zircon obtained for granophytic differentiates of the mafic igneous rocks. This close approach of the zircon uranium-lead isotope system to closed-system behavior offers strong support that the sill crystallized 1,433 ± 10 m.y. ago, and that, therefore, at least part of the Prichard Formation must be even older. The other dating methods give ambiguous results that reflect the complex geologic conditions to which the rocks have been subjected. A rubidium-strontium whole-rock isochron plot shows substantial scatter and requires assumptions about original inhomogeneity in strontium isotopic composition or subsequent open-system conditions in order to be interpreted. Potassium-argon analyses on amphibole do not record the primary crystallization age, but attest to the effect of later metamorphism.

**SAMPLE DESCRIPTION**

This study derives from an earlier investigation to determine the petrologic history of several gabbroic sills intruding the lower and middle parts of the Prichard Formation in the vicinity of the Crossport quarry near Eastport, Idaho (Figure 2). All samples included herein were obtained from the 400-meter-thick Crossport C sill, which displays a characteristic differentiation trend of gabbro to ferrodiortite to quartz diorite to biotite quartz diorite to biotite granophyre. The reader is referred to the paper by Bishop (1973) for a more detailed petrographic and geochemical description of the sills. Representative rock specimens from several members of the differentiation series were collected for radiometric age determination (Figure 3). We will only document the pertinent features of these samples as they relate specifically to the geochronologic results.

**DB-1** Dark, greenish black, fine-grained ophitic hastingsite gabbro, from 180 meters below the top of C sill, Crossport quarry.

**DB-2** Mottled, medium-grained hypidiomorphic quartz diorite with minor granophyric mesostasis, from 60 meters below the top of C sill, Crossport quarry.

**DB-3** Greenish gray, medium-grained granophyre composed of plagioclase (An80) and quartz enclosing elongate plumeite grains of ferrohastingsite with a later quartz-plumeite mesostasis. From 20 meters below the top of C sill, Crossport quarry. This granophyre occurs as a segregation body above the upper gabbro chill zone in a low-pressure region where the sill becomes discordant with the sedimentary layering.

**Figure 1. Location map.**

**Figure 2. Geologic map of the Crossport sills near Eastport, Boundary County, Idaho.**
development of granophyre mesostasis within the upper part of the differentiated quartz diorite.

**EXPERIMENTAL RESULTS**

Radiometric data were obtained by the potassium-argon analysis of ferrohastingsite, the rubidium-strontium analysis of whole rock, and the uranium-thorium-lead analysis of zircon. Established analytical procedures of isotope-dilution chemistry and mass spectrometry were employed; therefore, only a brief account of these experimental techniques will be given here.

The rocks collected for age determination were taken from apparently unweathered exposures within or adjacent to the Crossport quarry. Complexities of the radiometric systems attributable to postcrystallization metamorphism made it especially necessary to avoid the effects of surficial weathering. Thin sections were made for all rocks prior to processing in order to aid in mineral identification and sample selection. All whole-rock samples were taken from at least 5 kilograms of initial material, which was carefully split and pulverized to ~100 mesh size. The size of the whole-rock sample was always large compared to the grain dimension, but it may not be representative of the larger scale textural inhomogeneity commonly displayed by the quartz diorite and granophyre samples. The ferrohastingsite and zircon concentrates were obtained by processing a fraction of the whole rock through heavy liquids and a magnetic separator. Final purification of the zircon was accomplished by hand picking under a binocular microscope.

The experimental procedure for the potassium-argon method used is described in detail by Dalrymple and Lanphere (1969). Argon was extracted by direct fusion with an RF induction heater, and an \(^{40}\text{Ar}\) tracer was added prior to sample cleanup and mass-spectrometer analysis. The potassium content of the amphiboles was determined in duplicate by isotope dilution and reported as the average value. None of the argon determinations required substantial atmospheric corrections, and an accuracy reflecting reproducibility in potassium analyses has been assigned to the potassium-argon ages in Table 1.

**Table 1. Potassium-argon ages of ferrohastingsite from the C sill, Crossport, quarry.**

<table>
<thead>
<tr>
<th>Sample Number</th>
<th>K2O (percent)</th>
<th>(^{40}\text{Ar} / {^{39}}\text{Ar}) Radiogenic (moles/gram)</th>
<th>Percentage of (^{40}\text{Ar}) of Radiogenic Origin</th>
<th>Age, in Millions of Years</th>
</tr>
</thead>
<tbody>
<tr>
<td>DB-1</td>
<td>0.364</td>
<td>7.305 \times 10^{-10}</td>
<td>86.9</td>
<td>894 ± 15</td>
</tr>
<tr>
<td>DB-5</td>
<td>0.875</td>
<td>15.59 \times 10^{-10}</td>
<td>95.9</td>
<td>813 ± 9</td>
</tr>
</tbody>
</table>

Decay constants:

\[ \lambda_{K} = 4.33 \times 10^{-10} \text{ yr}^{-1}, \lambda_{^{40}\text{Ar}} = 5.51 \times 10^{-10} \text{ yr}^{-1}, \lambda_{^{40}K} = 0.0167 \text{ atom percent.} \]
The analytical technique for the rubidium-strontium method follows the procedure described by Peterman and others (1967). Rubidium and strontium concentrations were determined by isotope dilution, and strontium isotopic composition was measured from a separate, unspiked analysis (Table 2). All strontium data have been normalized to a $^{87}$Sr/$^{86}$Sr ratio of 0.1194 to eliminate the effects of mass-spectrometer fractionation. The accuracy of the $^{87}$Rb/$^{86}$Sr ratio is taken to be ± 2 percent of the ratio, and normalized $^{87}$Sr/$^{86}$Sr ratios are generally reproducible to ± 0.0005. The whole-rock isochron age (Figure 4) has been computed for four of the six samples using the least-squares program of York (1966). The two samples of granophyre depart significantly from the linear array on the isochron diagram and were not included in the age calculation.

The uranium-thorium-lead chemistry was carried out using a modified Krogh (1973) method of HF-HNO$_3$ bomb digestion, after which lead was purified by chloride-form anion resin exchange columns and anodic electrodereposition. Prior to dissolution, zircon was rinsed in hot 6N HCl and 6N HNO$_3$ for 20 minutes each to remove any soluble surface contamination or inclusions accessible to the acid. Lead concentration was determined by isotope dilution on an aliquot of the sample removed after bomb digestion, but this aliquot was otherwise subjected to a parallel treatment with the isotope composition fraction. A combined uranium and thorium tracer was added to the entire sample before digestion, and uranium and thorium were separated together from the eluate off the resin column of the lead isotope composition fraction by the nitrate-form anion resin method of Tatsumoto (1966). Mass spectrometric analysis for (a) lead was by surface ionization from a single rhenum filament with H$_3$PO$_4$-silica gel emitter, and for (b) uranium and thorium, by volatilization off two rhenum side filaments with a rhenum-iridium ionizing filament. Uranium, thorium, and lead concentra-

![Figure 4. Diagram of rubidium-strontium whole-rock isochron (age and intercept based on linear regression through samples DB-1, DB-2, DB-4, and DB-5 only).](image)

Table 2. Rubidium-strontium isochron age of whole rocks from the C sill, Crosport quarry.

<table>
<thead>
<tr>
<th>Sample Number</th>
<th>Rb (ppm)</th>
<th>Sr (ppm)</th>
<th>$^{87}$Rb/$^{86}$Sr</th>
<th>$^{87}$Sr/$^{86}$Sr</th>
<th>Age in Millions of Years</th>
</tr>
</thead>
<tbody>
<tr>
<td>DB-1</td>
<td>12.7</td>
<td>141</td>
<td>0.261</td>
<td>0.7093</td>
<td></td>
</tr>
<tr>
<td>DB-2</td>
<td>7.2</td>
<td>187</td>
<td>0.111</td>
<td>0.7057</td>
<td></td>
</tr>
<tr>
<td>DB-3</td>
<td>4.1</td>
<td>761</td>
<td>0.016</td>
<td>0.7081</td>
<td>1285 ± 165 (2σ)</td>
</tr>
<tr>
<td>DB-4</td>
<td>29.1</td>
<td>123</td>
<td>0.688</td>
<td>0.7175</td>
<td></td>
</tr>
<tr>
<td>DB-5</td>
<td>59.9</td>
<td>182</td>
<td>0.754</td>
<td>0.7210</td>
<td></td>
</tr>
<tr>
<td>DB-10</td>
<td>10.0</td>
<td>231</td>
<td>0.126</td>
<td>0.7085</td>
<td></td>
</tr>
</tbody>
</table>

Decay constant: $\lambda_B = 1.42 \times 10^{-11}$ yr$^{-1}$ age calculated by least-squares regression method of York (1966). See graphic representation of these data on Figure 4.

$^a$Samples not included in calculation of age.
Figure 5. Morphology of representative zircon grains from (a) DB-3 and (b) DB-10 (note euhedral to anhedral habit and internal fracturing). The diameter of each field is about 2 millimeters.

DISCUSSION

The several approaches to obtaining radiometric ages reported in Tables 1 and 3 clearly reveal complications in the ability of the various dating methods to record the time of crystallization of the Cospropot C sill. Nevertheless, when these new results are viewed together with previously established field and laboratory information on the geologic evolution of this area, a coherent interpretation can be made in terms of primary emplacement and subsequent metamorphism of these rocks. In fact, a geochronologic pattern, which may serve to refine both its depositional and structural history, is beginning to emerge regionally throughout the Belt basin. No comprehensive treatment of the subject will be offered in this report; rather we give emphasis to the new use of uranium-thorium-lead dating of zircon, which holds promise of adding substantially to our knowledge of the timing of mafic intrusion and, consequently, of the early stages of basin formation.

Previous radiometric ages that fall within the general range of 750-900 m.y. have been reported by Hunt (1962), Goldich and others (1959), and Obradovich and Peterman (1968). Yet only the uppermost stratigraphic units of the Belt Supergroup and certain mafic intrusions could actually have originated at about this time. Glauconites from the Sun River, Montana, area occurring as high in the stratigraphic column as the McNamara Formation give ages of approximately 1.100 m.y. by both the potassium-argon and rubidium-strontium methods (Gulbrandson and others, 1963; Obradovich and Peterman, 1968). A similar age has also been obtained by the rubidium-strontium whole-rock isochron method on samples of shale from the Helena Dolomite and the Empire, Sheeped, Mount Shields, and McNamara Formations. Only the Garnet Range Formation and the Pilker Quarzsite from the Alberton, Montana, area yielded a rubidium-strontium whole-rock isochron age as low as 900 m.y. (Obradovich and Peterman, 1968). It is evident that many of the young ages cannot reflect primary igneous and sedimentary events. The two potassium-argon dates on ferrohastingsite of 894 ± 15 m.y. and 813 ± 9 m.y. from the C sill add to this growing number of data, which suggest that a metamorphic thermal pulse tentatively correlated with the East Kootenai orogeny of Canada was widely felt throughout the northern part of the Belt basin. That the C sill did undergo extensive greenschist to amphibole kinematic metamorphism after emplacement is evidenced by the shearing of the rock and the growth of actinolite, epidote, oligoclase, chlorite, and carbonate. The samples chosen for potassium-argon analysis were themselves virtually free of these secondary minerals, but conditions allowing for argon loss from the ferrohastingsite may have prevailed generally at this time. Although not yet proven in the

Table 3. Uranium-thorium-lead ages of zircon from the C sill, Cospropot quarry.

<table>
<thead>
<tr>
<th>Sample Number</th>
<th>Concentration (ppm)</th>
<th>Isotopic Composition of Lead (atom percent)</th>
<th>Age, in Millions of Years</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>U</td>
<td>Th</td>
<td>Pb</td>
</tr>
<tr>
<td>DB-3 (−100+200) mm</td>
<td>1147</td>
<td>920</td>
<td>325</td>
</tr>
<tr>
<td>(−100+200) m</td>
<td>1037</td>
<td>849</td>
<td>282</td>
</tr>
<tr>
<td>DB-10 (−50+100)</td>
<td>1043</td>
<td>817</td>
<td>296</td>
</tr>
<tr>
<td>(−200+270)</td>
<td>1081</td>
<td>740</td>
<td>299</td>
</tr>
</tbody>
</table>

Decay constants:

\[ k_{204Pb} = 1.551 \times 10^{-10} \text{ yr}^{-1}, k_{206Pb} = 9.849 \times 10^{-10} \text{ yr}^{-1}, k_{207Pb} = 4.948 \times 10^{-11} \text{ yr}^{-1}, k_{235U} = 1.16 \times 10^{-11} \text{ yr}^{-1} \]

Isotopic composition of initial lead assumed to be 204Pb:206Pb:207Pb:208Pb = 1:1.2:15:436.0.

Mass spectrometry designated in parentheses; mm is less magnetic fraction, m is more magnetic fraction.
field, the metamorphism may have taken place in conjunction with some phase of the regional arching of the Purcell anticlinorium. On the basis of the rubidium-strontium and uranium-thorium-lead results discussed subsequently, we can definitely rule out the retention of an original crystallization age by these amphiboles.

Six whole-rock samples representing various rock types of the differentiated C sill were analyzed by the rubidium-strontium technique in an attempt to define an isochron age. The resultant data (Figure 4) are not collinear, suggesting that for at least some of the samples the basic assumptions of the method have not been satisfied; that is, either these samples did not achieve complete strontium isotopic homogeneity at the time of crystallization, or else subsequent open-system conditions have altered the petrology of the rock. In particular, the two samples of granophyre (DB-3 and DB-10) have anomalously high $^{87}Rb/^{86}Sr$ values relative to their $^{87}Sr/^{86}Sr$ values when compared to the remaining samples. Furthermore, we might have expected the highest rubidium-strontium ratios to be associated with these two samples of granophyre, but no such correlation reflecting the differentiation trend exists. Instead, the granophyre seems to have been enriched in strontium, which on the basis of isotopic composition derived from some source either external to the sill or after sufficient time had elapsed to alter the initial isotopic composition of the sill. It is possible that the granophyre was contaminated by the graywacke country rock during late-stage deuteric alteration, or that a redistribution of strontium or rubidium accompanied later regional metamorphism, such as during the East Kootenai orogeny.

The remaining four samples, DB-1, DB-2, DB-4, and DB-5, do yield a rather imprecise age of 1,285 ± 165 m.y. (2σ) when subjected to the least-squares program of York (1966). Statistical tests on these data clearly identify scatter on the isochron diagram beyond that expected from the analytical uncertainty alone, and it is probable that the problems found with the granophyre samples also occur to some degree here. For example, sample DB-5 would tend to have a $^{87}Sr/^{86}Sr$ ratio reduced if isotopic exchange took place among any of the other analyzed samples. In that case, the isochron will undergo rotation toward a lower slope, and consequently, a younger age. Even when allowance is made for the relatively high calculated error for the whole-rock rubidium-strontium isochron age, the age by this method barely has overlapping uncertainties with that by uranium-lead on zircon. Because we believe this latter method gives the most reliable primary crystallization age, some disturbance of the rubidium-strontium isotopic systematics may have happened pervasively throughout the sill. Although an isochron age calculated only for samples DB-1, DB-2, and DB-4 of 1,410 ± 140 m.y. does agree closely with the uranium-lead age of zircon, such a selective use of the data cannot be justified objectively.

The presence of appreciable amounts of zircon within the granophytic rocks of the C sill suggested that the uranium-thorium-lead analyses of this mineral might further help define the primary crystallization age. A distinctive pinkish, euhedral to anhedral population of zircon (see Figure 6) was separated from samples DB-3 and DB-10 as representative of two modes of granophyre occurrence. No differences in zircon morphology were noted between the two samples, and examination of thin sections and grain mounts leads us to believe that the mineral formed entirely during late-stage crystallization of the gabbroic magma. In particular, inherited cores with later overgrowths or bimodal distribution in size, shape, or color were not observed. An abundance of microscopic inclusions of chlorite (?) and a peculiar internal fracturing do characterize the zircon population, but probably relate to the deuteric conditions under which the granophyre crystallized and cooled. The several fractions of zircon actually analyzed were carefully handpicked to avoid the most turbid material.

The analytical data for the zircons is given in Table 3 and shown graphically on a concordia diagram in Figure 5. Commonly, Precambrian zircon with this amount of uranium is found to be slightly to moderately discordant, and the determination of an age requires some interpretation of the concordia plot. The presence of an impurity, such as chlorite, might also be expected to introduce further complications into the isotopic data. Thus, it was surprising to discover that all zircon fractions yielded almost concordant uranium-lead ages—206Pb/238U age by no more than 3 percent relative to the $^{207}Pb/^{206}Pb$ age. Consequently, we believe these data accurately date the time of sill emplacement. We prefer to follow the common practice of giving greater credence to the more precise averaged $^{206}Pb/^{207}Pb$ age of 1,435 ± 10 m.y., although, even if the daughter-parent ages are included, they all lie within the relatively narrow range of 1,417-1,436 m.y. (excluding sample DB-3 (~100 + 200) m., which appears to be discordant beyond analytical uncertainty). Perhaps some combination of vagaries in the isotopic behavior could have fortuitously produced these essentially concordant ages, but such an explanation seems extremely improbable. It is our contention that the uranium-lead age of zircon is the least likely to be disrupted of the ages from all the dating techniques used.

By contrast, the thorium-lead ages of zircon are all somewhat low in comparison to the uranium-lead ages.
Because the isotopic systematics associated with the $^{232}$Th decay scheme are much less constraining than those of the coupled $^{238}$U and $^{234}$U decay schemes, this discrepancy is easier to interpret in terms of disturbed thorium-lead systems. Apparently, the two radioactive elements or their decay products are in crystallographic sites at least partially independent of each other. They may occupy separate lattice positions in the zircon or, more probably, some of the thorium may reside in a different mineral phase altogether. The relatively high thorium-uranium ratios determined for these samples compared with many other analyzed zircons, as well as the observed presence of crystalline inclusions, tend to support such a hypothesis. The discrepancy could be explained, for example, if one-fourth of the thorium occurs in a phase which suffered substantial daughter-lead loss at the same time that the amphibole ages were reset.

In summary, we conclude that the uranium-lead data on zircon give the most reliable crystallization age of the C sill from the Crossport quarry near Eastport, Idaho. At the same time, the other less certain dating techniques do contribute to a better understanding of the total geologic history of the sill. A variety of factors, such as post-crystallization metamorphic events and isotopic inhomogeneity of the gabbroic magma, are thought to have complicated the results obtained by dating the potassium-argon age of ferrobasaltic glass and the rubidium-strontium age of whole-rock (and thorium-lead of zircon). While an ultimate goal remains one of deciphering the complete evolution of the Belt basin from its inception until the present, our study mainly establishes a rather precise age of $1,433 \pm 10$ m.y. on the C sill, which is one aspect of basin development. Furthermore, this result has obvious implications for placing a minimum age on the time of deposition of the lower to middle part of the Prichard Formation which the sill intrudes. In this regard the age of the sill is particularly interesting, because it suggests an earlier age than some previous estimates for the beginning stages of the Belt basin.

To extrapolate from the age of the sill to a stratigraphic age, certain assumptions must be made about depth control of concordant intrusions. This subject has been treated by Mudge (1968) and Greten (1969); and while these authors differ in approach to the problem, they both recognize an upper limit to the amount of overburden at the time of intrusion. Mudge demonstrated from his study of concordant intrusions in the western United States that this maximum limit is on the order of 2,500 meters. Also, he found that rarely do such bodies form where the cover is less than 1,000 meters. This range in depth would mean that the mafic sills here were intruded into the lower to middle part of the Prichard Formation some time during concurrent deposition of the upper part of the Prichard or overlying Ravalli Group rocks. It is possible that the Ravalli delta was already well developed and regionally extensive when renewed tectonic forces caused a reorientation of the basin and provided conduits for ascending magma. However, insufficient control exists at present to relate the age of the C sill to a more refined stratigraphic position than is indicated by the broad estimate of necessary overburden.

We can also speculate briefly on how much older the sedimentary rock at the horizon of the sill is than the intrusion itself. This approach requires that some estimate be made of the rate of sedimentation; that is, how long did it take to deposit the overlying 1,000-2,500 meters of rock? Widely differing rates of sedimentation have been measured for both modern
and ancient argillaceous and arenaceous rocks from depositional environments comparable to the Belt basin, but the most frequently found values are 0.1-0.01 millimeter a year. If we use these rates, the minimum time elapsed in accumulating the additional 1,000-2,500 meters of sediment would be 10-25 m.y. and the maximum time would be 100-250 m.y. Thus, even in our most conservative estimate, we need to entertain a minimum stratigraphic age of 1,450 m.y. for the lowermost portion of the Belt Supergroup, that is, the lower part of the Prichard Formation and the equivalent Aldrich Formation in Canada. Left unanswered here is the question as to whether the 1,300-m.y. rubidium-strontium whole-rock isochron age for the basal strata of the Big Belt and Little Belt Mountains, the Greyson Shale and Newland Limestone and the Chamberlain Shale and Neihart Quartzite (Obradovich and Peterman, 1968) reflects a younger depositional age for those units or the response to some later metamorphism.

The study by Ryan and Blankinship (1971) on the Hells Canyon Creek stock, British Columbia, Canada, yielded an imprecise rubidium-strontium whole-rock isochron age of about 1,300 ± 30 m.y. This stock, in turn, intrudes Moyie sills, which were themselves emplaced into the upper part of the Aldrich Formation. The Moyie sills and the Crossport sills may represent the same period of igneous activity.

A maximum limit to the age of the Belt Supergroup remains that of the plutonic and high-grade metamorphic crystalline terrane, which occurs as an unequivocal basement in a number of localities. Extensive radiometric dating of the basement rocks has firmly established the last major dynamothermal metamorphic event prior to Belt sedimentation at approximately 1,700 m.y. ago (Hedge and others, personal communication). Allowing for uplift and erosion to bring this terrane to the surface—some insight into this process is recorded in potassium-argon ages of mica at less than 1,600 m.y.—no great difficulty is encountered in reconciling the upward revision of lower Belhian time as proposed in this paper with known geologic relationships.

The uranium-lead dating technique with zircon from granophyre for establishing the crystallization age of a gabbroic intrusion—where problems were found with other dating techniques—holds promise for further application. It is not yet known if the isotope systematics will be as well behaved in bodies occupying other stratigraphic levels of the Belt Supergroup. Certainly, several more occurrences should be studied to evaluate the general usefulness of the method. A chronology for these mafic intrusive rocks may contribute not only to deciphering cycles of igneous activity but also to constraining the primary sedimentary record of the basin. Furthermore, by resolving between original crystallization ages and disturbed ages, the other dating techniques themselves may provide a clearer picture of regional thermal events.

REFERENCES


Reconnaissance Geology of the Blackbird Mountain Quadrangle, Lemhi County, Idaho

by

Earl H. Bennett

ABSTRACT

The area of the Blackbird Mountain 15-minute quadrangle in Lemhi County, Idaho, was mapped in support of a reconnaissance geology and geochemical survey. The area is significant to national mineral needs because it contains the Blackbird Mine, a readily available domestic source of cobalt.

Major rock units include three Precambrian units—the Yellowjacket Formation, the Hoodoo Quartzite, and an augen gneiss complex; and two Cretaceous-Tertiary intrusive rocks—the Crags pluton and the Leesburg stock.

The Yellowjacket Formation covers most of the quadrangle and is subdivided into a lower phyllite member and an upper dark gray quartzite member. The quartzite is intruded by the augen gneiss complex dated at 1,500 m.y.

In most places the Hoodoo Quartzite is thrust over the Yellowjacket Formation. The base of this thrust is believed to be mineralized in the Leesburg basin by the Leesburg stock.

There are two periods of mineralization in the study area. The Blackbird Mine is a cobalt-copper deposit that may be Precambrian in age. The Leesburg basin has placer and lode deposits of gold, silver, lead, and molybdenum mineralization that is Cretaceous-Tertiary in age.

The Blackbird Mine appears to be in a horst between the Slippery Creek fault and the White Ledge shear zone. Rocks that are believed to be higher ranked metamorphic equivalents of the Yellowjacket Formation are the orebody host. The relative competency of the phyllite and quartzite members of the Yellowjacket Formation may be important in explaining the genesis of the orebodies in the Blackbird Mine.

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Upper Proterozoic Stratigraphy of Northwestern Canada and Precambrian History of the North American Cordillera

by

G. M. Young*, C. W. Jefferson†, G. D. Delaney*, G. M. Yeo*, and D. G. F. Long†

ABSTRACT

The northwestern part of the Canadian Shield contains evidence critical to the dating of Proterozoic events in the northern Cordillera. Stratigraphic correlation of homotaxial successions from the northern margin of mainland Canada to Victoria Island and thence to the Mackenzie Mountains area has assigned the older Proterozoic rocks of the Mackenzie Mountains area to an age between 1.2 b.y. and 700 m.y. All of these correlated assemblages are shallow marine platform deposits, but there is some indication of westward deepening. Rocks of these widely separated regions are considered to have been deposited in an extensive basin, the Amundsen Embayment, that extended eastward into the shield from the great arc of the Mackenzie Mountain fold belt in the northern part of the Cordillera.

The youngest Proterozoic assemblage in the Cordillera includes important glacial deposits and intimately associated iron-formations that are part of the Rapitan Group. If these rocks are correlative with similar successions to the south, then deposition began about 800 m.y. ago. This boundary is, however, poorly established.

The Proterozoic rocks of the Wenecke Mountains region are closely juxtaposed against those of the Mackenzie Mountain area. Because of its contrasting thickness, facies, and structure, however, the Wenecke assemblage is thought to be older. It has close lithologic and stratigraphic similarities to the rocks of the lower and middle parts of the Belt Supergroup to the south.

Study of these regions has led to a scheme for subdivision of post-Aphelian supracrustals into three major packages, separated in most regions by unconformities. The oldest, called Sequence A, is between 1.7 b.y. and 1.2 b.y. old and includes the pre-Rae Group Proterozoic of the Coppermine Homocline, the lower and middle parts of the Belt-Purell Supergroup, the older Proterozoic assemblage in the Wenecke Mountains region, the Unkar Group of Grand Canyon, and the Apache Group of eastern Arizona. In all of these regions there is evidence of a major basic igneous event at about 1.2 b.y. ago. This event provides the means in most areas for dating the top of Sequence A. The second subdivision (Sequence B) is between 1.2 b.y. and about 0.8 b.y. It includes the Rae and Shaler Groups of northwest Canada, the older Proterozoic assemblage of the Mackenzie Mountains, the Missoula Group of the Belt, and possibly the Uinta Mountain and Chuar Groups of the western United States. The upper age limit of this sequence is poorly dated but in most regions is easily placed on the basis of geological relationships. Sequence C includes the Windermere Group and similar deposits that unconformably overlie older rocks along the length of the North American Cordillera.

Paleocurrent data, thickness, and facies changes all suggest westerly transport throughout deposition of Sequences A, B, and C. Continental separation has been inferred, along north-south fractures, in places ranging from the beginning of the early Proterozoic (2.5 b.y. ago) to the time of initiation of deposition of Sequence C (about 850 m.y. ago?). Whether oceanic crust was present in the Cordillera region in Proterozoic time is still a matter of speculation. Some evidence supports the alternative concept of basin floundering, related to regional tension during abortive (?) attempts at continental fragmentation.

INTRODUCTION

The complexities of Proterozoic stratigraphy in the North American Cordillera have been compounded by the dearth of meaningful radiometric age determinations and paleomagnetic data. Some of the problems in obtaining such critical data are due to rejuvenation phenomena, related to the preservation of these old
rocks in a young fold belt. Establishing a stratigraphic framework and "absolute" chronology are much more realistic propositions in stable shield areas. For this reason, the first part of this paper is a brief review of the stratigraphy of the northwestern part of the Canadian Shield. This is followed by a description of the Proterozoic stratigraphy of the western part of the Arctic Archipelago, the Mackenzie Mountains area of the northern Cordillera, and the Wrennecke Mountains region. In some of these areas the stratigraphic terminology for the Precambrian is at present inadequate. In order to discuss the stratigraphy of such regions, it has been found necessary to introduce some new stratigraphic terms. An attempt has been made to identify these new names throughout the text. The new names are introduced in a purely informal manner, but it is hoped that they will prove useful in erecting a formal stratigraphic scheme for these regions.

The description of the rocks in these areas is followed by a section concerned with establishing stratigraphic correlations in the northern part of the Cordillera. The final purposes of this paper are to take a broader view of Cordilleran Middle and Upper Proterozoic geology, to present evidence in favor of a threefold subdivision of post-Aphelbian supracrustal assemblages of western North America, and to speculate on the crustal history of the region.

**STRATIGRAPHY OF THE AMUNDESEN BASIN AREA**

**THE "BASEMENT" ROCKS**

The oldest rocks in the northwestern part of the Canadian Shield are the Archean rocks of the Slave Province (McGlynn and Henderson, 1970). This nucleus of ancient rocks (Figure 1) was stabilized at the end of Archean time. It is framed on the west by the lower Proterozoic (2.5 b.y. to 1.7 b.y.) succession of the Coronation Geosyncline, described in several papers by Hoffman (1968; 1969). Hoffman and others (1974) interpreted the faulted lower Proterozoic basins on the northeast and southern borders of the Slave Province as aulacogens. They considered these aulacogens to be failed arms of three-armed rift systems that were active during the development of the ocean basin that according to Hoffman (1973) became the Coronation Geosyncline.

Campbell and Cecile (1973), however, considered the lower Proterozoic rocks of the northern margin of the Slave Province (Figure 1) to be deposits of an intracratonic basin rather than an aulacogen. The East Arm of Great Slave Lake is the present site of what was described by Hoffman and others (1974) as the Athapuskow Aulacogen. Recently Badham (in preparation) has underscored the importance of large-scale dextral strike slip faults in the East Arm region, rather than normal faults as suggested by Hoffman and others (1974) for the early development of the lower Proterozoic basin in this area. The Coronation Geosyncline, according to Badham (in preparation), was situated at a long-lived continental margin. He considered the overlying dolostones and sandstones of the Wopmay Group (Figure 1) to have occurred at an Andean-type continental margin. Hoffman and others (1974) interpreted the Coronation Geosyncline in terms of a "Wilson Cycle" of continental fragmentation and ocean opening and closure. Resolving these important differences in interpretation may be crucial to understanding of the tectonic setting (intracratonic or oceanic) of the younger Proterozoic assemblages that are the main subject of this paper.

**AGE OF THE SHALE GROUP**

Supracrustal rocks and intrusive rocks in the northwestern part of the Wopmay orogen are overlain unconformably by the thick and complex succession of Proterozoic rocks that form the Coppermine Homocline (Figure 1). The sequence of rocks in the Coppermine Homocline, as documented by Baragar and Donaldson (1973), is shown schematically in Figure 2. This area provides the best data concerning the "absolute" ages of post-Aphelbian supracrustal rocks in northwestern Canada.

The Cameron Bay and Echo Bay Groups form the youngest deposits of the Coronation Geosyncline in this region. These supracrustal rocks are probably contemporaneous with the intrusion of the Great Bear Batholith (Fraser and others, 1970; Hoffman and others, 1974) about 1.74 b.y. ago. At the southern margin of the Coppermine Homocline, sedimentary and igneous rocks of equivalent age (about 1.7 b.y.) are unconformably succeeded by sandstones and dolostones of the Hornby Bay Group (Baragar and Donaldson, 1973).

Emplacement of the Muskox Complex about 1.2 b.y. ago (T. N. Irvine, quoted by Baragar and Donaldson, 1973) within the Hornby Bay Group apparently occurred before deposition of the unconformably overlying sandstones and sandstones of the Dismal Lakes Group. Deposition of these shelf deposits was followed by extrusion of the thick basaltic sequence of the Copper Creek Formation. These volcanic rocks have been assigned an age of 1.21 b.y. (Baragar and Donaldson, 1973). They are overlain by sandstones, siltstones, and lavas of the Husky Creek Formation that in turn overlain, with angular discordance, by the Rae Group (Figure 2).
The Rae Group consists of a lower clastic unit (siltstones, sandstones, and shales) and an upper, dominantly carbonate-rich unit. Baragar and Donaldson (1973) further subdivided these into a total of six mappable units. According to J. Derby and J. Dixon (personal communication, 1976, 1977) the upper two divisions of the Rae Group are actually Paleozoic. Detailed documentation is required before these problems can be fully resolved. The Rae Group is intruded by an extensive suite of basic sills that have given potassium-argon ages ranging from 605 m.y. to 718 m.y. If these dates are accurate, then some limits may be placed on the timing of events in the Coppermine Homocline.

To the north of the Coppermine Homocline, in the western part of the Arctic Archipelago, the Minto Arch forms a spine of gently folded Precambrian rocks flanked by Paleozoic and younger deposits. The Precambrian assemblage is known as the Shaler Group (Thorsteinsson and Tozer, 1962) and consists of shelf carbonates (commonly stromatolitic), mature sandstones, evaporites, and silty mudstones. The Shaler Group was divided by Thorsteinsson and Tozer (1962) into five formations. The lowest formation unconformably overlies schistose quartzites that are intruded by granitic rocks dated by the potassium-argon method at 2.4 b.y. (Thorsteinsson and Tozer, 1962). The Shaler Group is disconformably overlain by pyroclastics and lavas of the Natakusk Formation. Petrographic similarities, palynological results, and radiometric age determinations all support

Figure 1. Sketch map to show some features of the northwest part of the Canadian Shield. Western border of the exposed shield is shown by the line of squares. Lower Proterozoic rocks of the Coronation Geosyncline and associated basins are shown by the stippled ornament. Younger Proterozoic rocks of the Coppermine Homocline are shown in black. Major faults are indicated by the letters ‘‘F. - F.’’ The line of triangles in the west shows the approximate eastern margin of the Cordilleran fold belt.
Figure 2. Schematic representation of the Proterozoic stratigraphy of some areas of northwestern Canada. Note that the dated sequences of the Canadian Shield (Coppermine area and Victoria Island) are correlated with the undated rocks of the "Mackenzie Mountains Supergroup." Lava flows of the Coppermine River Group and the intrusive Muskox Complex in the Coppermine region form part of the extensive basic igneous event at about 1.2 b.y. ago. Location of these areas is shown in Figure 3. The small numerals in the Little Dal Group refer to subdivisions proposed by Aitken and others (in press) as follows: (1) mudcracked subunit; (2) basinal sequence; (3) grainstone subunit; (4) gypsum subunit; (5) rusty shale subunit (6) upper carbonate subunit. See text for explanation.
the contention of Christie (1964) and Palmer and Hayatsu (1975) that the Naktusiak volcanics are contemporaneous with the Coronation sills and other intrusions of the Franklin episode (Fahrig and others, 1971) about 650 m.y. to 700 m.y. ago. According to these data the Shaler Group could be any age from 2,400 m.y. to 700 m.y.

Subdividing the lower formations of the Shaler Group has led to a detailed stratigraphic correlation between the upper Proterozoic successions of Banks and Victoria Islands and those of the mainland to the south (Block Inlier and Rae Group of the Coppermine Homocline). The proposed correlation of the Shaler and Rae Groups permits much more precise dating of the former (between 1,200 m.y. and 700 m.y.). Details of this correlation and history of previous research are given in Young and Jefferson (1977) and Young (1977). For comparison a brief description of the stratigraphy of the Shaler Group is given below. For more detailed descriptions of individual units see Young (1974), Young and Jefferson (1975), Young and Long (1977a; 1977b), and Young (1977).

STRATIGRAPHY OF THE SHALER GROUP

The stratigraphic sequence of the Shaler Group was established by Thorsteinson and Tozer (1962). This work has been substantiated and refined by subsequent studies in the area (Young, 1974; Miall, 1975; Baragar, 1976; Palmer and Hayatsu, 1975).

Glenelg Formation

The Glenelg Formation may be subdivided into several widespread units. These extensive units probably merit formal status. On Victoria Island the lowest unit (Figure 2) is mainly fine clastics, with subordinate sandstones and carbonates; it has not, however, been studied in detail. The suggested correlatives in the Coppermine region (Figure 2) is better exposed and has been further subdivided (Baragar and Donaldson, 1973).

The overlying carbonate-dominated unit on Victoria Island and Banks Island was recently described by Young and Jefferson (1975) and Miall (1976). It is made up predominantly of gray to buff-weathering, fine-grained gray dolostone. Gray and black chert are common constituents, both as thin beds and as an early diagenetic replacement of stromatolites and bedded dolostones. Shale-carbonate cycles with abundant evidence of shallow-to-emergent conditions were noted in this unit by Young and Jefferson (1975). This unit also contains a characteristic suite of stromatolites (Jefferson, 1977) including cf. Bathypaera, Comophyton and large columnar branching forms.

The overlying fine-clastic unit is best exposed on Banks Island where it comprises a monotonous sequence of fine to wavy bedded green, gray and buff silstones and mudstones. The most common sedimentary structure is small-scale cross-bedding (ripple cross-lamination).

The fine-grained interval is succeeded by a thick accumulation of coarser grained clastic rocks. This sandstone-dominated sequence may be subdivided into two units. The lower of these is made up of sandstones that are pink and buff and have ubiquitous cross bedding. They are interbedded with purple silstones-mudstones and are arranged in a series of fining-upward cycles interpreted by Young and Jefferson (1975), Miall (1976), and Young (1977) as fluvial deposits. The upper part consists of white orthoquartzites and subordinate dark gray shales. Both the sandstones and shales are characterized by the presence of convoluted bedding, and widespread development of small-scale "pipe structures" may be related to dewatering processes (Young, 1977). The upper unit is probably a shallow marine deposit with possible local deltaic influence.

A thin shaly interval above these sandstones is followed by an extensive orange-weathering stromatolitic dolostone that was taken by Thorsteinson and Tozer (1962) as the topmost unit of the Glenelg Formation. This subdivision is made up mainly of stromatolites but also includes flat-chip conglomerates, minor gray shales, and fine-grained carbonates. Detailed studies of the shape, orientation, and possible paleo-geographic significance of these stromatolites were made by Young (1974) and Young and Long (1975). Variants of these stromatolites resemble Inseria, Gymnosolen, and Baculites, with the overall name cf. Inseria being applied to the biostrome (Jefferson, 1977).

Reynolds Point Formation

Accumulation of the shallow marine carbonates of the upper Glenelg Formation was terminated by influx of fine terrigenous material forming the overlying green, gray, and purple mudstones. The upper part of this terrigenous unit is composed mainly of cross-bedded sandstones together with minor silstones and mudstones showing evidence of shallow water-to-emergent (in places evaporitic) conditions. All of these terrigenous rocks constitute a major coarsening-upward sequence interpreted as a prograding marine deltaic complex by Young and Long (1977a).

The thick carbonate-dominated unit that makes up the main part of the Reynolds Point Formation was recently described by Young and Long (1977b). This unit contains a basal member of dark gray shaly micrite and dolomicate, with the development of spectacular thick and extensive stromatolitic biostromes. These car-
bonates are succeeded by a thick oolitic sequence with associated smaller stromatolite bioturms. The oolite beds alternate with wavy bedded shaly dolomitic units. ‘Molar-tooth’ structure (Smith, 1968) is abundant in shaly carbonates of the Reynolds Point Formation. Stromatolites resembling Gymnosophen, Batacuta, and Beuxia are present (Jefferson, 1977). In the upper part of the formation there are fine sandstones of probable tidal origin (Young and Jefferson, 1975). These clastics are overlain by the uppermost, thin member of the Reynolds Point Formation — stromatolitic and fine-grained carbonates that pass by transition into the evaporitic sequence of the overlying Minto Inlet Formation. Stromatolites in this unit resemble Acacella (Jefferson, 1977).

Minto Inlet Formation

The upper units of the Shaler Group are more poorly exposed than those below and have received less detailed study. The Minto Inlet Formation is characterized by abundant gray and white pure and impure gypsum and anhydrite beds. Some of these are oolitic and cross-bedded. ‘Chicken wire’ texture is common and euturbation folds are present in the evaporites. Associated beds include some stromatolites in gray limestones and dolostones. The carbonates range from mudstones to flat-chip conglomerates or grainstones. They are commonly intimately associated with the evaporites.

Wynnott Formation

The Wynnott Formation consists largely of gray and buff limestones and dolostones with some shales and evaporites. Most of the carbonate rocks are fine grained, but intraformational conglomerates (with fragments up to cobble size) are present. At least one unit of the upper Wynnott contains oolites and a spectacular development of oncoids up to 30 centimeters in diameter. Branching columnar stromatolites are present at several horizons. They commonly grow on the large oncoids. These forms resemble Gymnosophen and Batacuta. Some of the carbonates have striking exposures of desiccation cracks. Molar-tooth is common. About two-thirds up the formation there is an unusual, extremely finely laminated black, coaloid shale unit. The overall characteristics of the Wynnott Formation suggest deposition in a shallow marine environment.

Kilian Formation

The Kilian Formation in the northern part of Victoria Island (Figure 3) can be subdivided into a lower red silstone-mudstone-evaporite unit overlain by a shallow marine assemblage consisting of red and green mudstones, dolostones, and limestones (micrite and flat-chip conglomerates), cross-bedded and laminated sandstones, and desiccation-cracked mudstones. Carbonates of this unit also include branching columnar stromatolites resembling Gymnosophen and Batacuta. A thin lava and pyroclastic unit was discovered in the summer of 1977 in the upper part of the Kilian Formation. This unit is potentially very important for it may provide a much-needed ‘absolute’ date for deposition of the upper part of the Shaler Group. Paleomagnetic study of this unit has been undertaken by W. R. Morris.

The stratigraphy of the Kilian Formation appears to be different in the southern part of Victoria Island (Thorsteinsson and Tozer, 1962) where there is a cross-bedded sandstone at the top. These sandstones overlie evaporitic rocks which may be the lower member of the Kilian Formation or may be an evaporitic facies equivalent to the nonevaporitic upper Kilian to the north. This problem requires detailed study.

The rocks of the Shaler Group are cut by basic intrusive rocks of the Coronation suite (625 m.y. to 700 m.y.), which is thought to be comagmatic with the Nautskis lavas (Christie, 1964; Palmer and Hayatsu, 1975; Baragar, 1976) that disconformably overlie the Kilian Formation.

PROTEROZOIC STRATIGRAPHY OF THE MACKENZIE MOUNTAINS AREA

The tectonic framework and stratigraphy of the Proterozoic of the Mackenzie Mountains have been described by Gabrielse and others (1973), Aitken and others (1973), Aitken and Cook (1974a, 1974b), Eishacher (1976, 1977) and Aitken (1977). The Proterozoic comprises two major sequences, here informally designated as ‘Mackenzie Mountains Supergroup’ and ‘Ekwi Supergroup.’ The ‘Mackenzie Mountains Supergroup’ is composed dominantly of terrigenous to shallow marine clastics, carbonates, and evaporites. The ‘Ekwi Supergroup’ records a marked difference in sedimentation style with rhythmites, mixites, glacial deposits, iron-formation, shales, and shallow-marine clastics and carbonates. The boundary between the two supergroups is variable. Commonly it is an angular unconformity, but in some areas the discordance is slight and only apparent after mapping. Evidence from other areas, however, suggests a conformable transitional contact.

‘MACKENZIE MOUNTAINS SUPERGROUP’

Map-Unit H1

Map-Unit H1 is the oldest reported formation of the
"Mackenzie Mountains Supergroup." It constitutes up to 400 meters (Aitken and others, in press) of pale gray, locally cherry dolostone in several isolated localities; the base is not exposed. Stromatolites include several stratal-form varieties, Conophyton, and some columnar branching forms. See Aitken and others (1973) for a fuller description.

Tsezothe Formation

The Tsezothe Formation (Gabrielse and others, 1973) is a predominantly argillaceous sequence up to 1,500 meters thick. The formation can be subdivided on a regional scale into two members: a lower member dominated by gray argillaceous rocks and an upper member characterized by varicolored (red, green, gray) argillite, sandstone, and minor carbonate (Aitken and others, in press). A locally developed stromatolitic carbonatic member is present between these two members in the vicinity of Mount Edoni. Collectively the Tsezothe Formation is a coarsening-upward sequence, representing gradual basin filling or shallowing. It comprises a large number of smaller-scale, coarsening-upward cycles. The bulk of the lower member was deposited in a subtidal environment, at or near wave base, with local evidence of shoaling in the form of thin, discontinuous, sandstone bodies deposited as offshore shoals and barrier islands. The upper member is characterized by deposits of littoral environments, with small-scale cycles reflecting local shoaling in the form of tidal flat, lagoonal, barrier island, and biothermal development. Ample evidence of emergence is present in the upper member in the form of abundant desiccation cracks, and locally by the presence of paleosols.

Katherine Group

The recessive-weathering clastics of the Tsezothe Formation are conformably overlain by resistant sandstones of the Katherine Group, locally as much as 2,000 meters thick. The Katherine Group is divisible regionally into three units of formational rank (Aitken and

Figure 3. Sketch map to show the distribution of some post-Aphelian Proterozoic rocks of northwestern Canada.
others, in press) which can be traced into the type section of the Tigonankwene Formation (Gabrielse and others, 1973). The lower formation is further divisible (due to the presence of recessive units) into five distinctive members, at least in the northwestern half of the Mackenzie Mountains fold belt. Of these, only two can be traced with any certainty into the type section of the Tigonankwene. The group is interpreted as the product of a complex interaction of fluvial and marine processes. Sandstone members represent, in part, the deposits of rapidly prograding fluvial-deltaic complexes, which advanced at least four times over shallow marine, braided fluvial, and lacustrine environments. The phase of each of these major progradations is marked by the development of extensive ripple-laminated sandstones or large-scale, cross-stratified marine sand sheets, which are in turn overlain by argillaceous rocks with or without carbonates. The upper two recessive units of the Katherine Group (K4 and K6 of Atikten and others, in press) are host to stromatolitic dolostones. In K4 these include stratiform and large columnar forms, which were observed only in the type section of the Tigonankwene Formation; further north this unit is characterized by purple shales, silts tones, and sandstones with only minor carbonates. In unit K5 two stromatolitic horizons have been recognized; near the middle of the formation these include thin bioheral development of cf. Batalia and similar but unbranched stromatolites, and near the top, a distinctive orange weathering form of cf. Insera (Atikten and others, in press), which occurs in both scattered and contiguous (in places elongate) bioherms.

Little Dal Group

The Little Dal Group, up to 2.8 kilometers thick, is carbonate dominated and lithologically varied. It has been subdivided into six informal "subunits" of formation rank (Atikten, 1977; Atikten and others, in press). No significant disconformities have been recognized.

The basal, "mudcracked subunit," mainly shale and sandstone, records initial deepening following the deposition of the peritidal, terminal Katherine quartzites, followed by cyclical shallowing.

The succeeding "basal sequence" records marked deepening. It is dominated by red and gray, subtidal mudstones with abundant limestone nodules in its lower part, followed by deepwater, limestone-shale rhythmites, the latter locally yield Chwarta and primitive metaphytes (Atikten and others, in press). Debrisflows and slide-masses record slope environments. Spectacular reefs of stromatolitic limestones and dolomite up to 500 meters high are limited to the western part of the exposed basin facies, west of Mountain River. There, wedges of coarse, reef-derived talus indicate a growth relief of at least several tens of meters and a corresponding minimum depth for the basin floor. Southeast of Mountain River, the development of platformal facies is signalled by the appearance of two or more subregional stromatolite biostromes. Equivalents of the "basinal sequence" in the area of the type section of the Little Dal Group are dominated by grani stones, algal stromatolites, and shale carbonates with molar-tooth structure which record a platformal environment. In the type section area, stromatolites resembling Batalia and, less commonly, Gymnozoen and Boxonia, compose bioherms up to 150 meters thick and several kilometers wide. Nodular shale carbonates similar to those of the "basinal sequence" are laterally equivalent to the bioherms.

The "basinal sequence" and platformal equivalents are gradationally overlain by the "grainstone subunit," a regional blanket of shallow-water carbonate rocks. Two lithofacies dominate: dolomitized ooid grainstone, and plaly, mudcracked sandy dolomites of high-inter tidal origin. The latter increases progressively in importance southeastward, toward the type section of the group.

The "grainstone subunit" is overlain by up to nearly 500 meters of bedded, mainly white gypsum. Southeastward thinning of this subunit to 50 meters or less is accompanied by an increase in clay content and fine clastics.

The "rusty shale subunit" gradationally overlies the "gypsum subunit." Rock types include gray, greenish gray, black and purplish shales; mudcracked, ripple marked and lenticular bedded quartz rite sandstones; and orange silty carbonates bearing a distinctive stromato lite resembling Batalia. All of these rock types are notably pyritic and rusty weathering.

At the type section of the Little Dal, Gabrielse and others (1973) placed the division between the Lower and Upper Little Dal at the base of a shaly unit which corresponds in part to the "grainstone subunit," and in part to the "gypsum subunit" (see above). However, in keeping with the above subdivisions, the term "upper carbonate subunit" will be used to refer to the cliff-forming carbonates above the "rusty shale subunit." The reestablishment of peritidal conditions is recorded in up to nearly 900 meters of these carbonates that consist of oolitic grainstones, finely laminated shaly calcilutites with molar-tooth structure, stromatolites resembling Gymnozoen, Batalia, Boxonia, Conophyton as well as simple domal and stratiform types, nodular limestones similar to those of the "basinal sequence," and minor red shales and sandstones. Stability of the basin during deposition of the "upper carbonate subunit" is indicated by correlation of units
ranging in thickness from 20 centimeters to 30 meters for more than 300 kilometers along strike.

The top 100 to 300 meters of the “upper carbonate subunit” records evidence of shallowing and instability with lateral facies and thickness changes in stromatolitic carbonates, sandstone and red to green and gray mudstone members and volcanic flows and their reworked derivatives. Mafic dikes and silts that are common in underlying formations are absent above these volcanics and may have been part of the same igneous event. Shallowing, renewed clastic deposition, and renewed restriction of marine circulation are indicated in the uppermost Little Dal by desiccation-cracked red mudstones, teepee structures, casts of evaporite minerals, and gypsiferous limestone. Simple domal stromatolites are common in units that are interpreted as intertidal.

Redstone River Formation

This formation comprises evaporites and easterly derived teregtinous clastics that accumulated in embayments of a northwest-trending shoreline. The evaporites are at least 200 meters thick in basal sections, but pinch out shoreward. At least 600 meters of red mudstones overlie the evaporites in basinal sections. These thin markedly towards the northeast and towards “paleo-headlands” where they are replaced by sandstones, pebbly sandstones, conglomerates, and breccias which also pinch out toward the northeast. The majority of clasts in the conglomerates are derived from the Little Dal Group. However, the conglomerates are conformable with the Little Dal in regions overlying the Plateau Thrust Fault, so that the erosional surface must have existed farther to the northeast, where it later became a component of the more extensive sub-Rapitan unconformity. In some localities the overlying Coppercap Formation is separated from Little Dal carbonates by red mudstones (Redstone River Formation) less than 1 meter thick.

Coppercap Formation

The Coppercap represents relatively deep-water sedimentation of clastic carbonates and shales. It is about 200 to 300 meters thick in complete sections and constitutes three subunits which, in ascending order, are as follows: up to 60 meters of rhythmically bedded, orange-weathering clastic and algal dolostone; black shales and shaly limestones; and cliff-forming algalitic limestones and very finely laminated silicified limestones. In the Coates Lake area the basal contact is interbedded up to seven times; elsewhere it is sharply gradational. In most localities the basal carbonates are cryptagal-laminated, contain nodular structures that could be interpreted as fenestral or gypsum pseudo-morphs, and are variably copper mineralized.

In most places the Coppercap is unconformably overlain by the Rapitan. However, where the Upper Little Dal-Redstone River-Coppercap sequence is thickest, continuous sedimentation is suggested by a gradational to interfingered contact. Minor evaporites occur at the contact in the Coates Lake area.

“EKWI SUPERGROUP”

The youngest sequence of Proterozoic strata in Mackenzie Mountains comprises the Rapitan Group and the overlying Reeve and Sheepbed Formations. Although there is local evidence for a conformable contact with the underlying Coppercap Formation, the base of the Rapitan is for the most part an unconformity of regional proportions (Gabrielse and others, 1973; Aiik and others, 1973). With up to 4.5 kilometers of strata missing regionally and more than 200 meters eroded locally, this is the major stratigraphic discontinuity in the Proterozoic sequence in the Mackenzie.

The tectonism associated with this discontinuity began earlier, at about the time of initiation of deposition of the Redstone River Formation (Eibachner, 1976), but the regional stratigraphic break occurs at the base of the Rapitan. For this reason and to be consistent with usage in other parts of the Cordillera where the major stratigraphic break is placed at the base of a mixite-bearing unit, the Proterozoic assemblage of the Mackenzie Mountains region is subdivided into two sequences of supergroup rank. As noted above, the lower one is called the “Mackenzie Mountains Supergroup;” the name “Ekwisupergroup” is suggested for the upper sequence, which includes the Rapitan Group and all the overlying Proterozoic rocks. The stratigraphic nomenclature used in this section is shown in Table 1.

Rapitan Group

The Rapitan Group, which contains one of the largest iron deposits in North America, crops out discontinuously for about 630 kilometers in an arcuate north- and northwest-trending belt. Two major basins of preservation are recognized. These are the Snake River Basin and the Mountain River-Redstone River basin. Three unnamed formations compose the Rapitan Group as currently defined (Gabrielse and others, 1973). In accordance with anticipated formal definition and redefinition of the Rapitan in preparation (Eibachner, 1977), a tentative stratigraphic scheme is informally presented here (Table 1). As in the previous section informal stratigraphic terms, necessary to facilitate descriptions, will be enclosed by quotation marks.
"Thundercloud Subgroup"

The provisional name "Thundercloud Subgroup" is used to refer to the part of the "Rapitan Group" that contains evidence of glacial influence on sedimentation. It includes the "Snake River," "Sayunei," and "Sheral" Formations.

"Sayunei Formation" (Lower Rapitan)

The lowermost formation of the Rapitan in the Mountain River-Snake River basin, the more northerly of the two outcrop areas shown in Figure 3, the "Sayunei Formation," contains iron-formation at or towards its top. At least three members can be recognized. The "Mount Berg member," the lowest, is made up of green mixtures and mudstones with some orthoconglomerate. The member is known only from Thundercloud Range. The "Lukas Creek member," consists mainly of maroon mudstone rhythmites with sandstone, orthoconglomerate, and minor mixture. It is the principal member of the "Sayunei Formation."

The rhythmites include both grain-flow deposits and turbidites. Other common features include small-scale cross-lamination and load casts. The orthoconglomerates are in irregular lenses and are commonly graded; these features suggest deposition as channel deposits. Where this is the basal member, thick orthoconglomerates are common at the base. The "Mountain River member," which consists of maroon mixture with lesser amounts of orthoconglomerate, mudstone, and sandstone, is the uppermost member. Except for reddish color, it is lithologically and structurally similar to the overlying "Sheral Formation." The iron-formation typically occurs as thin beds or lenses in the uppermost part of the "Lukas Creek member" and the lowermost part of the "Mountain River member." Evidence of glacial activity, including striated and faceted clasts, dropstones, "till pellets," till balls (?) and possible ice contact features, is common throughout the "Sayunei Formation." (Young, 1976; Yeo and Young, 1976; Yeo, 1977; Eshbach, 1977). No evidence for contemporaneous volcanism (e.g. Condon, 1964) has been confirmed. The thickness of the "Sayunei Formation"

<table>
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<tr>
<th>Supergroup</th>
<th>Group</th>
<th>Subgroup</th>
<th>Formation (Snake R. basin)</th>
<th>Formation (Mountain R.-Redstone R. basin)</th>
<th>Member</th>
<th>Dominant Lithology</th>
<th>Formation</th>
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<td>&quot;Knott&quot;</td>
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<td>&quot;Rapitan&quot;</td>
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<td>&quot;Twinya&quot;</td>
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<td>&quot;Thundercloud&quot;</td>
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<td>&quot;Sheral&quot;</td>
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<td>&quot;Middle Rapitan&quot;</td>
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<td>&quot;Snake River&quot; (maroon mixture)</td>
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<td></td>
<td>&quot;Lukas Creek&quot;</td>
<td>maroon mudstone rhythmite</td>
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<td>&quot;Mount Berg&quot;</td>
<td>green mixture</td>
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Table 1. Provisional stratigraphic terminology for the uppermost Proterozoic rocks in the Mackenzie Mountains. Informal names are enclosed by quotation marks. The terms "Twinya," "Sheral," and "Sayunei" were used orally by Eshbach (Eshbach and Gabriele, 1977). The term "Rapitan" is used in an expanded sense as suggested by Eshbach.
is highly variable. It is thickest in the Mountain River area where it is up to 730 meters thick (Aitken and others, 1973). It generally thickens westward.

"Snake River Formation"

The "Snake River Formation," the homotaxial equivalent of the "Sayunei Formation" of the southern basin differs somewhat in gross lithology. It was first described by Ziegler (1959) as part of his "Snake River Tillite." It comprises both massive and stratified maroon and brown mixtures with orthoconglomerate and sandstone lenses and sandstone beds throughout. As in the "Sayunei Formation," features characteristic of glacial deposits are common. Carbonate, sandstone and greenstone class are found as are intraformational mudstone class and till balls (?). The maximum thickness of the "Snake River Formation" is not known but must be more than 730 meters. The largest deposit of iron in the Rapitan Group, the Crest deposit (Stuart, 1963; Gross, 1965a, 1965b, 1973), occurs in this formation. It is estimated at more than 20 billion metric tons averaging 47.2 percent iron (Stuart, 1965). The iron formation is found in up to five stratigraphic zones, each of which may be several meters thick. The "Snake River Formation" is generally more recessive than the "Sayunei Formation."

"Shezal Formation" (Middle Rapitan)

Overlying the lower Rapitan is a moderately recessive, thick gray mixite unit informally named the "Shezal Formation." In most places the basal contact is unconformable (Gabrielse, 1973), but locally it is conformable. This is the unit described by Ziegler (1959), in the earliest report on the Rapitan, the "Bonner Plume River tillite." Massive mudstones, massive and cross-laminated sandstones, and orthoconglomerate lenses are present. Features typical of Pleistocene water-laid tills or flowwalls are common in these mixites, which are both stratified and massive. Clast composition varies locally and vertically, reflecting the lithology of locally underlying units. Intraformational mudstone clasts are common, but maroon clasts are notably absent. Jasper and hematite clasts locally occur near the base. A glacial origin is ascribed to this unit also (Ziegler, 1959; Young, 1976; Eibacher, 1976, 1977). Its thickness is quite variable, over 775 meters near Snake River. It oversteps the lower Rapitan to the west and southwest and thins both eastward and westward across strike.

Glacial Deposition of the "Thundercloud Subgroup"

Features common in the (provisional) "Thundercloud Subgroup" which are typical of glacial deposits include: dropstones, striated and faceted clasts, till balls, possible ice-contact features, and crudely stratified mixites (water-laid tillites). Whereas none of these features is diagnostic evidence for glacial activity (Scherronhorn, 1974), their collective presence is very persuasive. The occurrence of till pellets (Young, 1976) is thought to be diagnostic of glacial deposition (Owen, 1970). No striated basement typical of terrestrial glacial deposits has been found.

Paleocurrent measurements give variable directions but indicate transport generally from the east and northeast. The "Thundercloud Subgroup" is interpreted as a marine glacial deposit as Ziegler (1959) originally suggested. Evidence of slumping and mass flow is common. Although such features led some earlier workers (Upitis, 1966; Condon, 1964; Gabrielse and others, 1973) to conclude that the Rapitan was not glacial, they are not incompatible with a glacial marine environment and are common in some Pleistocene glacial deposits (e.g. Eivason and others, 1977). Tuffaceous sediment has been reported from the Rapitan (Condon, 1964; Gross, 1973), but definite evidence of contemporaneous volcanism is not known. Volcanic rocks are, however, present in the upper Tindir Group (Mertie, 1973), which is probably in part a Rapitan equivalent in Alaska.

"Knot Subgroup"

The term "Knot Subgroup" is used here as a provisional name for what has previously been called the Upper Rapitan ("Twitya Formation" of this report) and the overlying Keele Formation.

"Twitya Formation"

Abruptly overlying the "Shezal Formation," the "Twitya Formation" consists of recessive-weathering, laminated gray mudstones with occasional sandy beds. It is commonly interfingered with the overlying Keele Formation, although in some areas the contact is quite sharp. There is no evidence for glacial activity during deposition of this unit. Ripples, wave marks, small-scale cross-lamination, and slump folds are common. The "Twitya Formation" oversteps the "Shezal" to the west and thins eastward. It nowhere overlaps the "Snake River-Sayunei Formation" directly. It is very thick, up to 1,170 meters near Redstone River (Gabrielse and others, 1973). The "Twitya Formation" represents a return to normal marine conditions after melting of the glacial ice. The increased abundance of coarser clastics upwards may reflect isostatic rebound. Paleocurrent data indicate highly variable transport directions with an overall suggestion of an easterly and northeastward source. Measurements from the northwest part of Snake River basin indicate transport from the west, suggesting
that the original basin of deposition was not very wide, at least in the northwest.

Keele Formation

Above the "Twinya Formation" and transitional to it, the Keele Formation consists of rhythmic, thick-bedded sandstone, carbonate (partly stromatolitic), and mudstone, with some orthoconglomerate. The latter commonly appears to be recemented. Sole markings, cross-stratification, and mudcracks are typical structures of the Keele Formation. It commonly appears as a strikingly banded, resistant cap-unit above the Rapitan Group. It is up to 610 meters thick (Gabrielse and others, 1973). Transport was dominantly westward. As the Keele and "Twinya Formations" are not separated by a stratigraphic break, except perhaps locally (Gabrielse and others, 1973), it is suggested that the Keele Formation be included within the Rapitan Group as part of the nonglacigene "Knorr Subgroup."

Sheepbed Formation

The uppermost Proterozoic unit, the Sheepbed, is also the least studied. It is a very recessive mudstone, up to 760 meters thick southwest of Coates Lake (Gabrielse and others, 1973). Minor sandstone and dolostone interbeds occur. It lies conformably on the Keele and is unconformably overlain by Lower Cambrian sedimentary rocks.

PROTEROZOIC STRATIGRAPHY OF THE WERNECKE MOUNTAINS AREA

The location of the Wernecke Mountains region is shown in Figure 3 and Figure 4 (area 5). This lies west of the area of outcrop of the "Mackenzie Mountains Supergroup" and "Ekwi Supergroup." The major problem in the Wernecke region is the establishment of stratigraphic relationships between the Proterozoic rocks exposed there and closely juxtaposed Proterozoic rocks (of very different aspect) of the other two supergroups.

WERNECKE SUPERGROUP

The lowermost Proterozoic succession that is exposed in the Wernecke Mountains (Figure 5) has an aggregate thickness in excess of 15 kilometers. On the basis of detailed stratigraphic studies undertaken during the summers of 1976 and 1977, the stratigraphy established by Green (1972), Blusson (1974), and Norris (1975) was revised by Bell and Delaney (1977). Three subdivisions of group status were defined, and these were designated A, B, and C from oldest to youngest. These subdivisions have now been given informal names (Delaney, in preparation): the A group is now referred to as the "Fairchild Lake Group"; the B group as the "Quartet Group"; and the C group as the "Gillespie Lake Group." Because this entire succession has long been referred to as either the Wernecke-type Proterozoics or the Wernecke succession, it has now been given the informal name the "Wernecke Supergroup" (Delaney, in preparation).

"Fairchild Lake Group"

The "Fairchild Lake Group" consists of at least 4,000 meters of light gray-weathering, thinly bedded to laminated siltstones and silty shales (Figure 5). Lenticular to wavy bedding is the most prominent style within this sequence of fine-grained sediments, although locally parallel planar bedding is present. Near the middle of this sequence of rocks is a member characterized by a thinly, rhythmically interbedded limestone and siltstone rock that weathers in a ribbed fashion.

The "Fairchild Lake Group" outcrops mainly as thin, commonly deformed and altered strips along the Bonnet Plume River area or in thin reentrants from the main zone (Figure 6). The only fairly complete sequence of this group is found southwest of Fairchild Lake, where a seemingly continuous sequence of about 3,500 meters of overturned sedimentary rocks has been documented (Delaney, in preparation). Most rocks of the "Fairchild Lake Group" are relatively undeformed, although locally a closely spaced bedding plane cleavage masks primary features in the rocks. Sillicification related to quartz veining is sporadic in occurrence throughout the sequence. Coarse-grained bladelike crystals of tremolite-actinolite are found as coatings on joint surfaces of rocks of the "Fairchild Lake Group." Along the Bonnet Plume River, near intrusive brecia complexes (Delaney and others, this volume), these fine-grained sedimentary rocks have been transformed into phyllites and schists, which commonly contain porphyroblasts of chloritoid.

Detailed correlation of the "Fairchild Lake Group" over any great distance is hindered by limited exposure, lack of distinctive marker horizons, and the effects of metamorphism. The nature of the contact between the "Fairchild Lake Group" and the "Quartet Group" is uncertain, as it is everywhere masked by structural complications.

"Quartet Group"

The "Quartet Group" consists of at least 5,000 meters of dark gray-weathering shales, shaly siltstones, and siltstones with some fine-grained sandstone beds (Figure 5). The base of the "Quartet Group" is marked by the presence of a 5 to 15 meter-thick white-weather-
Figure 4. Sketch map (modified from Stewart, 1972) to show the distribution of rocks of Sequence A (1.7 b.y. to 1.2 b.y.), Sequence B (1.2 b.y. to 0.8(?) b.y.), and Sequence C (0.8(?) b.y. to 0.6 b.y.). Sequence A rocks are shown by vertical ornament. The closer spacing and thicker dark lines in more westerly areas are meant to signify transition to deeper water conditions in that direction. Rocks of areas 12 and 13 are mainly continental. Sequence B rocks are shown by diagonal lines. These are mainly shallow marine platformal assemblages and in many areas also include evaporites. In the Mackenzie Mountains region (area 4) there is some suggestion of transition to deeper water conditions. Sequence C is shown by the dotted ornament. Note the contrasted distribution of rocks of Sequence C compared with those of the older sequences. The older sequences are preserved in smaller west-trending reentrants that in some cases reflect the original basin configuration. Arrows show generalized paleocurrent directions. See the text for details of individual areas and references.
ing unit consisting of thinly bedded limestone. Overlying this distinctive marker unit is a sequence of pyritic shales and silty shales, characterized by parallel planar laminations. These rocks gradually coarsen upward as a series of cycles composed of mudstone, siltstone, and fine-grained sandstone. A typical sequence in the upper part of the "Quartet Group" consists of dark gray weathering, medium to thick beds of wavy bedded mudstones that alternate with units of "clean" siltstone characterized by low angle cross-bedding. Many of these siltstone sequences terminate laterally in large-scale ball and pillow structures. Medium to thick units of massive silty shale with granule- to pebble-sized fragments of siltstone forming an open framework are scattered throughout this sequence. Certain aspects of this unit suggest that it may have been reworked. Common sedimentary structures in the "Quartet Group" include small-scale cross-bedding, load structures, sole marks, "egg-carton" ripple marks, and synaeresis cracks.

![Composite stratigraphic section of the oldest Proterozoic rocks exposed in the Wernecke Mountain region. Terminology is provisional (Delaney, in preparation). Note the dominantly fine-grained nature of the lower two groups. The upper part of the "Gillespie Lake Group" represents a platform sequence of carbonate rocks. Relationships between the "Fairchild Lake Group" and "Quartet Group" are uncertain, but there is a gradational boundary between the "Quartet Group" and the "Gillespie Lake Group."](image-url)
The lack of distinctive marker horizons in the “Quartet Group” make it virtually impossible to correlate this thick monotonous unit over great distances. Numerous low angle thrusts cause a repetition of sequences and further complicate stratigraphic reconstruction.

“Gillespie Lake Group”

The “Quartet Group” is transitional into the overlying “Gillespie Lake Group,” although the thickness and characteristics of the transitional unit are quite variable. At some localities, such as due south of Gillespie Lake (Figure 6), the transition from “Quartet” to “Gillespie Lake Group” occurs within a stratigraphic interval of 25 meters; north of Bear River the transitional sequence is at least 350 meters thick. At the latter locality the transitional unit assumes a distinctive striped appearance which is caused by alternation of thin to medium beds of orange-weathering dolosiltstone and gray, wavy bedded silestones.

In addition to the transitional facies, the “Gillespie Lake Group” includes three other major stratigraphic units (Figure 5). Above the transitional sequence is 700 to 1,000 meters of buff to buff gray-weathering,

![Geological sketch map to show the distribution of the rocks composing the “Wetnecke Supergroup.” Note that the “Fairchild Lake Group” occurs in a narrow zone along the Bonnet Plume River and in small reentrants from that zone. Note also the abundant faults that pose problems in reconstruction of the stratigraphic succession in this area.](image-url)
parallel planar-laminated to cross-laminated dolostones which contain thin beds and lenses of dark gray chert. Near the top of this sequence are several medium to thick beds which contain stromatolitic debris and, locally, intact columnar stromatolites. Conformably overlying the preceding sequence is 500 to 600 meters of buff to buff-gray and locally manton dolostones. This subdivision consists of thin to medium beds of parallel planar-laminated dolostones that alternate with thick laminae to thin beds of buff dolomicroite and gray shale. The uppermost unit of the "Gillespie Lake Group" consists of more than 1,000 meters of buff- to orange-weathering carbonates including crinkly laminated dolostones, oolitic and pisolithic beds, and stromatolitic bioherms and biostromes. Molar-tooth structure is ubiquitous throughout this platform sequence.

In the eastern part of the map area (Figure 6) rocks of the "Gillespie Lake Group" are locally overlain unconformably by unfossiliferous red-weathering shales and argillites of unknown affinity. These rocks may be equivalents of the lower part of the Rapitan Group.

Complicated structural relationships have severely hindered stratigraphic studies in the "Wernecke Supergroup." The most prominent structural system is a series of high angle faults which trend west-northwest and divide the area into a series of discrete blocks (Figure 6). Numerous low angle thrusts, which cut both the "Quartet" and "Gillespie Lake Groups," further complicate the reconstruction of complete stratigraphic sections for these groups. The rocks of the "Wernecke Supergroup" differ markedly from the Proterozoic succession in the neighboring Mackenzie Mountains to the east ("Mackenzie Mountains Supergroup"). A distinct break occurs between these two types of successions at the Bonnet Plume River where Mackenzie-type strata are exposed on its eastern side and Wernecke-type strata on its western side. Part of the explanation for juxtaposition of these contrasted assemblages may lie in Norris's (1977) tectonic model for this region. The model involves large-scale right-handed transcurrent faults so that the entire Wernecke assemblage may have been transported for tens to hundreds of kilometers from the south, along the Richardson fault array which parallels the Bonnet Plume River valley (Norris, 1977). The "Wernecke Supergroup" bears a much closer resemblance to the lower and middle part of the Belt-Putrell succession of the Canada-United States border region. In particular, the lower argillaceous units may be compared with the thick succession of fine-grained clastics in the lower Belt. The platform carbonates are strongly reminiscent of the Middle Belt carbonates: the Wallace and Sisyh Formations and equivalents. This speculation might be tested by detailed comparison of stromatolites from these two widely separated regions.

**STRATIGRAPHIC CORRELATIONS**

The history of earlier work concerned with correlation of the rocks of the Amundsen Embayment (Figure 3) was summarized by Young (1977), who thought that most of the rocks of the Brock Inlier (Figure 4, area 2) and the Rae Group of the Coppermine Homoclinal could be correlated with the Glesnelg Formation and lower part of the Reynolds Point Formation. As an expansion and refinement of the proposal by Aitken and others (1973) of a general equivalence between the Shaler Group rocks of the Brock Inlier and the older Proterozoic rocks ("Mackenzie Mountains Supergroup") of the Mackenzie Mountain fold belt to the southwest, Young (1977) proposed a tentative formational correlation of the stratigraphic units between these two areas. This correlation appears to be well-substantiated up to the lower oolitic units of the Little Dal Group and corresponding rocks of the Reynolds Point Formation of Victoria Island (Figure 2). The recently described, thick, evaporite red bed sequence in the lower part of the Little Dal Formation ("gypsum subunit") of Aitken and others (in press) was formerly reported as a separate unit below the Little Dal (Aitken, 1977). Resolution of the confusion surrounding the position of this evaporite red bed sequence and consequent identification of the great thickness of the Little Dal Formation allows the possibility that these evaporites might correspond to the Minto Inlet Formation of the Shaler Group (as opposed to the Redstone River Formation higher in the sequence, as suggested by Young, 1977). This idea was first put forward by Aitken (personal communication to C. W. Jefferson, 1977). If this correlation is accepted then the Redstone River evaporites might correspond to those of the lower member of the Kilian Formation. The Wynniatt Formation would then be correlatable with the upper part of the Little Dal Formation.

Problems still exist in correlating these upper units, and for this reason no connecting lines have been drawn on Figure 2. The stromatolites have not yet received sufficient study or have the evaporitic sequences been adequately studied and compared. Recent investigation of the Wynniatt-Kilian succession has shown it to be largely made up of rocks of shallow marine aspect. The facies of these rocks contrasts with that of the Coppercap Formation of the "Mackenzie Mountains Supergroup" — the putative correlative according to Young (1977). Even if it is accepted that the Redstone River Formation corresponds to the evaporitic lower Kilian Formation, the upper Kilian rocks are not similar to the deeper water reedimented carbonates of the Coppercap Formation. This could be explained as a facies change between the two areas, with deepening to the west; or perhaps the Coppercap is entirely younger than the Shaler Group.
REGIONAL CONSIDERATIONS

In spite of the problems mentioned above, rocks of the "Mackenzie Mountains Supergroup" in the northern Cordillera are considered to be reliably correlated with the Shaler Group of Victoria Island and areas to the south. These rocks are bracketed between 1.2 b.y. and 6.7 b.y. This correlation is much more firmly based than the formerly proposed correlation of the "Mackenzie Mountains Supergroup" with the Belt Supergroup to the south (Gabrielse, 1972; Donaldson, 1973).

The known age limits of many stratigraphic sequences in the Cordilleran region and in the northwestern part of the Canadian Shield do not lend themselves readily to "pigeonholing" in Stockwell's (1964, 1973) scheme for stratigraphic subdivision of the Precambrian of the Canadian Shield. For example the Belt-Purcell package appears to be Helikan-Hadynian; the Shaler Group also could be Helikan-Hadynian, but the available radiometric age determinations suggest that it is younger than most of the Belt-Purcell sequence. Many of the thick sedimentary accumulations of western North America appear to have formed at times interpreted throughout most of the Canadian Shield as orogenic periods.

Throughout the greater part of western North America there is a clearly recognizable three-fold subdivision of the post-Aphebian (younger than 1.7 b.y.) supracrustal assemblages (Figures 4 and 7). For purposes of discussion in this paper the subdivisions are referred to, in ascending order, as Sequence A, Sequence B, and Sequence C (Figure 7). The first two subdivisions are clearly evident in the northwestern part of the Canadian Shield where a boundary can be placed about 1.2 b.y. ago, after extrusion of the thick sequence of the Coppermine lavas. The importance of this event is indicated by the continentwide distribution at this time of mainly basic igneous rocks.

These include the Mackenzie Dyke Swarm that sweeps in a northwest-southeast direction across the full width of the Canadian Shield (Stockwell and others, 1970), the Muskox intrusion, the Keweenawan lavas, and possibly the slightly younger Purcell lavas at the base of the upper part of the Belt Supergroup. The Cardenas lavas of the Grand Canyon region and basic sills in eastern Arizona can also be assigned to this episode of basic igneous activity. Such an impressive array of widespread basic intrusions and extrusions must signify an important period of crustal tension, at least in the western half of the North American continent. These tensional forces may have contributed to the founding of sedimentary basins in what is now the western part of North America.

SEQUENCE A (1.7 b.y. — 1.3 b.y.)

The oldest post-Aphebian assemblages include those of the classical Belt-Purcell (Figure 4, area 7). Available radiometric age determinations, summarized in Harrison (1972) suggest that this thick pile of sediments was formed between 1.7 b.y. and 0.85 b.y. ago. Evidence of major breaks within the Belt-Purcell package is equivocal. Harrison (1972), however, suggested that perhaps the most important break is that between the middle and upper Belt — at the base of the Missoula Group. This event is one of the few that have been pinned down radiometrically, for a 1.1 b.y. potassium-argon date has been obtained from the Purcell lavas, close to the base of the Missoula Group.

Other Precambrian successions that could be assigned to Sequence A are the pre-Rae Group supracrustal rocks of the Coppermine Homocline (Figure 4, area 3), those of the Thelon and Athabasca Plains (Figure 4, areas 12 and 13), the Unkar Group of the Grand Canyon area (Figure 4, area 9, Figure 8), and the Apache Group and Troy Quarzite of Arizona (Figure 4, area 11).

In some regions such as the Wernecke Mountains (Figure 4, area 3) and the Tuchodi Lakes area (Figure 4, area 6; Bell, 1968), there are no definitive data on the age of the Precambrian sedimentary rocks. The rocks of both these areas appear to be lithologically most similar to those of the lower and middle Belt-Purcell rocks farther south (Gabrielse, 1972), which are here included in Sequence A. Alternatively these rocks could be a basal facies equivalent of the Shaler Group, which is here included in Sequence B. However, their close geographic proximity (i.e. of the "Wernecke Supergroup") to the shelf facies rocks of the "Mackenzie Mountains Supergroup" (Shaler equivalent), their higher grade of metamorphism and degree of deformation, and their highly contrasted lithologic character, all suggest that the "Wernecke Supergroup" is of greater age. Alternatively the "Wernecke Supergroup" could have been tectonically introduced into this region along major dextral transtent faults (Norton, 1977).

In a general way the older successions (Sequence A, between 1.7 and 1.2 b.y.) show a transition from a dominantly continental facies in the east (Figure 4, areas 12, 13) to a marine environment deepening to the west (Fraser and others, 1970; Price, 1964; Donaldson, 1973; Gabrielse, 1972).

SEQUENCE B (1.2 b.y. — 0.8 b.y.)

Rocks of the second great post-Aphebian depositional cycle were deposited unconformably on older se-
quences that bear evidence of the widespread 1.2 b.y. basic igneous event. The most widespread exposure of rocks of this cycle is in the Amundsen Embayment (Figure 4, areas 1, 2, and 3) and its westward continuation into the Mackenzie Mountains region (Figure 4, area 4). These rocks (Shaler Group and correlatives) are dominantly a shallow marine shelf assemblage. They were previously correlated with the oldest post-Aphebian assemblage (Belt-Purcell Supergroup) by Gabrielse (1972) and with the youngest (Windermere Supergroup) by Donaldson (1973), but the available radiometric age determinations and the lithologic characteristics of the rocks support correlation of these northern assemblages (Shaler Group and "Mackenzie Mountains Supergroup") with only the upper part of the Belt-Purcell Supergroup (Missoula Group).

A strikingly similar sequence of events took place in the Grand Canyon region where the oldest post-Aphebian succession (Unkar Group) is capped by the Cardenas Lavas, dated about 1.1 b.y. (Ford and others, 1972; Elsbeke and Scott, 1976). These rocks are overlain discordantly by the Nankoweap Sandstones which are in turn succeeded by the Chuar Group (Ford and Breed, 1973). The Chuar Group may be comparable in age to the Shaler Group in the north. The fossil *Chuaria* (Hoffman, 1976) and chitinoida (Bloeser and others, 1977) have been found in the Chuar Group. *Chuaria* has also been reported from rocks of the Uinta Mountain Group in Utah (Hoffman, 1976). These rocks are

Figure 7. Schematic section to show stratigraphic relationships of some Proterozoic rocks of western North America. Note that major unconformities separate the various sequences of the Proterozoic. Some of the dated igneous rocks are also shown. Note especially the basic igneous episode (at about 1.2 b.y. ago), represented on the figure by the Coppermine lavas and possibly the Purcell lavas. Note the relationship between the 700 m.y.-old Naktusiak lavas and the rocks of Sequence C — they are nowhere seen in contact. Note also that the system as currently known is one-sided, and that the unconformities are most strongly developed in easterly regions. Neither the nature of the substrate on which many of the Proterozoic rocks accumulated nor the configuration of the western parts of the basins are currently known.
considered to be about 950 m.y. old, an age not inconsistent with correlation of these rocks with the Shaler Group. *Chuaat* and chitinozoans have not so far been discovered in the Shaler Group, but J. D. Aiken (personal communication, 1977) has found *Chuaat* and primitive trilobites in the Little Dal Group in the Mackenzie Mountains area. Obtaining more information is needed on the areal distribution and stratigraphic range of these fossil forms.

The undated lower part of the Pahrump Group of Death Valley region (Figure 4, area 10) underlies, with apparent conformity, mixtites and associated rocks of the upper part of the Pahrump Group (Kingston Peak Formation) (Wright and others, 1974); hence, it may be considered part of Sequence B.

SEQUENCE C (0.8(?) b.y.—0.6 b.y.)

Obradovich and Peterman (1968) suggested that the youngest part of the Belt-Purcell Supergroup might be about 800 m.y. old, but the boundary between these rocks and the overlying Windermere is not well-defined radiometrically. In the absence of any more definitive evidence, and following Gabrielse (1972) and Stewart (1972), the boundary between Sequences B and C is taken at 0.8 b.y. ago. Diabases that cut the Shaler Group have yielded dates about 700 m.y. However, the area in which these igneous rocks occur is far removed from the nearest outcrops of Sequence C ("Ekwi Supergroup").

Along the length of the Cordillera the uppermost Proterozoic rocks are separated in most areas by an unconformity from older rocks. Gabrielse (1972), however, reported conformable relations in some western regions of the Cordillera. The younger rocks (Sequence C) in many areas contain thick mixtites (Schermerhorn, 1966) that have been interpreted by some as glacial deposits (Ziegler, 1962; Young, 1976; Eibach, 1977). According to Stewart (1972, 1976), these rocks record evidence of continental separation along the Western margin of North America about 850 m.y. ago. The generalized distribution of the youngest Proterozoic rocks (Sequence C) is shown in Figure 4. It differs from that suggested by Stewart (1976), because at the time of Stewart's publication the Shaler Group was considered by some to be part of Sequence C, rather than B. The stratigraphic succession in the Death Valley region (Figure 4, area 10; Figure 8) is similar to that in the Mackenzie Mountains region (Figure 4, area 4), in that mixtites and associated fine clastics overlies shallow-marine clastics and carbonates with conformable relationships in some areas and unconformable relations in others.

CONCLUSIONS

In western North America, post-Aphelien supracrustal assemblages may be divided into three sequences. The oldest (Sequence A) is between 1.7 b.y. and 1.2 b.y. old. These rocks show a major facies change from being dominantly continental in the east to dominantly marine in the west. Paleocurrent data, facies changes, and isopach maps in some areas support the existence of localized depocenters that took the form of reentrants on the eastern margin of a major depositional basin that must have lain to the west of the present outcrops (Harrison, 1972).

A second episode of restricted sedimentation (Sequence B) was preceded in most areas by a period of continentwide basic igneous activity about 1.2 b.y. ago. The upper limit of Sequence B is poorly defined at 800 m.y. Rocks of Sequence B appear to be present in virtually all regions where those of Sequence A are preserved. Possible exceptions occur in the Arizona region. Sequence B rocks are especially widespread in the northern part of the Cordillera and in the Amundsen Embayment (Figure 3).

The third period of Proterozoic deposition (Sequence C, 0.8(?) b.y. to 0.6 b.y.) is extremely widespread in a north-south direction but appears to have been restricted in an east-west sense. Deposits of this age are not known from areas east of the Cordilleran fold belt. Glaciogenic mixtites are present in some areas. An oceanic source has been proposed for the important iron-formation associated with the glacigenic Rapitan Group (Yeo, 1977; Delaney and others, this volume). Acceptance of this proposal would support the existence of at least a small ocean basin to the west during Rapitan sedimentation (Eibach, 1977). Stewart reported the marked contrast between the sedimentary style of the youngest Proterozoic rocks (Sequence C) and those below. This, together with the common occurrence of basic volcanism near the base of Sequence C, led him to infer continental separation along the length of western North America about 850 m.y. ago.

There are almost as many opinions as there are papers regarding the tectono-sedimentary history of western North America during the Proterozoic. Badham (in preparation) and Hoffman and others (1973) suggested that a continental margin existed in Aphelien (early Proterozoic) time at the site of the Cotonation Geosyncline. Gabrielse (1972) considered the initial breakup to have occurred during the Belt-Purcell sedimentary cycle (at the beginning of deposition of Sequence A). Burke and Dewey (1973) recognized the important 1.2 b.y. igneous event and suggested that continental fragmentation was initiated at the beginning of Se-
quence B deposition. They even postulated that the major depositional basin extending eastward from the Mackenzie Mountains to the western Arctic Archipelago might have been an aulacogen of the early Cordilleran geosyncline. They considered the western 500 kilometers of North America to have been accreted during the last 1,200 m.y.

A summary of the stratigraphic evidence presented in this review is given in Table 2 and Figure 7. Most of the paleocurrent data and gross facies changes support the existence of a large-scale sedimentary basin (geosyncline?) in western North America since early Proterozoic time. There are differences in the basins throughout Proterozoic time (limited north-south distribution of preserved basins of Sequences A and B, as opposed to great North-South extent of Sequence C), but most geologists have invoked larger depositional basins somewhere to the west throughout most of the period under consideration. Whether these basins were floored by oceanic crust at any time during the Proterozoic, or merely represent successive, mainly abortive, attempts at continental fragmentation must await more detailed studies, particularly in western parts of the Cordillera. There is some evidence of westerly source terrains (Eisbacher and Gabrielse, 1977) for the youngest Proterozoic assemblage. Close stratigraphic similarities between the upper Proterozoic successions of western North America and those of Australia and Siberia might indicate closer juxtaposition of these regions during the greater part of Proterozoic time.

ACKNOWLEDGMENTS

Responsibility for various sections of this paper lies as follows: G. M. Young for the description of Precambrian and the Amundsen Basin region, with contributions by C. W. Jefferson and D. G. F. Long; C. W. Jefferson and D. G. F. Long for the "Mackenzie Mountains Supergroup" rocks; G. M. Yeo for the younger rocks of that region; and G. D. Delaney for the Pro-

DEATH VALLEY

GRAND CANYON

Figure 8. Comparative generalized stratigraphic columns for the Proterozoic supracrustals of the Grand Canyon (after Ford and others, 1972) and Death Valley (after Wright and others, 1974) regions. The Unkar Group belongs to Sequence A but the lower part of the Pahump Group is undated. The Kingston Peak Formation and younger Precambrian rocks of Death Valley region are comparable to Sequence C rocks elsewhere. The age of the Chuar Group is not known with certainty but these rocks may correspond to those of Sequence B.
Table 2. Simplified stratigraphic table to show the proposed breakdown of some post-Aphebian (<1.7 b.y. old) rocks of the Cordilleran region. In some areas the sequences are separated by unconformities. A major episode of basic igneous activity at about 1.2 b.y. ago is used in many areas to date the boundary between Sequence A and Sequence B.

<table>
<thead>
<tr>
<th></th>
<th>Northern Cordillera</th>
<th>Central Cordillera</th>
<th>Southern Cordillera</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>SEQUENCE C</strong></td>
<td>“Ekwi Supergroup”</td>
<td>Windermere Supergroup</td>
<td>Kingston Peak Formation, etc.</td>
</tr>
<tr>
<td>(Rapitan Group, etc.)</td>
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<td></td>
<td>Shaler Group</td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>SEQUENCE A</strong></td>
<td>pre-Rae Group of Coppermine area “Wernecke Supergroup”</td>
<td>Lower and Middle Belt Supergroup</td>
<td>Unkar Group Big Cottonwood Formation</td>
</tr>
</tbody>
</table>

Basement rocks of about 1.7 b.y.

terozoic of the Wernecke Mountains region. The more speculative second part of the paper was mostly written by G. M. Young, but the ideas expressed are the fruit of what Paul Hoffman might call “pedagogic cross-pollination” among all the authors. J. D. Aitken critically read the penultimate version of this paper and “weeded” out some of the wilder ideas.

It is a pleasure to acknowledge the many people and institutions that have helped in the work: National Research Council of Canada has provided financial support to G. M. Young for a number of years. Department of Indian Affairs and Northern Development provided essential funds for expensive field work in remote areas of the northern Cordillera. The Northern Research Group at University of Western Ontario also contributed funds. Logistic support was provided in Arctic Canada by officers of the Polar Continental Shelf Project. All of this help was invaluable. Bob Hornal and Bill Padgham of D.I.A.N.D. in Yellowknife helped greatly in various aspects of our work. Chris Lord initiated and supported the work of C. W. Jefferson in the Cordillera. R. T. Bell of the Geological Survey of Canada did likewise for G. D. Delaney’s studies in the Wernecke region. The hospitality of the personnel at various mining camps in the northern Cordillera was also appreciated.

Among the many individuals who contributed by stimulating discussion, J. D. Aitken, J. A. Donaldson, R. T. Bell, Paul Hoffman, F. H. A. Campbell, and W. R. Morris deserve special mention. G. M. Young would like to acknowledge sterling assistance in the field by Eksu Parvisainen, Scott MacLenna, and his wife Maureen. In spite of this help some of the ideas are still sufficiently controversial that full responsibility must ultimately lie with the authors.

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Some Proterozoic Sediment-Hosted Metal Occurrences of the Northeastern Canadian Cordillera

by

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ABSTRACT

At least five types of metal occurrences are in Proterozoic sedimentary rocks of the northeastern Canadian Cordillera. The stratigraphic settings and descriptions of mineralization are presented as potential aids to exploration in similar rocks elsewhere.

Two distinct metalogenic associations are recognized in rocks of the “Wernecke Supergroup” exposed in the Wernecke Mountains of the central Yukon Territory. Uranium, iron, copper, cobalt, and barite are localized in and adjacent to breccias which cut the entire succession; lead and zinc occupy fissures in dolomites of the uppermost “Gillespie Lake Group.” Most breccia complexes are composite in nature, consisting of at least three genetic varieties: fault-zone breccias, stope breccias, and channelway breccias. Breccias are spatially associated with major lineaments. Mineralization is generally associated with one or more phases of alteration which include feldspathization, carbonatization, silicification, and hematitic alteration. The Wernecke breccias may have developed under a dilational tectonic regime.

Lower Little Dal Group carbonates host Mississippi Valley-type zinc mineralization at Gayna River. The major mineralized unit consists largely of dolomitized ooid grainstones. Known mineralization directly overlies a belt of stromatolitic reefs.

Stratabound copper occurs in the Mackenzie Mountains near the top of the pre-Rapitan “Mackenzie Mountains Supergroup,” at the transition between terrigenous red beds and evaporites (Redstone River Formation) and clastic and algal carbonates (Coppercap Formation). Mineralization is thought to have developed syngenetically to diagenetically during a marine transgression.

The Rapitan Group (Hadrynian), the basal sequence of the youngest Proterozoic assemblage in the northern Cordillera, contains an unusual and widespread, banded iron-formation. Abundant dropstones and the association with glaciallastic sediments show that this chemical sediment was deposited under glacial conditions. Three facies of jasper-hematite iron-formation are present: a laminated facies associated with mudstone rhythms, a nodular facies commonly associated with mireites and coarse clastics, and an irregular facies also found with poorly sorted clastics. Textural and chemical evidence suggests that the latter facies were enriched in iron during early diagenesis. The source of the iron is not known, but it might have been derived through hydrothermal leaching by seawater of oceanic crust inferred to have lain west of the present outcrop belt. An alternative source may have been through breccia diatremes which intrude rocks below the Rapitan. Water circulation beneath sea-going glaciers may have provided the overturn necessary to bring about upwelling of the enriched seawater. Precipitation of iron and silica would take place under the influence of the strong physicochemical gradients found near a glacial coast.

INTRODUCTION

Significant metal occurrences have been found in recent times in Proterozoic strata of the northeastern Canadian Cordillera. This preliminary report describes five such occurrences in the Wernecke and western Mackenzie Mountains (Figure 1).

The Wernecke Mountains physiographic area forms the transition from the northwest-trending Mackenzie to the west-trending Ogilvie. Wernecke strata include Proterozoic fine-grained clastics and carbonates and lower Paleozoic carbonates. Metamorphism is lower green schist grade, but primary features are evident.

The Mackenzie Mountains are an arcuate belt of Upper Proterozoic to Upper Paleozoic, epicontinental clastic and carbonate strata with rare volcanic rocks. Metamorphism is subgreen schist grade, and primary
Sedimentary structures and textures are well-preserved. The obvious tectonic structures, the open folds and thrust faults of broadly Laramide age, bear no apparent relationship to the metal occurrences discussed here.

Stratigraphic studies of Proterozoic rocks in the northern Cordillera are only now approaching regional coherence. The strata of the Wernecke and Mackenzie Mountains consist of at least three sequences of supergroup rank. In ascending order these are (informally): the “Wernecke Supergroup,” the “Mackenzie Mountains Supergroup,” and the “Ekwi Supergroup.” The regional stratigraphy is described elsewhere (see Young and others, this publication).

Rocks discussed in this paper are about the same age and general appearance as Belt-Purcell-Windermere rocks of southern British Columbia and northwestern United States, but no reliable radiometric ages have been obtained from them. No stratigraphic contact between the “Wernecke Supergroup” and the “Mackenzie Mountains Supergroup” has been observed, although both are overlain by the “Ekwi Supergroup.” However, the “Wernecke Supergroup” appears to be the oldest and compares in a general way to the Lower Belt, Ravlali Group, and Middle Belt Carbonates. The “Mackenzie Mountains Supergroup” is similar to the upper Belt Missoula Group. The “Ekwi Supergroup” in turn corresponds to the Windermere Supergroup. Details of these speculative comparisons are discussed in Young and others (this publication). Details of stratigraphy pertinent to the individual occurrences are given with the descriptions of those occurrences.

The following descriptions are arranged in ascending stratigraphic order. G. Delaney is responsible for the Wernecke breccia. R. T. Bell initiated the project and provided some field supervision. J. D. Aiken supplied material for the oral presentation on the Gayna River zinc property. C. W. Jefferson compiled the summary of the Gayna River zinc property from the literature and wrote the section on the Redstone copper belt. G. M. Yeo and S. M. McLennan described the Rapitan iron-formation.

The descriptions are presented as potential aids in exploring and understanding similar occurrences in other rocks of this age.

Figure 1. Location maps.
MINERALIZATION IN THE "WERNECKE SUPERGROUP"

GENERAL

Two distinct metallogenic associations are recognized in the lowermost Proterozoic succession exposed in the Werncke Mountains of the central Yukon Territory (Figure 1). Huge discordant breccias which dissect this proterozoic succession have associated uranium, copper, barium, cobalt, and iron enrichments; fracture systems in the upper dolomite sequence of this succession contain local significant enrichments of lead and zinc.

STRATIGRAPHY

The stratigraphy of the Proterozoic rocks, which host these mineral occurrences, is discussed in an accompanying paper by Young and others (this publication); what follows is merely a brief overview. These rocks are subdivided into three units of group status. They have a combined thickness of between 11 and 15 kilometers. The lowest subdivision is informally referred to as the "Fairchild Lake Group" (Delaney, in preparation). It consists of up to 5 kilometers of lenticularly bedded siltsstones with a ribbed weathering carbonate member near the middle. Overlying the "Fairchild Lake Group," with uncertain relationships, is a sequence of 5 to 6 kilometers of dark gray weathering shales, siltstones, and fine-grained sandstones, which constitute a series of coarsening upward cycles. This unit has been informally designated the "Quartet Group" (Delaney, in preparation). The "Quartet Group" is transitional to a 2 to 5 kilometer-thick sequence of orange to buff weathering fine-grained dolostones, the uppermost part of which consists of a platform facies. This unit is informally referred to as the "Gillespie Lake Group" (Delaney, in preparation). The entire succession outlined above has been designated the "Werncke Supergroup" (Delaney, in preparation). Complicated structural relationships, coupled with the dissimilar nature of this Proterozoic succession to that exposed in the Mackenzie Mountains, has hindered attempts at regional correlation (Young and others, this publication).

WERNECKE BRECCIAS

Although the existence of metalliferous breccias in the Werncke Mountains has been known for some time, it was not until 1976 that the extent and importance of these breccias as a metallogenic control were fully appreciated (Bell and Delaney, 1977). Subsequent field work by Bell (in press) and Delaney (in preparation) during the summer of 1977 has added a great deal of information concerning the characteristics and tectono-stratigraphic setting of the breccias.

The size of individual breccia bodies is quite variable. Some of the larger ones near Quartet Lakes (65°60’N., 154°23’W.) and Slab Mountain (65°00’N., 154°00’W.) are up to 4 kilometers by 1 kilometer. A distinct spatial association exists between the breccias and major tectonic lineaments. Most breccias lie along major fault systems, which trend west northwest and north to north northwest, or at their intersections. The west-northwest system of faults, which appears to be the dominant system, consists of high angle normal and reverse faults.

BRECCIA TYPES

Several distinct breccia types have been defined in the Werncke breccia complexes. These types are (1) fault breccia, (2) stope breccia, and (3) channelway breccia. The fault breccia consists mainly of elongate to equant, pebble- to cobble-sized angular fragments in a granulated matrix. This type of breccia is generally contained within clearly defined linear zones, and it is commonly possible to define a distinct closely spaced fracturing in the surrounding country rocks. The stope breccias are characterized by fragments which range in size from in excess of 30 meters to sand and granule material. Most of the blocks composing the stope breccia are highly angular and occur as isolated blocks in a granulated matrix or in a closed framework. Most margins of these breccia complexes are steeply inclined. Individual breccia bodies range from circular to elongate in plan view. Channelway breccias are characterized by several features: (1) They occur in cylindrical to ellipsoidal pipe-like structures generally 1 to 5 meters in diameter; these pipes are commonly aligned at a high angle to the horizontal, (2) Most of the fragments, which compose the breccias, are cobble-sized and surrounded to rounded with a high degree of sphericity, (3) Locally it is possible to define relict flow textures in the matrix. Figure 2 is a hypothetical sketch illustrating the interrelationships of the various types of breccias. Most breccias have a composite origin, resulting from the superimposition of the stope phase on the fault phase; these two in turn being cut by the channelway breccias.

INTERNAL ZONES OF ALTERATION

A series of zones of alteration is superimposed over the breccias; locally apophyses of this alteration extend into the surrounding country rock. The main types of
Figure 2. Schematic diagram showing stratigraphy and mineralization of the "Wernecke Supergroup" and the inter-relationship of the genetic phases in an idealized Wernecke breccia complex.
alteration are feldspathization, silicification, hematitic alteration, and carbonization. Feldspathic metasomatites, silicification, and quartz and quartz-feldspar veins are most commonly developed in breccias that cut the "Fairchild Lake Group" and the lower parts of the "Quartz Group." In some exposures, the light-colored feldspathic metasomatites assume an igneous appearance characterized by coarsely crystalline masses of albite (Laznicka, 1977). Carbonate alteration in the Wernecke breccias is characterized by a massive phase consisting of siderite, calcite, and dolomite and by a disseminated phase consisting of coarse-grained subhedral to euhedral rhombs of ankerite scattered throughout the matrix and framework. Hematitic alteration imparts a pinkish to blood red or dark gray coloration to both the framework and matrix. This type of alteration occurs as one of three distinct phases: (1) medium- to coarse-grained flakes of specularite, (2) fine-grained disseminated specularite, and (3) submicroscopic hematite.

Correlating the varieties of alteration discussed above with stratigraphic position reveals a crude zonation. This zonation is defined by the gradual upward loss of plagioclase metasomatites, quartz, quartz-feldspar veining, and related phases (lower "Quartz Group"), followed by the absence of carbonate alteration (upper "Quartz Group") and finally by the loss of any observable effects of hematitic alteration (middle "Gillespie Lake Group").

METAMORPHIC ALTERATION

At lower stratigraphic levels, generally corresponding with the zone of feldspathization and silicification within the breccias, a distinct metamorphic aureole is apparent in the surrounding country rocks. In these aureoles, which may be as much as 300 meters wide, the normally dark gray to gray weathering fine-grained sediments have been transformed into light green chloritic schists and phyllites. At some localities, knots of medium- to coarse-grained chloritoid are present in these rocks, and in at least one locality the presence of almandine garnet has been documented (Delaney, in preparation).

TYPES OF MINERAL DEPOSITS

As mentioned at the beginning of the review, uranium, copper, barium, cobalt, and iron occurrences are associated with the Wernecke breccias (Gross, 1963a; Green, 1972; Bell and Delaney, 1977; Archer and others, 1977). These mineral occurrences are found in irregular patchy bodies within the breccias or in fissure zones that appear to have emanated from the breccias. Most phases exhibit a distinct spatial association with a specific alteration type. Chalcopyrite and bornite together with associated pyrite and pyrrhotite are generally associated with the massive phase of carbonate alteration. The economically interesting occurrences are present at lower stratigraphic levels. Azurite and malachite stain both the breccias and surrounding country rocks throughout the "Wernecke Super-group." The most significant cobalt showings are contained within fault-gouge breccias, which commonly have a proximal association with carbonatization. Barite occurs as irregular pods and dikes, some of which contain medium- to coarse-grained octahedra of magnetite. The zones of hematization, in particular those that exhibit a marked iron enrichment, have been investigated for the potential of mining this commodity (Gross, 1963a; Green, 1972). Uranium generally occurs in the form of uranium titanates localized in quartz-feldspar veins; pitchblende coatings on fracture surfaces have been identified at two localities (Delaney, in preparation).

GENESIS OF THE BRECCIAS

On the basis of the observations summarized above, the Wernecke breccias probably developed under a dilational tectonic regime (Sawkins, 1976; Gabelman, 1977). The dominant fault systems, defined above, probably tapped fluids from the lower crust or perhaps the upper mantle. As these fluids ascended up the fracture systems, they caused a hydraulic stopping (Gilmore, 1976) and at the same time emplaced mineralized solutions. Later, gas-charged fluids broke through the stope breccia complexes in thin pipes that apparently vented. Although for the most part the mineral occurrences were probably derived from ascending fluids, some may contain a component resulting from remobilization of anomalous metal concentrations in the surrounding country rocks. Carbonaceous shales in the lower part of the "Quartz Group" contain anomalous concentrations of uranium and copper (Bell, in press).

Breccias that cut the Proterozoic "Wernecke Super-group" not only have provided channelways for mineralizing solutions but have also served as loci for the deposition of metals. The spatial association of the breccias with major lineaments suggests that the development of the latter is a tectonic precursor to the formation of the breccias. Recognizing the metallogenetic significance of the breccias has documented a new type of ore system that can be prospected for in similar rocks not only in the northeastern Cordillera but also throughout the world.
LEAD-ZINC MINERALIZATION IN THE
"GILLESPIE LAKE GROUP"

At several localities in the Wenecke Mountains, occurrences of lead-zinc have been documented in the "Gillespie Lake Group" strata (Green, 1972; Delaney, in preparation). Most of these showings occur in fracture systems with sphalerite and galena as the main ore minerals. To date, no significant tonnages have been outlined at any of these occurrences.

GAYNA RIVER ZINC MINERALIZATION

The following description is after Aitken (1977) and Newton (1977). The approximate location of the property, between the headwaters of Gayna and Arctic Red Rivers, is shown in Figure 1.

STRATIGRAPHIC CONTEXT

Occurrences of zinc and lead sulfides are widespread in carbonate strata of the total sedimentary column of the region. The mineralization of economic interest at Gayna River occurs within the lower part of the Little Dal Group of "Mackenzie Mountains Supergroup" (Young and others, this publication). The lower part of the Little Dal in the area consists of a "basinal sequence" (Aitken, 1977; Aitken and others, in press) overlain successively by a "grainstone subunit" and a thick "gypsum subunit." The "basinal sequence" consists mainly of limestone-shale rhythmites of deepwater origin and contains, west of Mountain River, a belt of giant reefs of stromatolitic limestone and dolomite up to 300 meters high. The "grainstone subunit" consists of two members: a lower one of mainly dolomitized ooloid grainstone; and an upper one of mainly platy, mud-cracked, sandy, supratidal dolomite. The most promising showings of sphalerite occur within the ooid grainstones overlying the reef belt. Perhaps significantly, small zinc showings are common within the underlying reefs.

MINERALIZATION

The following description is quoted from Newton (1977):

- Two types of mineralization within the dolomitic dolostones near the bioherms are of interest: (1) Brightly coloured sphalerite is associated with white sparry, arborescent matrix in secondary mosaics, rubble and crackle breccias related to solution and collapse, reworking and textural features. "Snow-on-the-roof" texture and complete enveloping of fragments by sphalerite are frequently present. (2) Pale to colourless sphalerite of undetermined origin is disseminated in fragments (some fragments may be 80% sphalerite) and in the matrix of a sedimentary breccia probably formed by slumping off reef margins. Rarely, galena envelopes fragments within the sedimentary breccia.

- The geological environment, mineralogy, and textural features allow for mineralization by circulating connate or meteoric waters.

STRATIGRAPHIC CONTEXT OF COPPER
MINERALIZATION IN THE REDSTONE
RIVER AND COPPERCAP FORMATIONS,
MACKENZIE MOUNTAINS

GENERAL

The transition zone between the Redstone River and Coppercap Formations is the site of copper mineralization along a discontinuous outcrop zone that extends more than 250 kilometers along the Plateau fault zone in the Mackenzie Mountains (Figure 1). The deposit at Coates Lake (Figure 3) was discovered by the Nahanni Sixty Syndicate in 1962. Coates (1964) studied the deposit in detail for Redstone Mines Ltd. Shell Canada optioned the property and staked much of the belt to the north, especially in the Keele River area. Other companies with major interests in the area include Rio Tinto and Canadian Nickel Company. The stratigraphy and tectonic framework of the general area have been established by the following studies: Gabrielse and others (1973), Aitken and others (1973), Aitken and Cook (1974). Aitken (1977), Eichberger (1976, 1977), and Young and others (this publication).

REGIONAL STRATIGRAPHY AND
PALEO GEOGRAPHY

The key stratigraphic units in this study are the Upper Little Dal Group and the Redstone River and Coppercap Formations. These compose the uppermost part of the informal "Mackenzie Mountains Supergroup" described by Young and others (this publication) and are overlain by the Rapitan Group, which forms the base of the "Ekwi Supergroup" (Young and others). The outcrop pattern of these rocks in the Northwest Territories is shown in Figure 3. The Redstone River and Coppercap Formations are absent in huchared areas and thicken markedly toward the center of the other areas. Paleocurrent directions generally indicate transport to the southwest and towards the centers of the outcrop areas of Redstone River Formation. Hence, the pattern of dots indicates the approximate shape of the shoreline during Coppercap time, and the huchared areas represent land areas.
Figure 3. Location and paleogeography of the Redstone Copper Belt. Copper mineralization occurs at almost every outcrop of the Redstone River-Coppercap contact within the "embayments." Paleogeographic interpretation is discussed in the text.
The axes of the "embayments" appear to have trended west to southwest. However, faulting along north-south axes as sketched theoretically by Eibacher (1977) could well have influenced "embayment" development.

Figure 4 is a composite sketch showing stratigraphy and facies changes of a typical wedge of Redstone River and Coppercap as observed in discontinuous outcrop. Faults are not shown because none were observed in the sections where the Redstone River Formation is present. The sketch represents cross-sections at about 45 degrees to the axes of the "embayments," parallel to the line of outcrop.

STRATIGRAPHY OF THE UPPERMOST LITTLE DAL GROUP

Redstone River "embayments" were developed on laterally continuous shallow marine carbonates of the upper Little Dal Group. Near the end of Little Dal time, the beginning of the "embayments" is suggested by thickness and facies changes of sandstone and shale members, stromatolite biostromes, and gypsumiferous limestones. These changes parallel those of the overlying Redstone River and Coppercap Formations.

REDSTONE RIVER FORMATION

The lower boundary of the Redstone River Formation is defined in basinal sections at the base of up to 200 meters of bedded gypsum and anhydrite. In marginal sections, where evaporites are very thin or absent, the base of the formation is defined by the appearance of a sequence of continuous clastics. Grain size decreases and thickness increases markedly toward basinward sections of these clastics, where at least 600 meters of cyclically bedded red beds overlie the evaporites. Throughout most of the formation, small-scale cycles form a continuous sequence of 1 to 5 centimeter-thick, cross-laminated to planar-laminated, graded beds.

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Figure 4. Composite diagrammatic sketch showing stratigraphy and facies changes in the uppermost Little Dal Group; Redstone River and Coppercap Formations. Stratabound copper mineralization occurs in the cryptalgal, fenestral carbonates at the base of the Coppercap Formation.
COPPERCAP FORMATION

The lower boundary of the Coppercap Formation is defined at the base of the thick sequence of carbonates overlying the Redstone River terrigenous clastics and is within the Cuprifereous Zone. The formation is characterized by orange to dark gray weathering detrital limestones and dolostones with laminated and massive graded bedding and foetid shaly limestone interbeds. It is 200 to 300 meters thick in complete sections and comprises three major subunits.

The basal subunit consists of sandy, shaly, and algal-laminated limestones, which commonly contain copper sulfides. These are overlain by up to 30 meters of rhythmically graded and bedded, orange-buff dolomites with shaly interbeds.

The second subunit consists of 10 to over 70 meters of recessively weathering, black, foetid, sulfide-bearing calcareous shales and grey, shaly limestones. Rare graded beds and irregular, nonpolygonal (?) shrinkage cracks are present. The base and top of this unit are gradational.

Rhythmically bedded gray calcarenites to calcilutites with shale interbeds compose most of the third subunit. Sedimentary structures include graded massive to laminated beds, slump balls, flame structures, scour at the bases of beds, flat chip intraformational breccia, and rare ripple drift cross laminae, generally as part of a Bouma cycle. The calcareous shale interbeds have streaky, crenulated, and flat laminae and are locally dark red. Another common rock type is finely planar laminated, partly silicified calcilutites with distinctly ribbed weathered surfaces. The top of the Coppercap is characterized by a variety of unique rock types, including: coarse allogenic limestone breccias; bioturbations of poorly preserved undulatory algal stromatolites (also present lower in the Coppercap); very coarse, silicified breccias; and mottled black, gray, coarse crystalline limestone.

SIGNIFICANCE OF THE COPPERCAP-RAPITAN CONTACT

The Coppercap Formation is unconformably overlain by the Rapitan Group along most of the belt. This is the major break which marks the top of the "Mackenzie Mountains Supergroup" and the base of the "Ekwi Supergroup." Conformable relations are, however, suggested in several areas by interfingering to gradational contacts and by lateral continuity of distinctive marker units at the very top of the formation.

The nature of the Coppercap Formation and contact relationships with the Rapitan are important prospecting tools for at least two reasons. First, the "gypsum subunit" of the Little Dal (see Young and others, this publication) has been several times mistaken for the Redstone River Formation, because both are red-bed evaporite sequences. However, the allogenic limestones of the Coppercap are apparently unique to the uppermost part of the "Mackenzie Mountains Supergroup" and are the best way to confirm the identity of the underlying Redstone River Formation. Second, copper mineralization appears to be best developed where the Redstone-Coppercap sequence is thickest. Conformable relations with the Rapitan in the thickest sections suggest the following: (1) factors causing the Redstone "embayments" were active throughout Coppercap time; (2) these factors influenced the thickness of the Coppercap Formation; and (3) the thickness of the Coppercap is more or less proportional to that of the Redstone River, not governed by sub-Rapitan erosion.

The Coppercap is a dominant cliff-former, in contrast to the Redstone River Formation; thus it is, in some ways, more useful for stratigraphic prospecting.

Jefferson and Young (1977) proposed that the Wynniatt Formation on Victoria Island (see Young and others, this publication) corresponds to the Coppercap and that the base of the Wynniatt was promising for copper mineralization. Contrary to expectations, the Wynniatt Formation is characterized by molar-tooth structure, ooliths, and stromatolites; hence, it possibly corresponds to the "upper carbonate subunit" of the Little Dal Group and is a poor candidate for Coppercap-type mineralization.

CUPRIFEROUS ZONE

The Cuprifereous Zone is defined as the zone of mineralization along the contact between the Redstone River and Coppercap Formations. This contact is intercalated to gradational, and copper mineralization occurs along the entire length of exposure in the study area. The contact records shoreline conditions during abrupt to gradual transgression of marine conditions over terrestrial red-bed environments. Contacts that represent shoreline conditions at lower levels within the "Mackenzie Mountains Supergroup" are also locally cuprifereous, though not to the extent of the Cuprifereous Zone. Included in the Zone are red beds, bedded evaporites, breccias, and carbonate beds at the top of the Redstone River Formation, and carbonates at the base of the Coppercap Formation.

The Cuprifereous Zone in the Coates Lake "embayment" was first described by Coates (1964). Webster and Mustard (1973) and Kirkham (1974) have provided additional comments. The zone is up to 110 meters thick and laterally continuous for at least 8 kilometers, with a possible original extent of over 30 kilometers. It
consists of up to seven, fining-upward, maroon, fine sandstone to mudstone cycles, each clastic cycle being separated by a cupriferous limestone. The top of the Cupriferous Zone is marked by a unit of white, calcareous, pyritic, trough cross-bedded sandstone that pinches out to the north and west.

According to Coates (1964), six limestone beds can be consistently recognized along a lateral distance of several kilometers, although individually they may swell or pinch out entirely. The gray limestone weathers gray to brown with locally conspicuous malachite and azurite staining. Primary textures include flat to crinkly and pustular cryptalgal laminae, gypsum crystal casts, and millimeter-sized pods that may have been fenestrae or gypsum nodules. Very recessive, white to pale buff, possibly gypsumiferous mudstones commonly underlie and overlie the limestones. These are in turn bounded by bleached, buff to pink, calcareous and dolomitic mudstones. The zone of bleaching ranges in thickness from less than 10 centimeters to more than 4 meters.

Copper mineralization occurs within the carbonate beds and in bleached clastics immediately above and below the carbonates. No copper mineralization was observed within the red beds. The primary metallic phases, in decreasing order of abundance, are pyrite, chalcopyrite, bornite, digenite, chalcocite, covellite, tennantite, and galena. Also present are supergene malachite, azurite, native copper, and iron oxides (Coates, 1964). There is an antipathetic and vertically controlled relationship between pyrite and copper minerals. The lowest two mineralized beds are the richest, with abundant bornite, digenite, chalcocite, and chalcopyrite but no pyrite. Overlying mineralized horizons contain only chalcopyrite and pyrite.

The results of drilling by Shell Canada, as part of its option agreement with Redstone Resources, were published in the Northern Miner on May 20, June 17, and July 1, 1976, and June 30, 1977. Grades of 3.0 to 6.2 percent copper and 0.3 to 0.6 ounces silver, over thicknesses of 3 to 5 feet (0.9 to 1.5 m) were reported over an area of about 19,000 by 5,000 feet (5,800 by 1,500 m).

Throughout the copper belt, the copper sulfides occupy primary or early diageneric pore spaces in the carbonates or bleached red beds. The pores include possible fenestrae or casts of gypsum nodules, sedimentary fractures, spaces between clastic grains of carbonate and terrigenous material, and dissolved gypsum prisms. The possible fenestrae or gypsum nodules are circular to oval in cross-section and range in diameter from 1 to over 20 millimeters. They are filled with dolomite, calcite, or quartz, rimmed with sulfides, or totally filled with sulfides. In samples where fine calcilutite layers are pulled apart and intruded by sedimentary calcarenite dykelets, the coarser grained dykelets are preferentially mineralized. However, finely disseminated sulfides in calcilutes and siltites are also common.

In breccia zones the breccia fragments themselves are mineralized instead of the large spaces between fragments. Some delicately mineralized carbonate layers have been penecontemporaneously faulted, with smearing out of sulfides along fault planes. Horizontal and vertical fractures are commonly filled with pure sulfides in strongly mineralized localities.

DISCUSSION

Desiccation cracks, algal laminae, and ripple marks, in sediments that are transitional between gyspiferous red beds and foetid detrital limestones, imply sabkha-like conditions during a marine transgression. However, the abundance of water required to supply the inferred alluvial systems does not fit a sabkha environment. Perhaps a better climate would be monsoonal, with clastics deposited during a wet season and evaporites developed during a dry season. This model is analogous to one suggested by Schmalz (1969), who noted that present-day thick evaporite accumulations develop in monsoonal climates.

The Cupriferous Zone is generally diachronous, although it may have occurred simultaneously across some areas. At two sections in particular, the Coppercap limestones clearly lap onto the Redstone River Formation. Distinctive beds within the Coppercap can be traced to where they abut the Cupriferous Zone, and the onlap relationship is visible in distant views of the cliffs. The intercalated cupriferous limestones and red clastics at Coates Lake are interpreted as the deposits (carbonates) of a gradual marine transgression that were repeatedly smothered by distal clastics of a gently sloping alluvial plain. The cyclicity of this sequence suggests relative stability and a long period of marine-terrestrial interaction. The lack of pure evaporites (a few salt casts and gypsum prisms are present) is an indication of continuous access to marine circulation and lack of ponding.

Although no carbonate material was observed in the Cupriferous Zone, the abundance of cryptalgal structures and the green reduction zone in red beds surrounding mineralization suggest that organic material played a role in sulfide precipitation. Bacterial reduction of sulfides is also a possibility that is supported by the presence of sulfides rimming gypsum nodules, in accordance with the model described by Annels (1974).

Primary textures suggest that mineralization occurred during diagenesis and that porosity was a key factor. Fine calcilutite, for example, would have an extremely high porosity when first deposited, but would lithify and rapidly undergo a decrease in porosity and perme-
ability. Some mineralization took place after sedimentary deformation, as in the case of sedimentary dykelets and some faults. However, mineralization appears to have been complete before some other sedimentary faults and brecciation had occurred. Chalcopyrite casts after pyrrhotite are good evidence of diagenetic mineralization.

If hydrothermal mineralization had been the primary agent, it is unlikely to have preserved delicate sedimentary structures and fine primary textures. Epigenetic mineralization in the Little Dal Group and Coppercap Formation is mostly associated with brittle fracturing and coarse crystalline carbonate veins with massive chalcopyrite layers up to 1 m or more centimeters thick. Other showings are cross-cut by late sulfide veins.

The important copper-controlling units in the "Mackenzie Mountains Supergroup" are the uppermost Little Dal Group shelf carbonates, the Redstone River Formation red beds and evaporites, and Coppercap Formation alloplastic limestones. These were deposited along a northwesterly shoreline in which easterly to northeastward emplacements began their development during the latest Little Dal time. Stratiform, syngenetic to diagenetic copper mineralization developed during the marine transgression of alloplastic limestones over red beds, under evaporitic conditions.

THE RAPITAN IRON-FORMATION

STRATIGRAPHIC SETTING

The Rapitan iron-formation crops out discontinuously in a north- and northwest-trending belt stretching some 600 kilometers through the Mackenzie Mountains of the Yukon and Northwest Territores (Figure 1). The iron-formation occurs in "Thundercloud Subgroup," the lower, glaciogenic part of the Rapitan Group. The regional stratigraphic context and detailed stratigraphy of these rocks is summarized elsewhere in this publication (Young and others). The "Thundercloud Subgroup" comprises an upper, gray, essentially barren, mixture-dominated (Schmerhorn, 1966) unit, the "Shezal Formation," and a lower, maroon mixture or rhythmite-dominated unit, the "Sayneui Formation" in the southeast, and the "Snake River Formation" in the northwest. The "Snake River Formation" consists of stratified and massive mixture with orthogonolomite and sandstone. The stratigraphically equivalent "Sayneui Formation" contains three members: (1) the lowermost green mixtures and mudstones of the "Mt. Berg member;" (2) the maroon rhythms and conglomerates of the "Luks Creek member," which is the most extensive, and (3) the maroon mixtures of the "Mountain River member." The boundary between the "Sayneui-Snake River Formations" and the "Shezal Formation" is marked by a color change from red to gray. This change is a primary feature possibly reflecting a change in depositional environment from oxidizing to reducing conditions or a cutoff in hematite supply. Detrital magnetite, probably produced by the Little Dal Formation basaltic, occurs in the clastic sediments below and above this color boundary. Finely disseminated pyrite, which produces the reddish color, is found only below the contact, whereas authigenic pyrite is common only above it in the "Shezal."

THE IRON-FORMATION

According to Stuart (1963), the dominant mineral of the iron-formation is jasper and hematite with minor carbonates, apatite, chlorite, and greenalite (?) along with a small clastic component. Jasper and hematite appear to be the primary minerals; there is no evidence of carbonate replacement (Dimroth, 1977a, 1977b), although carbonate-rich beds occur in associated clastic sediments. Casts of glauberite (?) and dolomite (?) in the iron-formation (Young, 1976) suggest that seawater was supersaturated at or shortly after the time of deposition.

Three textural facies of iron-formation are recognized. Laminated iron-formation typically occurs above thin-bedded rhythms and is commonly persistent laterally for hundreds of meters or more. It commonly appears to be transitional from "A to E" (Walker, 1967) rhythms in which the amount of jasper and hematite in the mudstone ("E") units increases upwards as in the pelagic iron-formations discussed by Dimroth (1977a) and others (Dunbar and McCall, 1971; Sieg&lscaron;ski and Scott, 1974; Shegel'ski, 1975). Nodular iron-formation is composed of elliptical or disc-shaped jasper nodules less than 1 centimeter in diameter with a hematitic matrix which is typically faintly laminated. This facies is more commonly associated with mixtures or coarse clastics in lenses of less lateral persistence than the laminated facies. Stuart (1965) recognized a third facies, characterized by irregular and intergrown masses of hematite and jasper. This is also associated with unsorted clastics. All three facies intergrade. The nodular texture and irregular intergrowth are interpreted as features of early diagenetic alteration of laminated iron-formation. Resedimented fragments of iron-formation in the associated mixtures sometimes show evidence of replacement of silica by hematite, supporting the concept of very early iron enrichment.
Figure 5. Graphic representation of major element variation in sections at Discovery Creek ("Snake River Formation") and near North Redstone River ("Sayunei Formation"). The iron-formation at Discovery Creek is predominantly nodular and the sediments (stippled pattern) are maroon mixites. The iron-formation at North Redstone River is predominantly laminated, with gray mixites ("Shezal Formation") above and maroon rhythmites ("Lukas Creek Member") below. Note the relatively greater iron content of the nodular iron-formation from Discovery Creek, which is diagenetically more altered.
The major concentration of iron-formation is in the "Snake River Formation" in the northwest where considerable exploratory work was done in the early 1960's (Stuart, 1965; Dahlstrom, 1972). Stuart (1965) estimated that reserves here may amount to over 20 billion tons of iron-formation, of which more than 8 billion tons averaging 47.2 percent iron could be mined by open pit methods within a 10 square mile area. Although iron-formation is common in the "Sayunei Formation," where it occurs at or near the transition between the "Luks Creek" and "Mountain River" members, the iron-rich beds are thinner and fewer in number. In the "Snake River Formation," nodular and irregular iron-formation are the dominant textural facies. Centimeter-scale banding has been traced for about 1.5 kilometers, and iron-formation zones can be traced much further. Dropstones, common within the iron-formation of both "Snake River" and "Sayunei Formations," indicate that the iron-formation was deposited under proglacial conditions. Clastic sills and dikes cut laminae and other structures and textures, suggesting that these features are primary or early diagenetic.

Major element analyses of the iron-formation show that it is markedly depleted in everything but ferric iron, silica, and phosphorous compared to the associated clastic sediments (Yeo, 1977) (Figure 5). Amounts of ferrous iron are negligible in the iron-formation, although detrital magnetite is locally common in the associated clastics. The iron content of the nodular iron-formation is considerably higher than that of the laminated iron-formation. Rare-earth element (REE) studies by Fryer (1977), showing that the nodular iron-formation is relatively enriched in the more soluble heavy REEs, suggest that it has undergone diagenetic enrichment as well.

OTHER LATE PROTEROZOIC GLACIGENE IRON DEPOSITS

Glacial deposits of late Proterozoic age have been described from many parts of the world. As pointed out by Young (1976) and Whitten (1970), iron-formation strata are associated with several of them (Figure 6). Many of these glacial deposits, including all of those with iron-formation, are at or near the edge of what were stable cratons by late Proterozoic time. The similarities among these deposits suggest that a new classification of iron-formation can be recognized to emphasize their unique features (Dahlstrom, pers. comn; 1976; Yeo, 1977). The best known of these are the Braemar-Holowilena iron-formations of the Yudnamutana Subgroup in South Australia (Dalgarno and Johnson, 1965; Parkin, 1969; Whitten, 1970) and the Banda Alta iron-formation of the Jacuído Group on the Brazil-Bolivia border (Dorr, 1945; 1973; Barbosa, 1976). The Jacuído Group is especially notable as it contains one of the largest iron deposits in the world (20 billion tons).

The iron-formation here is petrographically much like the Rapitan in that both nodular and laminated textures occur (Dorr, 1945). The Yudnamutana Subgroup is structurally and stratigraphically similar to its stratigraphic and tectonic setting and that of the Rapitan. In fact, the stratigraphy and metallogeny of the Adelaide Geosyncline show a striking resemblance to those of the northern Canadian Cordillera. In both cases, iron-bearing glacial sediments overlie thick platform sequences and are overlain by shaly units. Deposition of the Braemar-Holowilena was partly controlled by Surtian tectonism with faulting occurring prior to and during (?) glaciation (Dalgarno and Johnson, 1965; Parkin, 1969). Intrusive breccias in pre-Surtian sediments, reminiscent of the pre-Rapitan breccias of the Mackenzie and Wrence Mountains, have both copper and uranium mineralization (Parkin, 1969).

A MARINE CHEMICAL MODEL FOR DEPOSITION OF THE RAPITAN IRON-FORMATION

A definitive model for the Rapitan iron-formation and similar deposits throughout the world is not yet possible. The genesis of late Proterozoic iron-formations has been a problem in attempts to explain the origin of iron-formations by biologic and atmospheric evolution (e.g., Cloud, 1973). Condron (1964) interpreted the Rapitan iron-formation as the product of terrestrial weathering and subsequent enrichment by leaching and oxidation. Since the "Thundercloud Subgroup" is a glaciogene deposit, it seems unlikely that a terrestrial weathering model is applicable. Furthermore, such a model cannot explain the sudden change in primary coloration at the base of the "Shezal Formation," because the color indicates that depositional conditions changed from oxidizing to reducing.

Gross (1965b; 1973) considered the Rapitan to have affinities with Algoma-type iron-formations on the basis of its tectonic and stratigraphic setting and association with "tuffaceous" sediments. Gross (1965a) proposed that dissolved hematite and silica were carried by fumarolic waters along fault zones and precipitated on the sea floor. He noted that hematite-rich breccia pipes (some of which include locally abundant classes of banded iron-formation), which are common in the Wrence Mountains (Bell and Delaney, 1977), may represent these fumarolic channels. "Tuffaceous" (Gross, 1965a) hematitic arenites from the "Snake River Formation" actually consist of rounded and well-rounded (partly recycled) elastic grains mixed with plastically deformed
jasper in sand-sized blobs. The latter do appear shard-like. Volcanic clasts are probably derived from the underlying Little Dal Formation. The breccias themselves clearly predate the Raptitan. No instance of the brecciation which should occur in the Raptitan, at least below the iron-formation, has been observed.

Young (1976) inferred that the iron-formation is a sort of evaporitic deposit in which the concentration of dissolved or colloidal iron could be achieved by freezing large quantities of water. Such a mechanism has also been suggested for the Jacoligo iron deposit (Dorr, 1973). This model provided a mechanism for concentrating the iron but did not attempt to explain how the iron was dissolved or suspended, or what its source might be. A fourth model is presented below.

A number of characteristics, which, in combination, seem unique to the Raptitan iron-formation and other deposits in this class, are useful for interpretation. These include:

1. deposition as a continental margin in association with a rifting environment displaying tensional or quiescent features (Stewart, 1972; 1976; Sawkins, 1976a; 1976b; Eibach, 1977).
2. intimate association with glaciogene sediments (Young, 1976; Yeo and Young, 1976; Yeo, 1977).
3. original mineralogy dominated by chert (usually jasper) and hematite (Dorr, 1945; 1973; Gross, 1965a; 1965b; 1973; Gottin, personal communication 1976; Yeo, 1977).

Figures 7 and 8 outline a model that tentatively explains many of the features seen in the Raptitan iron-formation. With modifications, it may be useful in the interpretation of other late Precambrian iron-formations.

Stewart (1972) suggested that rifting along the western margin of the North American craton began between 570 and 830 m.y. ago. Sawkins (1976a; 1976b) proposed that rifting may have taken place as early as 1.200 m.y. ago. Active rifting may have been short lived, resulting in a relatively restricted basin, but it may be assumed that ocean crust was being formed at a mid-ocean ridge and that normal, related activity (hydrothermal circulation, deposition of metalliferous sediments, etc.; see Bostrom, 1970; Corliss, 1971) was taking place.

A glacial-marine origin for the lower part of the Raptitan ("Thundercloud Subgroup") is most likely (Ziegler, 1959; Young, 1976; Yeo and Young, 1967; Eibach, 1976; 1977; Yeo, 1977; Young and others, this publication). During periods of glaciation, cold meltwater and cooled seawater would move offshore while warmer seawater from the inferred ocean ridge to the west moved inshore to replace it (due to temperature and salinity differences). Hydrothermally generated, colloidal, iron-manganese hydroxy-hydrates and

![Diagram of Late Proterozoic glacial deposits of the world. Glacial deposits with appreciable iron deposits are shown as dark squares. Stable cratonic areas (probable continental areas) prior to the Pan African orogeny (≈ 600 m.y.) are shown in stippled pattern. The Pangean configuration of the continents as shown (after Hughes, 1975) is probably not accurate but is presented as a first approximation to the actual interrelationship of cratonic areas in the late Proterozoic.](image-url)
silica, which precipitate in vast quantities at ocean ridges as a result of basalt-seawater interaction (Biscoff and Dickson, 1973; Wolery and Sleep, 1976; Seyfried and Biscoff, 1977), could be carried towards the ice margin in such a circulatory system. Colloidal iron (and manganese) hydroxides could be held in metastable suspension for relatively long periods of time (Seyfried and Biscoff, 1977) and ultimately deposited in the floating shelf or iceberg zone of the glacier (Carey and Ahmad, 1960). Deposition at this location might be enhanced by higher pH and lower temperature and altered salinity (?). Chert deposition may have been controlled by seasonal or other episodic temperature changes (Krauskopf, 1959). Silica may have originated from an oceanic ridge hydrothermal system. Another possibility is that silica may have been present at much higher concentrations in normal seawater during the Precambrian. Present-day seawater is vastly undersaturated in silica (0.1 to 10 ppm according to Krauskopf, 1959), probably due to the presence of silica-secreting organisms. During the Precambrian, in the virtual absence of such organisms, silica may have reached near-saturation levels; perhaps as high as 100 to 140 ppm. This could provide a convenient source for the large volumes of silica in iron-formations and perhaps also for the ubiquitously silicified carbonates of this time. Decreasing pressures and temperatures would enhance supersaturation of silica in upwelling waters (Holland, 1973).

Manganese, accompanying the iron, would be expected to remain in solution due to its higher solubility (Krauskopf, 1957; Hem, 1972) and relatively slow oxidation below a pH of 9 (Stumm and Morgan, 1970). Its fate in the case of the Rapitan is unknown as no associated manganese deposits have yet been found. The Jacadigo Group in South America does have a significant manganese deposit associated with its iron-formation (Dorr, 1945; 1973).

The attractions of this model are numerous. It provides a logical and well-documented source for the vast quantities of iron and silica as well as a simple transport mechanism. The source is consistent with the dilutational tectonic regime thought to be in existence at this time. Concentration of iron and silica, a problem with many iron-formation models, is achieved before transport, making deposition a simple inverse function of terrigenous influx. For example, extensive deposits of finely laminated iron-formation would be expected where terrigenous input is minimal and episodic, whereas lenticular deposits of iron-formation and hematitic terrigenous material would be expected in zones of maximum terrigenous supply (proximal glacial marine facies). This is observed in the Rapitan. Banding could be related to seasonal fluctuations affecting the saturation levels of silica or iron.

One drawback to this model is the absence of iron-formation associated with the late Proterozoic glacial

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**Figure 7.** In the model presented here, iron and silica precipitate from metal-enriched seawater from a nearby spreading center upon mixing with cold, relatively alkaline, fresh water in the vicinity of a sea-going ice sheet. Finely laminated iron-formation is precipitated in turbidite deposits distal to the ice margin and sediment supply, while more massive, lenticular, nodular iron-formation is formed in the zone of maximum sediment supply near the ice margin. Iron-formation may also precipitate beneath the ice shelf. The circulatory pattern involved (after Weyl, 1970) is undoubtedly more complex in reality.
deposits found further south along the western margin of North America. The simplest explanation could be the proximity of the continental margin to the ocean ridge. A second drawback is the necessary assumption that oceanic crust actually was being formed to the west. An alternative source of iron- and silica-enriched water might be the brecciated fracture systems such as those common in the older Proterozoic strata of the Wernecke Mountains, as was suggested by Gross (1965a). Hematitic breccia is known to form entire mountains near the Yukon-Alaska border (Delaney, in prep.). At some point during deposition of the "Thundercloud Subgroup," the mechanism became inoperative. Perhaps sea-floor spreading or fumarolic activity ceased. This time is marked by the onset of deposition of the "Shezal Formation," the gray, pyritic matrix, laid down under reducing conditions. Similarly, the transition from the greenish "Mt. Berg member" to the reddish "Lukas Creek member" marks the initiation of the mechanism.

As pointed out earlier, this is a tentative model which leaves some questions unresolved. Acceptance or rejection must await more complete knowledge of the regional geology, especially to the west, and detailed geochemical comparison, particularly of trace elements, between metalliferous oceanic sediments and late Proterozoic iron-formations.

The Rapitan iron-formation is one of the largest iron deposits in North America. It was deposited under glacial conditions along an active (?) basin margin which ran parallel to the present-day Pacific margin. Its origin is probably related to this unusual combination of tectonic and climatic setting since very similar deposits are known from the late Proterozoic in other parts of the world. It is probably a marine chemical

![Figure 8](image-url)

Figure 8. The model presented here for glaucone iron-formation is reinforced by the experimental work on basalt-seawater interaction by Seyfried and Bischoff (1977). Their results are graphically summarized here. The diagram on the right shows the rate of increasing concentration of dissolved species in a seawater-basalt mixture of 50:1 at 260°C and 500 bars, conditions analogous to those which may exist at a spreading center. As can be seen, iron and silica are most effectively dissolved. The diagram on the left shows the amount of precipitation of those elements after dilution with normal seawater in varying amounts for 24 hours at 25°C. At lower temperatures and salinities the rate of precipitation would be even greater. It can be seen that iron precipitates most rapidly and manganese least rapidly, thus effectively separating them. Rates of precipitation of iron and silica would vary with dilution and temperature, giving rise to the laminations of the iron-formation.
deposit derived through hydrothermal interaction with oceanic basalts or ferruginous hydrothermal solutions in breccia complexes, and precipitated under the strong physio-chemical gradients which must exist in the vicinity of a sea-going glacier.

CONCLUSIONS

A wide variety of economically promising mineralization types and styles has been discovered in Middle to Upper Proterozoic rocks which are exposed in a remote and little explored terrain. Some of these occurrences (the Redstone copper and Gayna River zinc) are similar to types well known in other parts of the world. Others are either a new kind of metallicogenic system in North America (the Wernecke breccias) or allow new insights into a widespread but hitherto little studied class of deposit (the Rapitan iron-formation).

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Regional Aspects of the Helikian (Precambrian Y)
Little Dal Group and Correlatives,
Northwestern Canada

by

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ABSTRACT

An Helikian age has customarily been assumed for the Little Dal Group of Mackenzie Mountains because of its position unconformably beneath probable Windermere equivalents including unequivocal tillites. Preliminary stromatolite biostratigraphy suggests that the group spans the Middle/Upper Riphean boundary, at 950 ± 50 Ma, thus supporting the earlier assignment. Correlation with lithologically similar successions at Brock Inlet and Victoria Island is supported by stromatolite assemblages.

Little Dal sedimentation commenced with marked transgression and deepening. The initial deposits of basinal character display persistent tendencies to development of, first, diagenetic nodules of limestone in subtidal shales and, second, limestone-shale rhythms. Fossil metaphytes and Cheaslea occur in the latter. Equivalents on the platform to the southeast are dominated by grainstones and stromatolite biostromes. Northwesterly shoreline trends characterize the basal and middle Little Dal; however, suggestions of northeasterly trends appear at intervening levels.

Gradual basinwide shoaling led to deposition of an ooid grainstone blanket, followed by supratidal platy dolomites and then thick gypsum. Gypsum deposition was followed by a pulse of peritidal clastic sediments whose isopach trends parallel those of the basal clastic strata of the group. As clastic supply diminished, deposition of peritidal carbonate sediments with minor clastic intervals recommenced and persisted to the end of Little Dal sedimentation.

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Depositional Environments and Stratigraphic Setting of Rocks of the Tsezotene Formation and Katherine Group, Mackenzie Fold Belt, Yukon and Northwest Territories, Canada

by

Darrel G. F. Long

ABSTRACT

Proterozoic rocks that underlie the Rapitan Group (presumed Windermere equivalent) in the Mackenzie Fold Belt are assumed to be Belt-Purcell equivalents. They record a prolonged history of stable-platform conditions. The Tsezotene Formation, a predominantly argillaceous sequence occurring near the base of this succession, records—on a large scale—gradual basin filling (or shallowing) prior to deposition of predominantly arenaceous rocks of the overlying Katherine Group. Superimposed on this general coarsening-upward trend within the Tsezotene Formation are a large number of smaller scale cycles, which reflect local shoaling in the form of subtidal, tidal flat, barrier island, and lagoonal environments. The overlying Katherine Group represents the deposits of a rapidly prograding fluvial (deltaic) sand sheet, which advanced at least four times over shallow water marine sands, shales, and minor carbonates. The destructive phase of each of these major progradations is marked by extensive ripple-laminated sandstones which are in turn overlain by argillaceous rocks with or without carbonates. The last destructive phase is followed by deeper water lithologies in the overlying Little Dal Group.

Conference Discussion

Rolland R. Reid, moderator

Rolland R. Reid, University of Idaho: Tonight we will ask Bob Meyer, the chief geologist for the Bunker Hill Mining Company, to remark on the geology of the Bunker Hill Mine.

Robert L. Meyer. Bunker Hill Company: The Bunker Hill, largest of the Coeur d'Alene district mines, is located at Kellogg, on the south fork of the Coeur d'Alene River and near the western limit of the district. In terms of mineral belts associated with groups of mines in the district, the Bunker Hill Mine occupies a nearly western position on what has been labeled the "Page-Galena" belt. Towards the east, other mines on the Page-Galena belt include in sequence the Crescent, Sunshine, Polaris, Silver Summit, Coeur d'Alenes, Coeur Project, and Galena.

Discovery of the Bunker Hill ore deposits is shrouded in legend, including that of a jackass with more knowledge than the prospector he served. If all the florid declarations are stripped away, the probable facts are that late in August or early in September 1885, Noah Kellogg, prospecting out of Murray, discovered the Bunker Hill outcrop. In order to avoid sharing his discovery with his grubstaking partners, he arranged for Philip O'Rourke, a former Leadville miner, to locate the Bunker Hill claim on September 10, 1885, while Kellogg acted as witness and wrote the location notice. Through a series of partnerships and sales, the Bunker Hill and Sullivan Mining Concentrating Company was incorporated in July 1887. Five years later, deepest access to the orebodies having been reached from creek level in Wardner, and with the mill at the present site of Kellogg being served by aerial tramway, an ambitious decision was taken to drive a long entry to tap the orebodies at mill elevation from the Kellogg side. Tunnel work began in 1893 and was completed in 1902. In 1917 the lead smelter began operations, and in 1927 the electrolytic zinc plant was completed.

In 1956, the corporate name was shortened to the Bunker Hill Company. Through merger in 1968, the Bunker Hill became a wholly-owned subsidiary of Gulf Resources and Chemical Corporation.

Since its discovery in 1885, the Bunker Hill Mine has operated continuously, except in times of labor stoppages, producing through 1976 a total of 34 million tons of ore containing 2.6 million tons of lead, 933,000 tons of zinc, and more than 125 million ounces of silver. Production normally comes from eleven or more separate orebodies mined over a vertical range of 3,200 feet above sea level to 800 feet below sea level. Active mining has taken place down to minus 1,800 feet. Main entry is through the Kellogg Tunnel at 2,400 feet elevation, and access to orebodies below that level is by means of three major inclined shafts and other auxiliary inclines. In total, over 100 miles of major horizontal openings, as well as some 6 miles of shafts and raises, are maintained.

The Bunker Hill ore deposit and its constituent orebodies occur within Ravalli Group rocks of the Precambrian Belt Supergroup.

Several major structural features dominate the general broad framework in which the Bunker Hill Mine is located. To the north is the Osburn fault which brings older Priehard Formation rocks opposite the mine formations. To the east, the general bounding structure is the Alhambra fault, a major thrust, and to the west and south is a large anticline. The mine area lies on the north limb of this fold, which establishes a west-northwest to northwest trend for bedding planes. As is true with many folds south of the Osburn fault, the axis of this anticline plunges southwest; thus, the formations on the north limb dip steeply to the north-west, or are overturned steeply to the south or southwest.

In the Bunker Hill Mine, the north limb is occupied by Revett Formation rocks and a lesser amount of St. Regis Formation. Although past literature places the bulk of Bunker Hill orebodies within what is referred to as the "Revet-St. Regis transition zone," recent stratigraphic work by Brian White establishes the Revett Formation, with few but important exceptions, as the host for Bunker Hill orebodies.

Central to the location of many of the orebodies is the Cate fault, a strong shear zone that is braided in places. In both plan and section the Cate shear zone is sinuous, with an average strike of N. 27° W., and average dip of 41 degrees west-southwest. Although throughout the life of the mine the close relationship between the Cate fault and major orebodies was obvious, in fact constituted one of the cornerstones of development, the causality of that relationship has more recently come into question through structural research, of which you will hear more from Dwight Juras.

North and northeast of the Cate fault, in its footwall, a series of moderate- to steep-dipping persistent shears join acutely with the Cate in upper levels or parallel it at lower elevations. This set of faults appears to have a common structural setting. Our evidence indicates principally normal movement, although some right lateral component is possible. Much, if not all, movement appears to be post-ore, and preliminary stratigraphic correlation suggests offsets of 1,000 to 2,000 feet.
Orebody in the Bunker Hill Mine are varied in their structural settings, fabric, and mineralogy. A growing body of evidence associates the structural elements of many orebodies with at least one broad but subtle west-northwest fold. Fold-generated structure fabrics are mineralized in the hinge and near-hinge limbs of the broad flexure to form large complex ore zones. Ore-associated elements include sphalerite-galena-siderite-filled reverse shears, replacement ores of stratiformlike fabric composed of both sphalerite and galena, and principally sphalerite replacement as fine crackle breccia and regular dense soaking. Development of these various fabrics appears to be dependent on the location relative to the hinge, the lithology of the host unit, and perhaps the stratigraphic horizon in the Revett Formation.

Farther out on the north, overturned limb, smaller parasitic folds, broken by reverse shears, are the sites for localized but rich sphalerite-galena ore shots. Collectively, orebodies of this type have been classified as "Bluebird" ore, characterized by a mineralogical suite consisting of sphalerite and galena, commonly with sphalerite in excess of galena, and the presence of pyrite in variable amounts. Gangue minerals are bluish black quartz and some siderite.

An important set of orebodies are true veins or fissures which strike northeast or east and dip southerly from 40 to 70 degrees. Those located in the floorwall of the Cate fault are referred to as "link" veins for their apparent structural relationship with the Cate and Dull faults. Evidence indicates they originally developed with quartz-sulfide fillings of extensional openings. Mineralogically, they contrast significantly with the Bluebird type; galena is the principal sulfide, with lesser amounts of sphalerite, chalcopyrite and tetra- hedrite, and rare boulangerite and pyargyrite.

Gangue is principally white quartz with some siderite. Original filling textures have been destroyed by subsequent stress, resulting in the development of steely and schistose galena which flows around boudins and rods of siderite and chalcopyrite. Both mineralogically and structurally, veins of this system have been termed "Jersey" type.

A third type of orebody, important in the past but unfortunately with no currently productive analogues, consists of thoroughly brecciated zones. They are generally associated with junctions of major faults or narrow corridors between faults in thin-bedded, brittle quartzites. Recurrent shattering appears to have been accompanied by waves of sulfide-carbonate filling and replacement. From a core of nearly solid galena with only remnants of sphalerite, siderite, and quartz, the intensity of brecciation decreases outward through a zone of rather dense reticulated siderite-filled fractures partially replaced by galena and sphalerite to an outer fringe of narrow sulfide-siderite fractures too widely spaced to be mined economically. The model for this type of orebody was the March ore shoot, by far the most productive. Over a slope distance of 4,500 feet, the March produced nearly 7 million tons of ore averaging 10.3 percent lead and 4.5 ounces of silver per ton.

Are there any more March orebodies? Bunker Hill believes that if not a March, there are still important orebodies to be discovered. The company's faith in this premise is reflected in the ambitious research program begun two years ago and continuing today.

Reid: Our next presentation will be on the Lucky Friday Mine at the Hecla Company, and Roger Lillibridge will make the presentation.

Roger Lillibridge, Hecla Mining Company: I have been asked to give a short talk on the Lucky Friday Mine, which is located on the extreme east end of the Coeur d'Alene district. The outcrops at the surface are St. Regis rocks; whereas, at depth, from say 2,000 feet down, the mine is in the Revett. The history of the Lucky Friday is fairly simple. The claims were located about the turn of the century. The entire property consists of three claims. I cannot imagine anyone who could position three claims to encompass that beautiful orebody. They did a fine job.

In 1942, the Lucky Friday Mines Company was incorporated, and at that time production began in a small way. Production was very small until the vein was opened to about the 2,000-foot level. At that point, the vein enters Revett rocks. The vein responds to the quartzitic environment of more brittle rocks and expands to considerable length. From 2,000 feet to the present bottom level of 4,450 feet, the vein length is approximately 1,600 feet. At the Lucky Friday we do not have the same problem as they do at Bunker Hill, because we have just one vein, maybe a few very short splits, but primarily one vein. That is the only thing we have to worry about at this time. I would like to find more, all of us would, but at the present we have just the one vein. Access is by means of a very short adit into the No. 2 shaft. Presently there is a project afoot to sink a new shaft starting at the 4,450-foot level, and we are planning this shaft for a total depth of 5,000 feet. If all our predictions hold true we will be a very deep mine by the time we are through. Maybe deeper than 9,400 feet.

Currently we are producing approximately 700 tons a month: grade—14.9 ounces of silver, about 10.5 percent lead, and 1.4 percent zinc. Total production through 1976 is about 3.5 million tons, 55,000,200 ounces of silver, 34,000,700 units of lead, and 3,000,800 units of zinc. Not quite the production of the Bunker Hill, but quite respectable. Current production is from the 4,050- and 4,250-foot levels, and the 4,450-foot level is being developed.

From the surface to the 2,000-foot level the vein was very spotty and very short. I believe that on the surface
or on the No. 1 level, they had a vein of some 50 feet long by 6 inches thick. Now this is encouraging and that’s all. There were some pretty hearty people in charge at that time. John Sekulic did a fine job. There was not a great deal of change until the vein was opened up on the 2,000-foot level. At that point the vein entered Revert rocks, and the length and width of the vein increased drastically. We have a rough diagram of the 3,050-foot level of the Lucky Friday. On the north-east, the vein is apparently limited by the Star fault and actually a small subsidiary fault that we call the North Control fault. It has a strike of approximately S. 80° E., and dips comparatively flat, about 75 degrees. To the southwest we are presently developing some additional ore on a hanging wall section of the vein that has considerable promise to it. Throughout the life of the mine we have seen a much higher silver content on the west end of the orebody and since this extension of the vein is on the west it also has a much higher silver content. Silver-lead ratios run as high as 500 to 1. Beautiful ore. Mostly tetrahedrite. In the central part of the vein the mineralization consists of comparatively fine-grained argentiferous galena, some fairly dark sphalerite, a considerable amount of iron, some siderite, and a considerable amount of quartz. Other gangue minerals are not evident. The vein intersects the beds at a low angle.

In most places, however, since there are at least six folds, high angle, very tight synclines and anticlines in the mine area, the fact that the vein crosses beds at a low angle gives it a comparatively irregular strike. This does give us a considerable problem in development, since development is with a system of laterals and crosscuts.

A schematic of the 4,250-foot level shows that the irregular nature of the vein has diminished somewhat, although we see at least one right angle turn caused by the most southerly of the anticlines in the vein area. These small folds are on the south limb of a major district anticline that strikes generally southeast, as most of the folds do in the eastern part of the district.

Reid: Continuing with the Hecla Company operations we’ll next have a presentation on the Star-Morning Mine by Richard Leep, a geologist for that mine.

Richard W. Leep, Hecla Mining Company: The Star Unit is a combination of two mines, the Star and the Morning, operated by Hecla Mining Company. Ownership is by the Bunker Hill Company (70 percent) and Hecla (30 percent). The Unit is located midway between Burke and Mullan, Idaho, with access via a 10,000-foot adit from the old Hecla Mine location in Burke. Milling and surface facilities are in Burke.

The Star mines were acquired by the Sullivan Mining Company, a 50-50 operation by Hecla-Bunker Hill, from the Star Mining Company in 1924. In the early 1950’s, Bunker Hill obtained full ownership, and Hecla became the operator for a service fee. The Morning Mine, which was bought from its original locators in 1887 for the sum of $10,000, had become an ASARCO property when it was leased by Hecla in 1961. At that time both the Star and the Morning were combined under the present operating agreement. In 1966 Hecla bought the Morning from ASARCO, and a new 50-50 agreement was drawn up between Hecla and Bunker Hill, to become effective in 1981.

The Star Unit is serviced by the No. 4 shaft from the 2,000-foot level (Burke adit elevation) to the 7,900-foot level, and by the No. 5 shaft from the 2,000-foot level up to the 1,450-foot level. Secondary access is provided by the old Star shaft and winze on the western edge of the orebody.

Production averages about 1,100 tons per day, with a grade per ton of 2.7 ounces of silver, 5.5 percent lead, and 6.2 percent zinc. Most of current production comes from seven major veins. The most prominent of these is the Main Vein on the Star-Morning fault. The Main Vein has been mined continuously from the surface (elevation over 5,400 feet above sea level) to the 7,900-foot level (elevation 2,100 feet below sea level). Maximum stress length was nearly 4,000 feet near the surface, and the strike length in the lowermost levels is about 1,000 feet. Vein width varies greatly, but averages 4 to 6 feet. The immense size of this sulfide ore shoot, as well as its near vertical dip, is unique.

The remaining veins are south and east of the Main Vein and are associated closely with the Morning East fault. Of primary interest is a series of mineralized fractures which depart into the south wall of the Morning East fault at small angles (15°-30°). These fractures are in places mineralized for as much as 500 feet along strike and extend nearly 1,000 feet vertically. The Noonday North Split, Noonday, Motor Barn, Space Split, and Grouse veins are the major producers of this category. Sporadic mineralization has also occurred along the Morning East fault, with the major production between the 5,900-foot and 7,100-foot levels.

Most of the ore mineralization in the mine occurs within the Revert Formation of the Belt Series. Some veins were mined near the surface in the St. Regis Formation, but with depth the Revert-St. Regis contact moves to the east and out of the picture. Beds strike northwest and dip steeply (60°-80°) to the northeast. This pattern is fairly constant throughout the mine, and only gentle folding is evident. Veins cut bedding at high angles, with strikes N. 70° W. to N. 80° W. and dips 80° N. to vertical.

Mineralization occurs as fillings, in fault zones, and in fractures, often with subsequent fault movement indicated by well-developed slickenides in the vein material. Some indication of resulfurization of vein material is present but is not the general rule. Ore
mineralogy is primarily fine-grained marmatite and galena with small amounts of tetrahedrite. Gangue minerals consist of 50% sphalerite, 25% quartz, calcite, barite, siderite, chalcopyrite, and pyrite. Subtle zoning is evident in most veins, and zinc is generally prominent on the west end of shoots and lead on the east. Some remobilization of galena is evident, and some good samples of "chertized" galena have been collected. Rare showings of disseminated galena have been detected in the wall rocks.

In summary, the Star Unit is a unique and persistent member of the Coeur d'Alene district family of mines. Since production began in the early part of the century, it has produced nearly 25 million tons of ore containing 68.4 million ounces of silver, 1.6 million tons of lead, and 1.5 million tons of zinc (half of the zinc produced in the district). Although low in grade by district standards, it has persevered and remains an important force in the economy of the "Silver Valley."

Reid: Now we will move to the ASARCO properties in the district. First we will have a presentation on the Galena Mine by Joe Lucchini.

Joe Lucchini, ASARCO: The Galena Mine is situated in Lake Creek Gulch, approximately 2½ miles west of Wallace. The name Galena is somewhat of a misnomer, as the current production is in excess of 3½ million ounces of silver.

The property was first staked early in the 1880's. There is now a total of 72 claims on the entire property. The first mining began in 1917, and the ore first found was in the upper regions or upper levels of the mine. It was a low-grade lead-silver ore, which was often mined at a loss. In 1918, the Callahan Mining Company obtained an option and purchased the property; and from that time until 1932, it was explored on ten levels to a depth of 1,600 feet. The mine then remained idle from 1932 to 1947. In 1946, the property was leased to the partnership of ASARCO and Day Mines. Callahan retained a 50 percent share; ASARCO and Day Mines split the remaining 50 percent, with ASARCO getting 75 percent and Day Mines 25 percent. After extensive geologic studies indicated the possibility of ore at depth, shaft sinking was begun and extended to just below the 3,000-foot level in 1947. Exploration work began and silver-lead and lead-silver veins were discovered south of the shaft on the 3,000-foot level. By late 1951, none of these were of any great economic significance; but in the fall of 1953, a major structure was crossed in a crosscut to the southeast. This was the Silver Vein, and this vein became and still is the largest producing vein in the mine. The upper levels of the Silver Vein are mostly mined out, but the vertical extent is from the 2,050-foot level down to the 4,600-foot level, and it continues below.

Production at the mine began late in 1953. There are three present shafts at the Galena Mine: the Galena shaft which extends to 1,507 feet below sea level; the Callahan shaft, which is an internal shaft that goes down to the 2,800-foot level (this shaft is used primarily for ventilation and can be used as an escape way); and the No. 3 shaft, the newest and the deepest shaft, reaching 5,316 feet at the bottom, which is 2,088 feet below sea level. Currently, the 4,000-foot level is the lowest level being developed. Mining is progressing throughout the mine on approximately 10 of 52 veins of silver-copper ore levels. The main at the Galena Mine is composed mostly of silver and copper. The grade potential averages approximately 28 ounces of silver and 1 percent copper.

The veins, which are ore shoots, contain mostly siderite in which are found lesser amounts of chalcopyrite, tetrahedrite, and some galena. Other gangue minerals are pyrite and quartz. Several other minerals found are jamesonite, boulangerite, and stibnite.

The veins have formed along preexisting joints and fractures in two primary rock formations of the Belt Supergroup, the Revett and the St. Regis. Most of the Galena Mine's ore is found in the upper Revett, which is predominately quartzite. It is estimated that approximately 80 percent of the ore is found in this area. There is a transition zone, which is primarily argillite, between the Revett quartzite and the St. Regis. The transition zone is a gradational feature, as there is no definite contact between the two formations. A minor amount of the ore is found within this area. The veins have a nearly vertical attitude and range in widths from less than 1 foot to more than 20 feet for an average of 7 feet.

There are four main areas of mineralization within the mine. The mine is approximately 1 mile in length to the east of the Galena shaft and 1 mile to the west. On the west side of the mine are the Argentine and West Argentine areas, which are about 2,000 feet apart. On the east side of the mine are the Silver Vein and the southeast areas. One feature these mineralized zones have in common is that the veins generally strike in the northeast-southwest quadrants. The bedding often has a considerable influence on the rake of the veins and generally dips to the north.

In exploiting for some of the veins, it is necessary to rely on previous work that has been done on the upper levels and on diamond drilling. The diamond drilling is laid out on a predetermined pattern, computer-developed by the U. S. Bureau of Mines. The holes are oriented to the northwest-southeast, which should eventually cross the general strike of most of the veins in the mine. The holes are spaced at intervals of approximately 250 feet.
Production in the mine utilizes cut-and-fill mining. The mill was built in 1926 and was enlarged in 1960, from an original capacity of 450 tons of ore per day to 750 tons per day. The concentrates are sent to El Paso, Texas.

The Galena Mine will continue to be a major metal producer in the Coeur d'Alene district as well as a large contributor to the economy of the area for many years in the future.

Reid: Continuing with the ASARCO properties, we will next have a presentation on the Coeur Mine, presented by Joe Wallace, geologist for the Coeur.

Joe Wallace, ASARCO: The Coeur Mine is located approximately 1 1/2 miles south of Osburn, Idaho. The property is owned by Coeur d'Alene Mines and leased from them by ASARCO.

In the early 1960's, ASARCO examined the available data from mines to both the east and west of the Rainbow property, which is now the Coeur. Projecting the geology from these mines into the property, ASARCO decided that there was a pretty good chance of favorable host rocks being found. In the early 1960's ASARCO decided to sink a shaft and the work began in late 1965. When the shaft was completed, development of the lower levels (2,800, 3,100, 3,400, and 3,700 feet) proceeded. The favorable host rock, the Revett Quartzite, was found as well as very good mineralization. As the lower levels were developed initially, it was thought that the wedge of the Revett Quartzite and the contact between it and the St. Regis was fairly comparable with the bedding—in other words, more or less along general bedding planes. As development continued upper levels were driven above the 2,800-foot level. It turned out then that the wedge of quartzite (the northern edge of it) was actually steeper than the bedding itself. This led to the possibility of either a larger gradational contact in the upper levels or some possible facies change in the Revett-St. Regis contact zone. As the bedding in the St. Regis and Revett are fairly consistent in strike and dip throughout the Coeur Mine, there is very little major folding or faulting within the rocks themselves. The lack of distinct rock type changes—due to the gradational nature of the contact zone and the lack of definite marker beds in the mine—precluded any direct positive correlation from level to level.

This slide shows the general relationship of the major vein structures to the bedding and to the St. Elmo fault to the north of the mine area, and the Polaris fault to the south. The structures in which the veins occur are thought to be primarily tension fracture openings between the two faults. The vein structures themselves have been followed into the softer argillites and argillaceous quartzites on primarily the east end of the general structure; however, the mineralization tends to pinch out as the softer rocks are approached by the vein. As far as mineralization in the Coeur Mine, the quartzite generally becomes more pyritic nearer the veins. Near the veins are some strong and prominent lenses of pyrite and massive pyrite. The veins generally cut the bedding at approximately 35 degrees and dip to the south. The veins strike about N. 80° W., that is, in the 356 vein and most of the 400 vein. The 400 vein does have a pronounced fish hook on the west end that again probably relates to the movement along the Polaris fault in some manner which we are not sure of yet. The 483 vein, on the far east end, has not been fully developed. It has been cut on a couple of levels. The north end tends to turn toward the east, and the south end toward the west indicating possible ties with the Polaris and St. Elmo faults.

The primary gangue materials in the vein are siderite with quartz present in many places, mainly as tension fracture fillings within the vein. Occasionally, some calcite is scattered through the vein and as fracture fillings in the rocks. Argentiferous tetrahedrite is the primary mineral being mined at the Coeur, with chalcopyrite occurring in many parts of the mine and most strongly in the 356 vein. Freibergite from the tetrachloride family has been positively identified in one or two locations during initial development; we just think of the mineral as mainly argentiferous tetrahedrite. The tetrahedrite is quite massive, very fine-grained, and only on one or two occasions have we ever found any tetrachloride crystals or tetrahedral forms in the mineral.

We are currently mining the 356 and 400 veins, with some development occurring on the 483 vein. The 400 vein is approximately 1,000 feet long, 2,100 feet in vertical extension, and on an average about 8.5 feet wide. The 400 vein is unusual in that it appears to bottom out below 3,400, and as it becomes quite a bit narrower and shorter, the value also decreases considerably. The upper extent of the 400 vein has not been completely determined.

The 356 vein by contrast is generally narrow, averaging in places 5 feet in width, but in many places being more like 1 to 2 feet in width. It is about 700 feet long and 1,100 feet in vertical extent, and it appears to be dying out above the 2,800-foot level, although, the vein seems to be getting better as we go deeper. The 3,700-foot level (our lowest drifted level) has ore that is 5 to 8 feet wide in places and of reasonably good grade. Our average mine grade per ton runs 20 ounces of silver and 0.7 to 1 percent copper. The silver-copper ratio runs about 23 to 1 (23 ounces silver per ton to 1 percent copper). We have several other veins intercepted in drill holes and a couple of places with some minor development in the way of exploration, and these are in the works yet as far as future development goes. We have presently seventeen stopes which are mineable or pre-
pared for mining, and five other raises that have yet to be prepared for actual stopping to begin.

Reid: Next we come to Day Mines. Garth Crosby, chief geologist for Day Mines, will discuss the Tamarack Mine.

Garth Crosby, Day Mines, Inc.: The Tamarack Mine is located in the northeast part of the district, about 6 miles north of Wallace, near the mining town of Burke. The property is owned 100 percent by Day Mines and is a merger of several predecessor companies. The Tamarack and Custer Mining and Milling Company was one of thirteen companies merged into Day Mines.

The property has produced about 3 million tons of lead-zinc ore, averaging 10.5 percent combined lead and zinc between the years 1901 and 1958. There were some interruptions in production.

Silver-lead ratio averages one-half, that is, ½ ounce of silver to 1 percent lead. The local area contains about a half dozen major mines along two major mineral belts with a production of approximately 23 million tons. So, this is a major portion of the Coeur d'Alene district. The Hecla mineral belt is made up of the Interstate Mine on the west end, then the Tamarack, Standard Mammoth, Hecla, and Marsh mines. Other minor, much smaller producers can be included in that mineral belt, bringing the total to about fifteen mines, both major and minor.

The Tamarack Mine is developed by a series of adits numbered 1 through 7, the upper six being in the Ninemile drainage and No. 7 is in the Canyon Creek drainage, just below the town of Burke. Three shafts, progressively downward from the No. 5 level are 600 feet, 1,200 feet, and 900 feet deep. These workings exist through a vertical range of about 4,500 feet and a horizontal range along the vein structure of about 3,000 feet. Seventeen separate veins have been mined from outcrops at elevations of 5,000 feet down to 1,100 feet. Major production has issued from veins found in the Burke Formation, and only minor production has issued from upper Prichard rocks. This Burke-Prichard stratigraphic position is one of the two major positions known to be productive in the Coeur d'Alene district.

The mine is set in an interesting framework of geologic elements. Major district-scale faults, known as the Puritan, Standard, Oreana, and Niles faults, are located along the large antiform plunging generally northwest. Thus, the sandy and quartzitic rocks to the lower Burke and upper Prichard course through the area of the Tamarack Mine. The veins are generally located along the north limb of this antiform.

The mine is located at the southeast edge of the Gem stocks, monzonitic in composition. The ore minerals are galena and sphalerite, marmatite sphalerite. The gangue minerals fall into two general categories characteristic of the mines around the Gem stocks. Quartz-pyrite and siderite are typical in higher elevations and at greater distances from the Gem stocks. As the Gem stocks and depth are approached, gangue minerals are pyrrhotite magnetite, granulite (as iron silicate), garnet, and biotite. Galena and sphalerite remain texturally unique in the higher temperature gangue minerals at depth in the veins. Zoning in individual veins is characterized by high lead-zinc ratio in upper portions tending to high zinc-lead ratio in the deeper parts. Similar lead-zinc and zinc-lead ratios apply for the mine from top to bottom as a composite of all veins. Alteration is chiefly bleaching, some chloritization, and a very extensive hornfels development, especially near the stock and at depth.

Current activities in the mine are composed chiefly of exploration and development along two geologic trends. Targets are above the Tamarack 7 level where all the work is being done. The targets are ores in the Burke quartzites or somewhere near the Burke quartzites in upper Prichard rocks. This creates a unique exploration project for the Coeur d'Alene district where we are actually searching upward instead of downward in a district that has great depth extension. Some narrow ores have been found of limited horizontal extent, about a foot or less wide on the average. But minor amounts of stopping are being started on two of these veins in the hope that as they progress upward into the more brittle, quartzitic Burke rocks, there will be some improvement.

A second area is being entered in the northwest end of the mine, where showings in the Burke rocks were cut in diamond drill holes, and is presently being opened with a cross cut.

On the field trip tomorrow, we will see the unique rock and ore relationship that is improvement and impoverishment of veins as the rocks change. The narrow veins are very high-grade sulfide veins, assaying 50 percent combined lead and zinc over narrow widths. Both the width and grade of the veins change very distinctly and in a very clean-cut fashion with the change in rock types. Changes occur over very short horizontal distances, less than 5 feet, and in some places less than 1 foot.

Reid: The Sunshine geology staff was unable to make a presentation tonight so I have asked Earl Bennett of the Idaho Bureau of Mines and Geology staff to make a brief presentation based on public information on the Sunshine Mine.

Earl H. Bennett, Idaho Bureau of Mines and Geology: Most of the information I will be discussing is out of Bulletin 16 which was published several years ago by the Idaho Bureau of Mines and Geology. I would like to ask Merle Hutchinson and Don Springer to please
correct me, if I make any glaring errors during this presentation.

No discussion of the Coeur d'Alene district would be complete without the history of the Sunshine Mine. The original location was made by Dennis and Truman Blake in 1884, who located the Yankee lode on Big Creek. They worked the area for several years rather profitably and then the company was taken over by Daniel Pierce in 1910. Pierce did a great deal of development work through persistent driving of the Pierce tunnel in which he finally intercepted a fairly large ore body.

In the early 1920's the Sunshine Mine drove an adit parallel to the Pierce tunnel and after sinking a shaft developed the first fabulously orebody on the 1,700-foot level. In May 1935 the company started to sink the Jewel shaft, which consisted of four, 4½-by 5½-foot compartments and is the main access to the mine today. By 1937 the Sunshine was the largest known producer of silver from any single mine in the world. At the present time, the deepest working level is the 5,400-foot level.

More recent highlights of the Sunshine include the tragic fire in 1972 in which 91 miners perished, a year long strike from March 1976 to March 1977, and the corporate battle which currently is underway for the takeover of the Sunshine by the Hunt brothers of Dallas, Texas.

As shown on this map, the Sunshine Mine is located on Big Creek between Kellogg and Osburn and is in the heart of the informally named Silver Belt, which is approximately 6 miles long and 1 mile wide. The Silver Belt and the Sunshine Mine are located between the Osburn and the Placer Creek faults. Vertical displacement on the Osburn is around 14,000 feet and displacement on the Placer Creek Fault is approximately 9,000 feet. Divergent faults that transect the area between these northwest-trending strike-dip faults include, from the north to the south, the Big Creek fault, the Silver Syndicate fault, and the Polaris faults, all of which figure prominently in the location of the ore shoots. Rock units within the mine area are the familiar Belt Supergroup units including the Prichard Formation and the Ravalli Group that includes the Burke Formation. The Ravalli Group is overlain by the calcareous Wallace Formation. The ore in the Sunshine Mine is located in ore shoots in the Revett quartzite-St. Regis transition zone. This is the main ore host throughout the Coeur d'Alene district.

Mineralization in the Sunshine is confined to the Big Creek anticline. This asymmetric fold is overturned to the north with the axial plane inclined approximately 58 degrees to the south. The fold is localized between the Big Creek and the Silver Syndicate faults. This fold is similar to the Tyler Ridge flexure. You can see that the two major ore shoots within the mine are the Yankee and the Sunshine veins. The Silver Syndicate fault is especially important as an ore-bearing structure within the Sunshine. This fault strikes N. 70°-80° W and dips 65°-70° S., and it is characterized by silver-bearing galena that contains large blebs of tetrahedrite.

There are four major ore zones within the Sunshine: the Sunshine-Polaris vein system, the Chester vein system, the Yankee Girl veins, and the Silver Syndicate vein. Major production at this time comes from the Chester vein system. The Chester vein extends through an area called the "hook" and tapers into the Silver Syndicate fault which is another major orebody within the mine. Gangue minerals within the mine consist primarily of quartz and the carbonates siderite and ankerite. Sulfides include silver-bearing tetrahedrite or freibergite, galena, sphalerite, chalcopyrite, and pyrite. There are also several sulfosalts of minor importance.

With increasing depth there is a change in certain mineral ratios. Copper seems to be quite stable throughout the depth of the mine, but silver and antimony values decrease with depth. Quartz, pyrite, and chalcopyrite also become more abundant in the lower levels of the mine. I have some production figures but only through 1959. I did not have time to get the rest out of the Idaho Mine Inspector's Handbooks, but from 1904 to 1959, 6,143,467 tons of ore were processed netting 187,500,000 ounces of silver, 123,186,726 pounds of lead, 53,800,000 pounds of copper, 41,500,000 pounds of antimony, and 1,922 ounces of gold. Net smelter returns for this period were $151 million.

Reid: I am sure the speakers will be happy to answer any questions you might have. Please go to the microphone in the center and identify yourself and your affiliation.

Dave Beatty, California Institute of Technology: What process do you think created those high manganese zones?

Dwight Juras, Bunker Hill Mine: I think that the manganese is probably derived from manganese-siderite in the vein.

Beatty: Was it not present originally in the sedimentary rocks?

Juras: It could very well have been.

Beatty: Why aren't they offset by the faults?

Juras: They are. That was a reconstruction, pushing the north side west so that it matched the bands on the south side of the fault.

Jerry Farnum, Bunker Hill Company: We have had ore minerals flowing from the rocks into the veins and vice versa and also gangue minerals flowing from the rocks and from the veins. Concerning the sedimentary environment for the original sandstone, I think the barrier bar boys were with the sandstones.
Very quickly, by way of challenging the group, I am going to throw out a few working hypotheses without any explanation, just the bare bones. First, district ores were derived by metamorphic remobilization, or Red Sea or sedimentary-type orebody from deep in the Prichard Formation. See if that gets any comment. Two, the mineral belts in the district are rather strictly west-northwest. Three, certain gangue minerals are derived from the country rock and account for the favorability of certain horizons for precipitation ore. Now what I hope to accomplish is to let people start anywhere they wish.

Jack Roylance, Buxton Inc.: We have heard a lot in the last couple of days about the questions of whether metals came out of rocks into veins, or whether mineralization disseminated from veins into adjacent rocks.

If we assume that diffused mineralization in host rocks did migrate into nearby open spaces to form veins, how do we explain the diffused mineralization?

I will present my answer to this last question, even though I do not have any first-hand exploration experience in the Coeur d'Alene district.

Metals that form insoluble sulfides are anomalously concentrated over wide areas of the Coeur d'Alene district. These anomalous concentrations are generally found in the upper Revett sandstone, near the transition to the overlying St. Regis Formation (argillite, mudstone, siltstone).

I suggest that those metals came from deep in the earth's crust and rose with hydrothermal solutions along a fault zone that is now marked by the Osburn fault. Introduction of the metals, I believe, was preceded by a migration of gases out of the Prichard Formation while this formation was being compacted.

These gases may have flowed with other compaction-expelled fluids up into the overlying Burke and Revett sands. Gases originating in the thick, black Prichard shales and clays may have been composed principally of methane. Probably hydrogen sulfide was mixed with the methane, because the Prichard Formation is characteristically pyritic.

Where would these gases go? They would rise to the top of sand units where clay beds would restrain upward movement of bubbles (which may have formed during the rise of gas through sand). Oilfield-type traps may have accumulated hydrogen sulfide-bearing gas.

The upper part of the Revett sand, by this model, should have been a favorable site for concentrating hydrogen sulfide and, consequently, insoluble metal sulfides. If later a Sullivan (B.C.)-type event occurred in what is now the Coeur d’Alene district, there may have been an extraordinary amount of metal-rich, saline fluids coming up the proto-Osburn fault system. As the fluids reached the upper levels of the Revett, they would encounter local concentrations of hydrogen sulfide gas. Possibly clay beds overlying the Revett sand did not maintain fracture permeability along the fault zone and hot, metalliferous fluids were forced to travel laterally through upper Revett sands.

Precipitation of metal sulfides in the upper Revett sands would be controlled by the chemistry of the formation pore waters as well as the chemistry of the introduced hydrothermal solutions. Hydrothermal solutions can vary with time and may travel by different paths as the mineralizing event progresses. The interplay of these variables could explain metal zonation in the diffused mineralization of the Coeur d'Alene district rocks.

Coeur d'Alene vein systems also show concentrating metallic compositions, such as the difference between the Silver Belt and Bunker Hill's lead and zinc vein systems and the copper-rich Snowstorm veins. There, contrasts may reflect the earlier interplay of hydrothermal and formation waters, if metals later moved from Revett rocks into veins.

Farmin: Very provocative statements. Does anyone have a reply? My reply is that I do not believe the Osburn fault zone is a conduit for the metals, because we do not find metals in the Osburn fault.

Jurase: Some of the things I am going to say may be a little Mickey Mouse but I want to try and revise some of the thinking that some of you may leave with. I would first like to start with some of the statements that Rolland Reid made about the regeneration of west-northwest-trending folds.

In the Bunker Hill Mine we have a very complex structural situation, which my talk tried to simplify and allow you to understand the structures in Bunker Hill. The structures are very complex, and a very detailed sequence of events was not appropriate to that talk. In trying to pin down the age of the metals in the Bunker Hill Mine, I have tried to pin down the exact timing of the metals, and the exact timing is with the west-northwest folds, previous to southside-down on bedding. There are no other structures in Bunker Hill cut by southside-down on bedding faults but the structures that are link veins and fracture cleavage and reverse faults. This kind of mineralization is followed by normal faulting which is post ore. Other warping has occurred previous to the normal faults and after the normal faults. A regeneration of west-northwest folding would have changed the plunge of the fold. It is impossible in my mind to put the age of mineralization in a regeneration of fold events. There are more detailed things that I could give in Bunker Hill, but I think it would be inappropriate at this time. Maybe I can expand upon it in another paper later, but I cannot show you examples on the screen at this time.
I would like to make a comment that I tried to get out of Brian White, but he had left and I got a confirmation that I could make this statement. The workings in the Bunker Hill Mine are quite extensive and for those who visited the mine and saw the model, you can verify that fact. Brian White has studied the Revett Formation away from the veins, covering a large area of the mine, and he can make the statement that the mineralization that appears in strataform in the other orebodies in the district does not exist in the Bunker Hill away from the veins. Where you find the veins you find strataform, and Brian White has gone into massive stratigraphic correlation throughout the mine. He does not believe he has missed any strataform except in certain locations, which I will get to in a minute.

We will reiterate the fact that some of these structures that Mohan showed about the leaching of metals into small fractures is impressive, but it does not prove the timing of the strataform or sulfide rock. I implied before that the strataform sulfide rock could have been distributed into the rock from the fractures of the folding, and then later fractures cutting this sulfide rock, or strataform rock, could have leached the mineralization into them. Has he studied the thin section of the leach zones, and how does that compare with the nonleach zones in relation to the fabric? Nobody commented on that. Would Rollie Reid mention something on that before I go on to the next part of what I wanted to say.

Reid: The fabric in those small leached zones, petrographically, is not different from the rest of the metamorphic fabric. It is simply devoid of sulfide, which has migrated to nearby low-pressure zones represented by later fractures. But we have galena replacing sericite and engulfing sericite and if the galena is taken away in the leach zones, what fills in the places the galena was in, in the leach zones? I would reply on the total pressure in the rock system to close up the voids by just a little bit of late cataclastic adjustment, perhaps on the existing grains.

As long as I am responding, let me touch on a couple of other things. I am of the opinion that we could have recurring folding on the Osburn fault zone, not necessarily deepening the plunge of those folds. I will take a chance to comment on Jerry Farmin's remark a bit earlier about the absence of any evidence of conduit in the Osburn fault zone for sulfides to rise from depth. First I would like to say that as McMannis, Palmquist, and I showed in the Montana Archean in 1973, the ancestral Lewis-Clark line got under way fairly early, the earliest movement started at 2.2 billion years ago, with very strong recurring movements again at 1.7 billion years. With very ancient diabase coming up, one is virtually certain that there is a connection to very deep, even subcrustal upper mantle rocks, in the Lewis-Clark line. I envision a recurrence of movement in that ancestral zone at some point during Belt time. We have got Art Setonson's 1.2 billion year date for the Osburn fault zone in the Page Mine, which again suggests even then a connection with fairly deep, perhaps upper mantle rocks. I can envision mineralizing solutions coming up along the ancestral Osburn fault zone and then all traces of the conduit being obliterated in a very thick cataclastic zone of brecciation that represents now the Laramide Osburn fault zone, and high stress simply driving all evidence of any channels out of that now thoroughly crushed, brecciated, and altered rock.

Mohan Ramalingaswamy, Consultant, Vancouver, B.C.: I would like to comment that lead to zinc ratios in the vein material are very erratic. They vary tremendously; whereas, in the strataform zones they are fairly uniform. From these ratios we can derive that lead has preferentially mobilized, with respect to zinc, and you have got vein orebodies. There are large fault structures in the Newgard area, but there are not big-sized veins as you find in the J-Vein area. Do you have any comments on that?

Juras: That was to be my next comment. Those areas which you described in the mine have something developed structurally that was not present in our understanding when you were working in the Bunker Hill Mine. First of all, the Newgard area previous to the research, before you arrived at the mine, was not known to be the core of the Tyler Ridge flexure. The whole Newgard area is in the Tyler Ridge flexure. Composite maps of the Newgard area show that the mineralization in the Newgard area changes from bed to bed as you follow it up. It is not in the same bed. Second, the area of the J-Vein which you described has recently been shown to me to also be a crest of the parasitic fold of the Tyler Ridge flexure. We are trying to expand on that a little bit more.

Ramalingaswamy: On that I would like to say the Newgard orebodies are in the hinge of that fold. If there was strataform mineralization in the sulfide unit that has been folded, it is going to go to the hinge anyway.

Juras: How does it get to the hinge zone?

Ramalingaswamy: It goes to the hinge zone. The hinge zone of a fold is the area of least stress. Sulfides being unstable will naturally recrystallize in the hinge zone.

Juras: I see, but these same rocks away from the hinge zone do not have any mineralization, say on 23 level.

Ramalingaswamy: They might have mineralization, but it may be of very low grade.

Juras: What you are saying is that the only place in the mine where the orebodies occur is where the mineralization was strataform of higher grade?
Ramalingaswamy: Yes.

Juras: But that is opposite to what occurs near most veins.

Ramalingaswamy: But there are hardly any veins in the Newgard.

Juras: I disagree with that.

Ramalingaswamy: And also the Newgard orebody actually has fairly uniform lead to zinc ratios.

Juras: Everybody talks about lead-zinc ratios, and I have shown in my talks, first of all, that the zinc mineralization precedes the lead mineralization.

Ramalingaswamy: How do we know that?

Juras: Because I see cross-cutting relationships.

Ramalingaswamy: Yes, that is a relationship of sulfide stabilites. You cannot relate sulfide stabilities to cross-cutting relationships. Galena cuts across sphalerite but that does not mean galena came in later. It is entirely structural and sphalerite is fairly stable up to 500 degrees; it does not recrystallize below 500 degrees, but galena recrystallizes at 100 degrees.

Juras: But if it is correct that zinc preceded lead, lead-zinc ratios would not have any meaning.

Ramalingaswamy: Now, for example, I think it was Dr. Green from the University of Idaho who mentioned a paragenetic sequence in a vein. If you look at the bottom of the sulfide stability diagram, it is exactly that from arsenopyrite, tetrabedrite, cubanite, galena, argentite, that is the same zonation you see in the vein Dr. Green was describing. So that is not a paragenetic sequence.

Juras: Cross-cutting relations are not a paragenetic sequence?

Ramalingaswamy: No. Because cross-cutting depends on the sulfide stability. Suppose you take a sphalerite- and galena-rich bed and all you do is deform it, then what are you going to get? The galena will come out and go into a fracture and cut across sphalerite. This is a metamorphic texture; it is not a paragenetic sequence.

Charles Hauntz, graduate student, University of Idaho: I would like to speak just a moment on this business of ratios. A question was asked me this morning concerning lead-silver ratios in the Lucky Friday Mine. I need to clarify the answer I gave. I suggested that the lead to silver ratios in the blue rock in the Lucky Friday Mine are 1 to 1, 1 ounce of silver to 1 percent lead. I said that the lead to silver ratios in the veins range from 1:1 to 1:500, 1 percent lead to 500 ounces of silver, and that is true but some clarification is needed. The lead to silver ratios in the Lucky Friday vein have a method or a system to their madness. Particularly in the west end the lead to silver ratios are 1 to 500. In the middle of the vein they are essentially 1 to 1. You can correlate that with the mineralogy of the vein. The west end of the vein contains much more tetrahedrite, and the center of the vein contains essentially galena with very little tetrahedrite. I am suggesting that the 1 to 1 lead to silver ratio in the blue rock exists because the only mineral in the blue rock is argentiferous galena. The reason for widely spread ratios in the veins is the presence of many other silver-bearing minerals. That does not speak to Mohan’s lead-zinc ratio, because I do not have any zinc in blue rock.

Ramalingaswamy: Now I have seen the majority of the lead-silver ratios do not vary at all in the vein as well as in the stratiform. Now you say the silver varies somewhat, that is because the veins probably have tetrahedrite. It is not necessary for the vein to occur strictly at the junction with the stratiform; this silver-bearing material or tetrahedrite might come from a lower copper-rich unit or tetrahedrite-rich unit, which got mixed up, due to deformation features.

Juras: I am going to restate something which I think. Mohan was saying. Lower in the Bunker Hill Mine, trace amounts of lead and zinc in the rocks are not visible to the naked eye. In the areas where the vein exists, or in the hinge zone, sedimentary features ("soaked rock") have been replaced by galena and sphalerite. This is a remobilization into the hinge area, which I think, Mohan is confirming. Mohan shows slides of the impoverished zones in the "soaked" rock. The "soaked" areas, he implies, are derived from the rocks which have only trace amounts of lead and zinc. Thus, Mohan implies that the metals have moved from the rocks where they occur only in trace amounts to the vein areas where the metals may again be leached and appear as impoverished zones adjacent to the veins.

Ramalingaswamy: Okay, what you are trying to do is prove me wrong. I have not heard a single statement showing that sulfides originated from the vein and stratiform sulfides came from the vein. Dr. Reid and yourself are assuming periods of full length faulting in that area. Sulfides are fairly unstable and do not retain their original textures. Let me say that during the period of folding when the Tyler ridge flexure took place, sulfides concentrated in the hinge zones. Another period of faulting or folding took place when the sulfides were reworked.

Juras: Why did not the sulfides come out during the first north-south folding when the deformation was of the same metamorphic grade? Why did they only come out during the second folding? Why are the veins in the district mostly west-northwest?

Ramalingaswamy: Well, that is probably a favorable point for you; I cannot answer that. Just like before,
plan views of the "J" area have shown that when the stratiform sulfide units hit the "J" zone, you get vein sulfides, and on the 27 level when the stratiform sulfides are on the hanging wall you do not get any veins. Maybe they have not intersected a favorable structure.

Haunz: Mohan (Ramalingaswamy) showed us slides of the "J" vein with a large, wide zone depletion of the disseminated rock. I would just like to say that yester-
day underground in the Lucky Friday for the first time in three years, we found blue rock in contact with the vein. There is no depleted zone around the main vein. The blue rock is not depleted around the main vein. Again, I think that there is a difference between Mohan’s disseminated lead-zinc that he sees in the Bunker Hill and the disseminated lead that I see in the Lucky Friday.

Juras: In one area of the "J" vein I have seen stratiform (galena) abutting against the vein, but I have also seen impoverished zones in stratiform on one side of the vein. The other side of the vein may not appear depleted at the stratiform intersection.

Jim Duff, Bunker Hill Company: For a period of about a year I was the geologist in the "J" area of the Bunker Hill Mine, so I had an opportunity to get a first hand look at a lot of this type of mineralization. I would like to make a couple of points to clarify some of the things that Mohan and others have said. Some of the things I will say are based on information that has become available since Mohan worked there. First of all, there are different types of stratiform mineralization in the "J". We see stratiform, I should not use this term stratiform perhaps, I think I will say disseminated, but we see disseminated galena mineralization not only in quartzites, in hard, vireous quartzites, but also in very argillitic rocks, what we call sulfide argillites. The disseminated galena mineralization can be very extensive, as far as I know it can extend for a thousand feet or more away from the vein areas. I do not know how far exactly. The relationships between that type of mineralization in the veins is not always clear. In one particular case where we have the disseminated mineralization in the sulfide argillite, that particular unit comes right into the "J" area, and as far as I am aware, there is no particular increase or decrease in the lead content of that sulfide argillite unit as it touches the "J" area. Now I do not know how far away from the "J" area the mineralization extends. It is possible it may represent some sort of leakage away from the "J" area.

Another thing is that we do have a lot of veins, par-
ticularly lead veins, in the mine with no disseminated stratiform mineralization associated with the veins. So, you do not have disseminated mineralization, particu-
larly lead mineralization, because the disseminated-
type mineralization is very low grade. Something on the order of 1 or 2 percent or even less lead, maybe 1 ounce of silver per ton and 1/4 percent zinc, and with the low grades, since the grade does not have that great a range, the ratios will not be as variable as in the "J" vein. For example, where the lead grades range from a few percent up to 20 percent and a consistent zinc grade of about 1/4 percent, the lead-zinc ratio in vein mineralization will have a tremendous variation, but you will not see that same extent of variation in the lead-zinc ratios in the stratiform disseminated mineralization.

Farmin: Nobody has ever shown or told me about stratiform tetrahedrite; therefore, that would explain the 1 to 1 ratio and the Lucky Friday discrepancy there. The other is that this stratiform zone that stands a thousand or more feet away from the "J" vein might be very close to another vein.

Norm Radford, Bunker Hill Company: Perhaps this should be entitled confessions of a geologist. We all know that every property in the district has some localized unique stratiform areas as described in the Lucky Friday Mine. The Bunker Hill seems to be well-endowed with stratiform areas, so it is well-studied. We have about thirty ore deposits being worked in the Bunker Hill, but perhaps only three or four of these orebodies show a significant stratiform. On the tour yesterday, we took everyone to the 11 level Newgard-Quill area, which is our classic stratiform location. Then we took everybody down to the "J" area, which is another classic stratiform area. The first thing we always say is "you know we have tourites coming. Where are we going to take them?" Well perhaps the most interesting place to study is a stratiform area. We do not take them to the other 28 places, or 25 places, where the stratiform is rare, if ever, seen. Now the stratiform is primarily seen, as Jim Duff just pointed out, in the "J" or "K" area, in the footwall of the Kruger fault, and in the hanging wall of the Cate fault where we see the Quill, the Newgard, and an area called the Roxcor, which has some stratiform.

We have done a great deal of work on lithology in the Bunker Hill, and I do not think that we did any work with the intent of everyone getting in a hassle over where the St. Regis-Revet transition zone is located. I think the most important thing that we were trying to do is identify those beds that are the most favorable loci for ore. I work at the Crescent Mine and at the Silver Bell. On very rare occasions we see limited areas of stratabound galena with no tetrahedrite. We see, in the veins, galena, tetrahedrite, siderite, and quartz. We may see some quartz in the vein hanging or footwall, or we may see irregular blebs of chalcopyrite or tetra-
heedrite, but we see absolutely no stratabound tetrahedrite or chalcopyrite.
Ed Fields, Bear Creek Mining Company: In order to look at some of the questions and problems that have arisen here, I think it is important to look on a district-wide basis and on a region-wide basis. For instance, we are arguing about the various characteristics of the Bunker Hill Mine, but also look at the Lucky Friday Mine where you do have stratiform mineralization which is not apparently related to any structural features. It is found far and away from the vein structure, and away from some of the major structures, so that area does not seem to have a close correlation or relationship to structures and that may be a true stratiform circumstance. Maybe getting a little farther afield into an undeformed area, go into the Spar Lake copper deposits. Many of the features you see there in terms of mineralization, in terms of the texture of the mineralization, are very similar to the stratiform-type mineralization here; only there you are dealing with copper rather than lead. There you are dealing with an undeformed sequence, an undisturbed sequence relatively flat-lying, where you get cross-beded mineralization with the copper, mineralization which comes up through bedding and suddenly it is truncated, absolutely truncated in the middle of a bedding unit. Some of the controls that have been offered there are diagenetic controls of the mineralization which may have been controlled by some hydrologic interface, where you had a water interface within the pile of sediments and the mineralization came up to this water interface and did not carry out any further. There is very good evidence for stratiform mineralization which looks similar to the disseminated mineralization here. Maybe a third point worth talking about is some of the aspects that Rollie Reid mentioned. At the Cobar Mine in Australia, which is a stratiform copper-zinc deposit, there are arenaceous, maybe slightly dirtier-looking units than what we are dealing with here, but it is a true stratiform copper-zinc deposit. There you have very penetrating deformation, penetrating fracture features which cut the rock, and these actually offset and pull apart very thin stratiform beds. The beds are pulled apart and actually remobilized along the fracture cleavages to form a cross-like structure on a very small, a few centimeters, basis. There you have remobilization along very closely spaced penetrative cleavage which actually almost elongates it in the opposite direction of the actual bedding dip of the orobody. That makes an interesting feature.

Farmin: Anything else? We have spent a lot of time on stratiform. How about flow controls? Do you believe that?

Larry McMaster, Duval Corporation: If the Bunker Hill geologists do not go along with this stratiform idea and remobilization of sulfides into the vein, just where do you propose that all of this silver and lead came from? If all you can say is down below, I find that disappointing.

Farmin: I think ultimately the metals did come from extreme depth, from some deep-seated magmatic source. All I am saying is that I am not sure where they came from in the mantle, but it seems quite an interesting idea that at the Sullivan in higher stratigraphic positions you have fissure veins. In the St. Eugene Mine you have typical lead-silver veins that came from below and deposit only in quartzites. It was one giant leak between that kind of thing and the Sullivan orebody underneath that prompted me to make the bold statement I made about Red Sea-type deposits which obviously come from below somewhere and were remobilized up into the quartzite that is a favorable location for deposition.

McMaster: To direct somebody else in this, Don Springer was saying that some of these veins below the ore shoots or orebodies are very narrow, a quarter-inch wide or so, and then when they hit the favorable horizon they spread out to 7 or 8 feet. It seems amazing that that much fluid material could flow through a small fracture. Maybe it is possible, but it seems very impractical. Now, can he comment on this, and perhaps this is only one small locality, I do not know. I am not that familiar with the district.

Jarat: This is exceedingly common and occurs vertically as well as laterally. The vein is thick in quartzite, decreases to practically nothing in argillite, and is thick again going into quartzite. It happens in both directions.

I commented in my talk about the intersection of the fractures in the district with the bedding at the plunge of the fold. In the Silver Belt district, the plunge of the fold is about 15 degrees. The intersection of the bedding and fractures would be about 15 degrees to the east, in the Sunshine and Crescent Mine areas. Why couldn't the solutions have moved up the plunge of the fold? This way the solutions should not have to come up from the thin areas of the fractures but would come up the plunge of the bedding-fracture intersection. This is very similar to a case at the Bunker Hill Mine. We still have the problem of solving whether the zinc in the Newgard area has come up the Tyler Ridge flexure or a reverse fault on the overturned limb of the flexure.

McMaster: Another thing about your comment on it coming from below: usually it is from hydrothermal solutions. I assume this is what you are attributing it to. Usually you get some kind of vertical zonation; speaking of typical porphyry copper, you have a sequence of base metals that grades away from the heat source. The heat drives the system. From the tour that I had and just talking to some of the other people, there does
not appear to be very much vertical zonation in the ore shoots in the Coeur d’Alene district.

Juras: I agree on that. If you check the reports from the Pine Creek area, vertical zonation has been reported. As the veins were mined deeper, the pyrite predominated and the veins became uneconomical. If you look in the Silver Belt, the Crescent area is capped by galena. I believe the Sunshine Mine is the same way. Garth Crosby might be able to comment on the fact that the Star Mine, which may have set below the Dayrock Mine, has a vertical zonation.

Crosby: I hesitate to comment on Hecla’s orebody. I can comment on the Hercules. The Hercules has very high silver-lead ratios in the upper portions as compared to the lower portions. Also, orebodies in general across the Coeur d’Alene district show higher silver-lead ratios and higher lead-zinc ratios in the upper portions as compared to their lower portions. The higher silver-lead ratios were found at lower elevations in the Star-Morning Mine. Let me add a word of caution, for I have not studied this particular problem, but these ratios were pointed out to me some years ago. If I may, let me reply to the stratahound subject, for this should be in the record. The Dayrock Mine, which is in the Dobson graben, contains veins along the St. Regis-Revett contact characterized by what we have described simply as disseminated ore. As a matter of fact it is not ore when it is disseminated. In general, when a vein is being followed in a drift along its strike, and the vein is a good, commercial grade it will often feather out and fade away into disseminated material until no significant structure remains. The disseminated portions of these veins will broaden often to several times the vein width, maybe as high as 30 to 40 feet. As we progress along the strike, the drift begins picking up vein structure, and slowly the mineralization becomes more compact, more consolidated, and we are back in a 3- or 4-foot-wide vein of good sulfides, and we have ore again. This happens repeatedly throughout that mine in both the St. Regis and the Revett rocks, chiefly in the upper portions of the mine. Calculations show that there is as much metal in the 30- to 40-foot widths as is contained in the compact vein.

Now, another remark on metal ratios. In the disseminated zones the silver-lead ratio was often close to 1:1, closer than the average silver-lead ratio for the average ore of the mine, which was distinctly less than one except for the Hornet orebody. Tops of orebodies and tops of single shoots of orebodies, often have skyrocketing silver-lead ratios. Probably this occurs because lead is absent. In a few places, tetrahedrite exists in these areas, supplanting the lack of lead.

The comment was made regarding the nature of these orebodies as they peter out. I have seen the bottoms of many of the orebodies. The interesting feature seems to be quite consistent in that the metal content of the vein and the vein structure both give up almost completely at the same place. At the bottom of the Tamarack Mine some silicification and pyrite are found, but the vein structures actually do not exist. They have pinched out in many places even more completely than where the veins pinch out laterally. On the trip yesterday, we did see narrow vein structures leading away from the Tamarack Mine; in fact, they appear continuous to other locations, 1,000 to 2,000 feet eastward, in which some improved mineralization has been found. As to where the ore came from, I feel it has been a more complex mechanism than Jack Roylance has suggested. I cannot believe that the original shapes of orebodies even resemble the shapes of the orebodies we are mining today. I think energies have been applied to the metals ever since the metals arrived. These energies have been moving this material around in no small way. One of the evidences is the Dayrock Mine with orebodies of very low plunge angles parallel to the Dobson Pass fault. The overall plunge of the mine is the same as the fault, 27 degrees, an anomaly in the district. Now, in addition, a barren section of the vein can be followed to the east. This is a 100-foot section above the fault, barren of ores. This seems to be a tremendous coincidence, suggesting the possibility that some of the energy that has gone into the movement of the orebody has left evidence of an energy that was parallel to the fault, which has remobilized these metals. I do not like the word “remobilized” as well as I like “mobilized.” I think mobilization is underway far more continuously than we can recognize.

Farmin: The plunge of the Bunker Hill Mine, in general, parallels the plunge of the Cate fault, so that is a parallel. The other thing is that the Dayrock Mine, when the Dobson Pass fault is reconstructed by normal movement, lies directly above the Fresco Mine which is the next lead-zinc mine, so that the zonation is in mixed lead-zinc up through lead with a high silver horizon sandwiched in between. I do not know whether I made this point adequately yesterday, but I cannot understand why there would be zinc in the upper Prihard and overlain by lead in the lower Burke, which is stratigraphically below the St. Regis-Revett, and have the same sequence repeated in the same way between the lower portions which are more argillitic. There is a direct parallel there.

Don Springer, Independent, Osburn, Idaho: I would like to make a few observations, generalizations, and like most generalizations, they are not necessarily true. I believe that most of the silver orebodies we see in the district have or exhibit an incomplete galena envelope. This may or may not be a vein. There are gigantic holes in this envelope and it is by no means complete. In
some instances the galena is in the less favorable host rocks. The Crescent Mine, the Sunshine Mine, the Galena Mine, all have lead orebodies in the upper extremities in the Wallace Formation. I do not know what this contributes, other than perhaps a little on the zoning. In some places, around or on the edges of the silver ore shoots, there is a development of stibnite before the vein itself diminishes to subeconmic dimensions. In regards to disseminated tetrahedrite, I perked up. I think I know where there is an occurrence. I will have to go back and check.

Farmin: Don, I would like you to comment on other metals beside lead, zinc, and silver, which may be mobile. If you comment on say a source or just a localization of where you have a high copper to lead-zinc ratio, or maybe some other elements such as bismuth or antimony, would there be any particular area in the district where you have this type of high ratio, which may be a central source area?

Springer: I have not recognized anything like this. One of the questions that was asked on the field trip was whether mercury had any bearing on ore, or whether mercury had been used as an exploration tool. My answer is that from some of the work I have heard about, the entire district is surrounded by a doughnut of mercury. The mercury is low within the district and high on the fringes and then low again further away, so mercury could possibly be used for finding the Coeur d'Alene mining district. I think we already know it is here, but it might be useful for finding another Coeur d'Alene mining district in Bonner County or across the border, or somewhere else. I thought it was interesting that there is one location where some red gouge was seen, I am not certain now whether it was the Big Creek fault or the Placer Creek fault, and it ran about 1/2 pound per ton. So there are some concentrations of mercury. Some of the tetrahedrite in the silver veins is mercuric. But, as far as some of the more exotic, I should not say exotic, but minerals that are not mined here commercially, I know of none other than the usual that we find associated in the veins.

Michael Paloumaki, graduate student, University of Idaho: I would like to ask Dr. Reid how you concentrate stratabound galena and possibly copper at right angles to the cross beds as in the Lucky Friday, Bunker Hill, and possibly Spar Lake by a shearing mechanism.

Reid: I cannot answer the Spar Lake part of that question, because I have not seen it. I heard Ed Fields say a while ago that the Spar Lake ore had not been metamorphosed, and I do not think the mechanism I envision can be applied to Spar Lake, as I understand that deposit. For the blue rock within the district, I have proposed that cataclastic shear planes, closely spaced, over a fraction of a millimeter apart, have served as the loci for the essentially surface diffusion flow of galena, stress activated, outward from the veins. As it has traversed these cataclastic shear fractures which are completely penetrative, it has encountered chemically favorable interfaces with the bedding that it is traversing, and has precipitated at such chemically favorable intersections. I spoke on a negative correlation with sericite and somewhat positive correlation with heavy minerals in some of the layers. So, each bedding interface is represented at a large number of cataclastic crossings, and each chemically favorable bedding interface precipitates sulfide at that crossing. These now lenticular sulfide grains parallel the cataclastic schistosity on the microscopic scale. This sums up by an integration, if you will, to an appearance of control by various kinds of bedding features, whether it be bedding, cross bedding, slump features, intraformational conglomerates, or whatever in the primary bedding structures.

Ramalingussamy: On Charlie Hauntz's thin sections, I noticed mostly quartz on the bedding features, maybe a little feldspar. He did not say much about that. Right in the band there was sericite. That was the one parallel through your shearing mechanism, but what are the hidden metals that you were talking about, the heavy minerals? How does the galena migrate through the rock and just concentrate there?

Reid: We did one electron probe analysis of the heavy minerals in a typical dark mineral layer and found an assemblage of rutile and zircon. The sericite is not parallel to the bedding in this situation, although it is related to the bedding in that some layers were originally more clay rich than others. The sericite, now, parallel the initial metamorphic fabric which cuts across the bedding in the axial-plane cleavage direction, later paralleled by the cataclastic shear. I am proposing a surface diffusion flow of the sulfide on the surface of the silicates. Perhaps you have had occasion to see a thin layer of frost on a window glass in the wintertime, and see clear areas in the frost where water has, in the frozen form, skated on the glass interface to make water crystals in the center of a recrystallized zone. That kind of a solid surface diffusion mechanism, which can work even below 32° F., could likely work for stress-activated sulfides. As for a nucleation, I think the sulfide may nucleate more readily perhaps on sericite grains, on the lattices of certain heavy minerals, and less favorably on clay lattices. So, I think it is a question of favorable nucleation sites for precipitating the sulfides.

Jim McGregor-Dawson, Minerals Exploration Company: How far away from the veins have you seen actual stratiform galena mineralization? From your diagram you seem to be getting it around the veins all the time. Does it extend away from the veins?
Jurasa: I could recognize stratiform mineralization from the descriptions that are available here, but I don’t know how far they extended because the drifts or stopes do not go beyond the veins at any time. So all I see are pockets of stratiform mineralization left over in the ‘D’ area. That is why you find slight variance, like 5 to 40 lead to zinc ratios.

McGregor-Dawson: In some of the curves you note a decrease in lead as you approach the veins, but you also have a decrease on both sides of your lead high.

Jurasa: That is because of the two veins you see and the pocket of stratiform in between. That is why you get a decrease on both sides.

Jim Aitken, Geological Survey of Canada: We have been considering genesis here, but none of the last three speakers has said anything about the limitations on what is known about the actual age of the mineralization. It seems to me that knowing the age, at least within limits, would completely eliminate some hypotheses of origin. I would like to ask the last three speakers to summarize what is known of that age. Has anyone attempted a lead age? We were told that there is a post-or stock up here. What is the age of that stock and so on? It seems to me that this would eliminate a lot of noise from the system.

Jurasa: Long, Silverman, and Kulp have assigned a Precambrian age on the basis of lead isotopes. I guess Dr. Zarrman took samples of the ‘D’ area and he said it to be Precambrian. But, I do not know; the primary source of the lead may be Precambrian, but the previous three speakers, in my opinion, said that the mineralization occurs along certain fractures or faults. I think that you cannot relate sulfides to silicates. Suppose sphalerite and galena were present recently in, let’s say, the metamorphosed impure sandstones. Then you metamorphose it and what happens to the sulfides. They tend to recrystallize and they tend to go into zones where the stress is released. Therefore, say, galena cuts across sphalerite. You cannot assign any paragenetic sequence to it, because galena automatically is less stable than sphalerite. It has to cut across sphalerite, which is what I have seen in the mines that were previously assigned long paragenetic sequences which to me does not make sense.

Fields: I would like to inquire whether there has been any detailed geochemical studies on the stratiform mineralization versus the vein mineralization to different pathfinder elements? And, also, secondly, what are some of the other examples of stratiform mineralization that are known in the district? We looked at some from the Lucky Friday Mine today, but I would like to see if there are any of the other mines that have this type of mineralization.

Ramalingaswamy: I have seen stratiform mineralization in the Galena Mine when I took a field tour with Dr. Cheney and students from the University of Washington, and today I have seen some blue rock in the Star-Morning Mine. And close to the veins I could see, I am not sure, but I could see dark bands in the impure quartzites. Just like what happened in the Bunker Hill, I could not see them in the beginning. I had to train my eye and collect the samples and cut them and see all these beautiful soft sedimentary structures. Therefore, it is possible that the stratiform sulfides could be present everywhere in the district.

Grant Young, University of Western Ontario in London, Ontario: I would like to address Dr. Mumma. I know you said you did not want to comment on this, but I am going to press you anyway regarding the genesis of some of the sulfides. From the slides that you showed and from the way you interpreted the slides of polished hand samples and so on, it looks as if you have a very intimate association between these sulfides and normal clastic sediment set down in a highly active regime. Do you have any comment, and can you discuss how recrystallized sulfides got there? Do you think they were eustatically introduced, or do you think there was chemical precipitation with deposition? If the latter, it seems rather unlikely because of the high flow regimes suggested by some of your sedimentary structures.

Martin Mumma, Eastern Washington University: Actually I have written a section on genesis in my paper, and I did not think that they were like what Patterson and Razak said. They mention stromatolites in the Bel sediments, and Renfro’s article mentions that stromatolites are needed for the preservation of sulfides. I think it is not detrital; it is just precipitation.

Young: Well, I find that very hard to buy from the slides that you have shown, because it looks like the heavy minerals, the sulfides, are concentrated in foreset laminae of cross-beds.

Mumma: No, but these are reworked. These portions have already been subject to greenschist metamorphism, and sulfides do not tend to retain any original detrital characteristics whatsoever after recrystallization. You cannot equate sulfides with silicates. It might have been detrital, I do not know; or it might have the original characteristics obliterated, and now we do not see them, because they are all completely restituted.

Young: Are you saying that the sulfides were precipitated because of the metabolic activity of algae elsewhere and then were resedimented in the environment you find them in now?

Mumma: No, just before the consolidation of the sediments.
Young: I do not follow that. I noticed that in his first slide he had age determinations or a time spectrum on the side of the column and I note that the lower boundary of the Missoula Group was placed at 1,300 m.y. I wonder if you could defend that.

Mumma: The boundary there is put on the basis of some of the Purcell Silts that have been dated at around 1,100 m.y., and those mud cracks are most common above that boundary, so this is more or less a matter of interpolation. There is no specific, as far as I can determine, potassium-argon date or whatever kind of a radiometric date that would confirm that particular boundary.

Young: A second point was in regard to the things you talked about as burrows. I think that if your interpretation is correct, it is extremely exciting new information, but I would like to be convinced that they are in fact burrows. I do not think any of the evidence is totally convincing, but maybe you can expand on this.

Mumma: I would like to be totally convinced of that also, but the things that I have looked at seem to be mud cracks and do not follow the general pattern that I see in some of these things that we call encased burrows. A lot of mud cracks follow a regular cycle. There is a fairly definite spacing between the cracks, depending upon the size of the polygons and so on. They start and they stop at particular levels, and in general the mud cracks that I have seen, particularly in the St. Regis, tend to be rather straight sided and do not tend to be compressed and compacted as a lot of these strange things that I see and that for the lack of a better term, at this point, I call burrows. I am well aware of the literature dealing with the Precambrian, and in no case can I find definite proof that particular metazoans existed anywhere before about 850 or perhaps 900 million B.P. The Ediacaran Fauna is well documented; that you are probably very well familiar with, as being familiar with Australia confirms this and some other things also. So, the presence of metazoans, polychaetes, or anything like that is entirely conjectural, but if someone else can come up with a better explanation for the organization of these particular things, then I am certainly open to suggestions.

Ernest Gilmour, Eastern Washington University: I have a number of questions, but to start off with the junior author of the third paper, I would like to understand a little bit better what the general configuration is, say, from a basin point of view, of the Revett, and how does the fan-delta model explain the deposition of it. Two quite specifically different interpretations of the Revett-St. Regis interval have been presented.

Don Winston, University of Montana: Now, the interpretation that we developed is basically an analogy with the work that we have done on the Missoula Group for which we have stratigraphic information. Our interpretation is very close to the interpretation to the north. Now, horizontal laminations are not just formed on beaches. Horizontal laminations are processed by deposition characteristically in the upper flow regime, and upper flow regime can exist on beaches. The shallow water with the shallow waves brings the sediments up and carries them back down the beaches. Horizontal laminations are also found in river systems, particularly in braided streams where in very shallow water the flow regime produces horizontal lamination for the same reason. One of the critical things about identifying beaches versus horizontal lamination produced by rivers is in looking for the low-angle foreshore cross beds and the successions of low-angle foreshore cross beds, which Marty Mumma has shown in his slides. These we can see forming on modern beaches. They are also characteristic of the Cretaceous beaches that we can see stretching from Montana down into New Mexico. So, let me ask Marty if he has seen foreshore cross beds in the sands which are characteristic structures of beaches.

Mumma: Yes, certainly. There are all kinds of those things that you can find in practically any trench that you dig on beaches.

Gilmour: What is the scale of the foreshore cross beds? Do they stretch from the upper part of a thick quartzite unit down to the base of that same quartzite unit in the way they do in Cretaceous and in the way they do in modern reconstruction?

Winston: In some instances they do. They tend to be relatively large bundles of planar cross beds. Not exactly tangential at the end like you see in some of the beds that you are referring to, but the bundles are almost mirror images of many of foreshore cross beds that I have seen, for instance, in many of the beaches on Galvaston Island in North or South Carolina and along the Florida coast where there are coarse carbonates. There may not be really as much difference between the interpretation that you have and the interpretation that I have. It may actually be a matter of scale more than anything else. I have worked with the Missoula for some time, and I am in firm agreement with you. But it is probably a fan or a braided stream complex. The sediments that you see in the Revett bear no resemblance to the sediments that are in the Missoula. You do not have the salt casts, the small-scale veins, and all those kinds of things that are in the Missoula. I think that the sedimentary structures in the sands are similar. I have looked at some of the Revett around here and elsewhere, and I have never seen what I would consider to be a good foreshore cross bed. Let me ask Tom Bowden if he has seen good foreshore cross beds.
Tom Bouden, Anaconda Company: No, not that I know of.

Winston: Another possibility then is to develop this sort of stratification in the fan-delta system where the channels, perhaps as they come down and spread out, become too shallow to form good bars and dunes, transverse bars and dunes, but instead flow almost like a sheet wash and form horizontal stratification. Now there is a great difference between the two kinds of deltas that you showed in the slides. You showed a Mississippi delta, which has vegetated banks, and a meandering channel that develops very nice point bars as the channel shifts back and forth. On the other hand, the Corpus Christi fan as you showed was devoid of vegetation, and the channels of the Corpus Christi look very broad and poorly defined. There is another fan-delta, the Copano fan-delta that has been described, just up the coast, in a very similar situation, and the characteristic sedimentary structures are cross beds and horizontal lamination. The traction load deposits as cross beds and horizontal laminae, and the suspended load can pass out into the bays and lagoons as in the case of the Corpus Christi delta into Corpus Christi Bay. So, we envision then a detrital surface. It was shown several years ago that the river systems predate vegetative cover on the land surfaces; hence, they must have been different from the ones we see operating today. In other words, it is the vegetation that has produced soils and mostly suspended load streams versus what must have gone on in the geologic past. In the Precambrian with the rain drop splash, sheet wash runoff, there was very little retention of soils on the landscape. Most of the material was stripped and carried down the broad, braided streams as traction load. That is what, in fact, we see in the record.

Gilmour: I wanted to question Mumma after I developed a series of answers from Winston. I wanted to develop something about the geometry of the Revent through Winston’s paper and find out what sort of a progradational deposit they were talking about, so that I could go on to the barrier bar and ask a similar question about that. I am still not clear on the source area and the configuration of the basin, but maybe I can ask the question without it. I also am concerned about the time that is shown on the section. I am concerned also that the Revent-St. Regis package of sediments involved several hundred million years, so I want to know what type of a barrier bar are we talking about?

Mumma: And you want to know exactly what they have in mind, the barrier bar or what. Most of the barrier bars that I am familiar with and the things I have seen in Texas, as far as the literature is concerned, because I have not been there, is that those things are measured by thousands of years.

Gilmour: Now we are talking about a large barrier bar package which took literally millions and millions of years to develop, and I am just wondering what sort of framework people have in mind for this sort of a model?

Jerry Harbour, Gulf Resources Corporation: Did you do any comparative petrography between the blue rock and the nonblue rock, specifically in respect to the ankerite content?

Hauntz: Yes, and using thin section and mobile analysis I was looking specifically to see if there was an increase in galena with a decrease in ankerite. The problem I ran into was that the ankerite percentage in the blue rock is 1 to 4 percent and that the galena percentage is 1 to 2 percent. The statistical error in the mobile analysis method is 1 percent, and because of that I could not determine any direct relationship between an increase in carbonate, or decrease in carbonate and an increase in the galena.

Ian McCartney, Cominco Ltd.: You envision galena migrating from veins to sedimentary laminae in response to physical processes. I would like to know what characteristics such laminae might possess that would enable galena to concentrate there. I could see it happening more easily if galena was transported by chemical processes such as hydrothermal solutions. If it moved physically it would tend to migrate along and become concentrated in cataclastic features rather than in sedimentary laminae.

Hauntz: First of all let me say I have not been able to quantitatively establish a difference between laminae that contain galena and laminae that do not. I think there is a difference in sericite in that laminae. I think that the grain size, quartz grains, are larger in the laminae that contain the galena. Probably though, they would average 2 millimeter rather than 5 millimeter. There is no significant difference. I think you have the same problem whether you are using a hydrothermal mechanism or a mechanical mechanism. Why does the galena pick the particular place it picks to be precipitated or deposited?

Paul Barton, U. S. Geological Survey: I would like to ask you how these laminae terminate against unmineralized rock. Laterally through a bed that contains galena in these bedded laminae, there are some features that I cannot explain. It seems that when you have one of these beds, rather than the
galena being in the laminae all the way through the bed there is usually a 1- to 2-inch zone at the top of the galena-bearing bed that is barren or at least relatively barren in comparison to the blue rock. I cannot explain why. In thin section I see no difference, or at least no significant difference between the two places.

Lee Barker, Duval Corporation: Do you see any of the blue rock in drill holes in the hanging wall of the vein or do you not have enough information in that area?

Hauntz: To my knowledge, there was only one drill hole in the hanging wall that showed blue rock and that was very high in the mine. The access to that locality is no longer available, so I have not observed it personally. Blue rock does show up in bore holes to the north of the vein, and it looks very similar to what we see on the 4,250-foot level.

Barker: How do the lead-silver ratios vary from the vein and the blue rock, or do you know this?

Hauntz: The lead-silver ratios in the main vein of Lucky Friday are very variable. I do not have specifics; however, I think the ratio is probably from 1 to 1.5 to 1 percent lead to 500 ounces of silver in some places. In the blue rock, however, the lead to silver ratio is fairly constant. Again not a statistical average but a ball park figure is 1 to 1 ratio. That may vary but not significantly.

Charlie Jefferson, University of Western Ontario: Listening to the evidence presented for and against the stratiform original nature of these ores, it appears to me that if we are looking at this scientifically, we should set up various hypotheses and see ways in which we can disprove each one of them. The one that stands up is the one that stands closest to the truth.

Reid: We have essentially two hypotheses presented here: one is cataclasism with concurrent diffusion redistribution of galena from previously existing veins; the second is stratiform galena in the rocks redistributed into veins. We can look at several features which have been presented here that seem to me difficult to explain in terms of moving material from the veins into the sedimentary rocks. One of these is the feature that Ramalingaawamy presented showing leach zones beside the veins that cut these blue rocks, those with stratiform galena.

Jefferson: And how would you explain veins which are barren except for areas where they cut galena-rich layers in the quartzites?

Reid: By extrapolation from the small-scale features, I see some late redistribution of sulfide during the late part of the tectonic relaxation stage, driven by residual strain energy in the rock.

Ramalingaawamy: I would like to point out features that you consider cataclasitic in the slide that Charlie Hauntz showed. The sericitic laminae display schistosity, and the cataclastic schistosity and the mica schistosity are parallel. You might consider the form of the galena platelets to be simply another expression of schistosity; as you know, it is very plastic. If you take galena in the rock as it was among these layers and simply deform it, you produce a foliation along those layers. Galena being as soft as it is would enter into all of these laminae and between the sericite veins, just as you described in cataclasis.

Reid: I have thought about that and I have also looked at some higher temperature analogs and read about them, the high temperature sulfides of Broken Hill for example. Sulfides, which have recrystallized under a metamorphic environment, have developed essentially a granoblastic habit; and perhaps that is what we expect if the sulfide is there from the beginning—that is, the development of granoblastic habit not showing replacement features against other grains. Operating from that kind of premise, I prefer to think that the galena has been introduced late and has not been there from the beginning.

Juras: I am not going to ask a question of Reid. I want to make some statements about some of the fabric which I observed in Ramalingaawamy's work. It may not answer all the questions, but some of the ideas that are being considered need discussion. If galena has spread away from a vein and this sulfide appears stratiform, younger fractures which cut that stratiform may leach the lead from that stratiform. This does not indicate that the stratiform sulfides existed before the original veins—those large veins which occur away from those filled fractures which Mohan showed. The question arises whether those fractures which Mohan showed were related to the stress system of the folding. Maybe those fractures are not fold-related but are younger than the folding. The major premise of this comment is that the stratiform sulfides may have been distributed out of the veins and then the quartz-filled fractures, which are younger, leached the galena at the spot where the galena is along the bedding plane.

Jefferson: If you consider the simple hypothesis, you could say then, maybe the lead went into the sedimentary rock and then out again, and then in again and then out again. I still think that it is simpler to consider the one-step hypothesis where lead is concentrated at an early stage in the sedimentary rocks. The rocks are deformed and hydrothermally influenced to develop the observed structures and textures.

Juras: I like the simple hypothesis, but another statement that you made was that lead veins occur only in places where the lead is stratiform. That is incorrect. At Bunker Hill we have lead veins where there are no quartzite stratiform lead areas.
McGregor-Dawson: You talked about the galena; what is happening to the sphalerite? You mentioned some sphalerite in the sediments. What is the deformation of the sphalerite?

Juras: Sphalerite behaves in the same textural ways as galena. I should have said that; it is subordinate and I just relied on descriptions of galena to carry the topic.

McGregor-Dawson: You do not get any cataclastic deformation in the sphalerite?

Juras: As you recognize, sphalerite behaves texturally in the rock in the same ways as the galena does.

Wilfred Walker, Jet Propulsion Laboratory: Charlie Jefferson just wanted a simple way of doing things, and yet Dwight Juras has already pointed out that we have at least two periods of deformation. Presumably by what you say, Dr. Reid, metamorphism accompanied them. I see nothing wrong with both hypotheses working. Material coming from the rocks initially and being mobilized into veins, probably in D-1 or D-2, and then with cataclasis in D-3, being mobilized somewhat out of them. I do not think that we need just two hypotheses, as Charlie said. One or the other, I see every indication of having them both.

Foramin: Back in about, maybe 1,700 m.y. B.P., we had a worldwide intrusion of anorthosites along the edges of the ancient platform that formed triple junctions similar to those on the east coast of Africa. The dome was formed at the triple junction. The western triangle of the triple junction was Belt Island; the east-west arm was an aulacogen, and it deposited the later Revett sediments and so on. Along with that anorthosite intrusion, we had the earlier diabases which, in some fashion, tapped the mantle and got the metals up there, forming a Red Sea-type deposit, conceptually parallel to that at the Red Sea triple junction. Later the sediments were metamorphosed by first north-south folding and then west-northwest folding and the metals finally reposed in the west-northwest folding fabric. How's that for a theory?