BELT Symposium III Abstracts

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MBMG 381
Richard B. Berg, Compiler

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Cover: Stromatolites in the Altyn Formation
East of Apikuni Falls, Glacier National Park
Scale is in 5-cm divisions.

Photo by Robert J. Horodyski

B E L T

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Preface

Most of the papers presented at Belt Symposium III, held August 14-21, 1993, at Whitefish, Montana, are published in Montana Bureau of Mines and Geology Special Publication 112 entitled "Proceedings of Belt Symposium III." The twenty-two abstracts included in this open-file report are for papers for which a manuscript was not submitted for inclusion in the proceedings volume. The abstracts included here are compiled from the Program and Abstracts for the Symposium with no editorial changes.

January 11, 1999

Butte

Richard B. Berg

GEOLOGY OF THE MONTANORE STRATABOUND CU-AG DEPOSIT, LINCOLN AND SANDERS COUNTIES, MONTANA.

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A stratabound Cu-Ag deposit hosted in quartzite and siltite of the uppermost beds of the lower Revett Formation was discovered by U.S. Borax geologists in 1983. The outcrop is located next to Rock Lake in the southern Cabinet Mountains. The find partially resulted from a reappraisal of favorable horizons for Revett stratabound Cu-Ag mineralization and the release of U.S. Geological Survey Bulletin 1501 (Wells and others, 1981) indicating the possibility of significant mineralization within the Cabinet Mountains. Borax drilled 27 holes in the deposit over five drilling seasons, all penetrated rock of economic grade. The horizontal length of the deposit is over 11,000 ft, and the width varies from about 800 ft at the outcrop to over 4,000 ft at the northwest end. The property was purchased from Borax in the fall of 1988 by a Noranda Minerals/Montana Reserves joint venture, and renamed the Montanore Project. Noranda then initiated a decline for deposit access, and the permitting needed to allow production.

The geologic reserves of the deposit are estimated at over 200,000,000 tons. About 134,500,000 tons of ore at a grade of 0.74 percent Cu and 1.93 ounces per ton Ag are within the extralateral rights appurtenant to validated mining claims owned by Noranda and that cover the apex of the deposit. The remainder, located outside the extralateral rights, and under the Cabinet Mountains Wilderness, cannot be legally mined. The Montanore Project area, as discussed in this abstract, is shown in Figure 1.

Belt Supergroup strata in the project area consists of greenschist-grade metamorphosed sedimentary rocks assigned, in ascending order, to the Prichard, Burke, Revett, St. Regis and Wallace Formations. Glacial till fills valley bottoms. The decline, when completed, will transect the upper part of the Prichard, the Burke, and the lower Revett. Upper lower Revett through the Wallace are penetrated by drill holes as well as exposed on the surface. Figure 2 is a cross-section along the decline and through the deposit.

The Prichard Formation consists of predominately gray to black, thin-bedded to fissile siltite, minor medium-bedded silty quartzite, and rare quartzite. Pyrrhotite and sparse pyrite, are common in the finer-grained rocks as disseminations and as concentrations along bedding and fracture surfaces. Up to 1% arsenopyrite, which formed concurrently with the pyrrhotite (J.E. Clemson, pers. commun., 1991), is present in one 50-foot-thick zone. Multi-element analysis of samples from this zone for Au, Ag, Cu, Pb, Zn, As, Sb, Bi, Hg, and Mo found only high arsenic values. Biotite porphyroblasts are abundant in the pelitic rocks, and chlorite is notable in the silty quartzite and quartzite. Sedimentary features include small-scale scour and fill, soft sediment slumping and ripple marks. In several parts of the uppermost Prichard, thin, homogenized layers show no internal stratification and obliquely cross-cut bedding. These features may result from debris flows or sediment fluidization. The Prichard in the project area is tentatively assigned to the upper member and transition members as described by Cressman (1989). Thrust faulting may jumble the section and hinder the correlation.

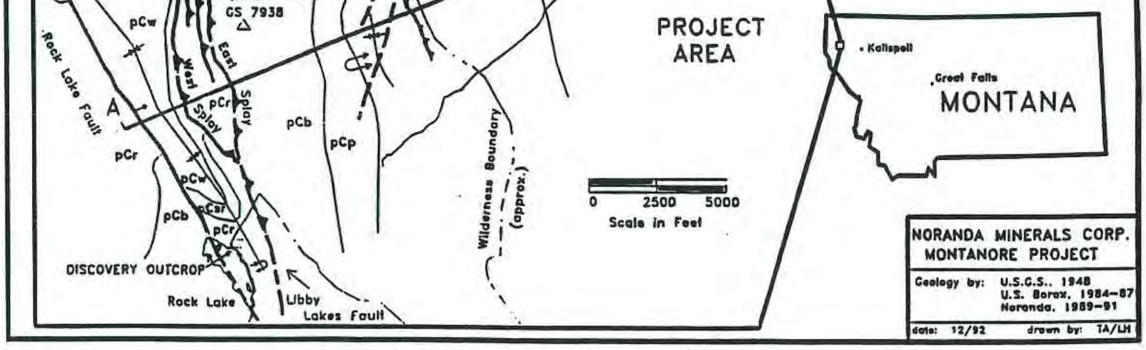

Fig.1 Location Map with Generalized Geology

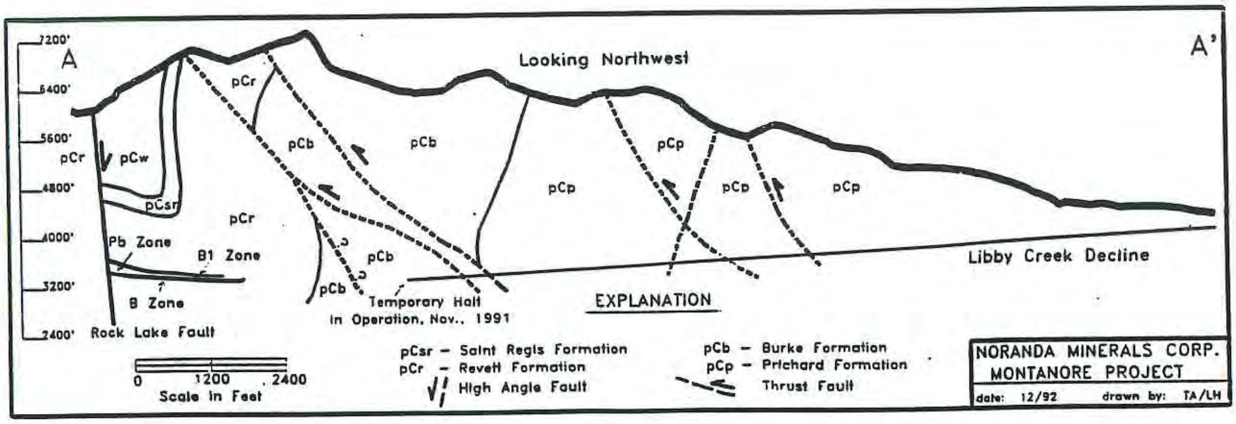

Fig. 2 Geologic Cross-Section Along Libby Creek Decline

The Burke Formation consists primarily of siltite, subordinate silty quartzite and sparse argillite. It has been divided into two informal members similar to those described by Mauk (1983). In the lower Burke, which is about 800 ft thick, siltite beds range from 3 in. to over 3 ft thick, and average about 1.5 ft thick. The siltite beds often grade up into thin argillite caps. The upper Burke is over 3,000 ft thick and consists of two broad zones of interbedded silty quartzite and siltite with an intervening siltite unit. Magnetite is common in the upper Burke. The top of the Burke is mostly truncated by thrust-faulting.

All three members of the Revett Formation are present, and combined, are 2,500 ft thick. The upper portion of the lower member, along with the other two members is exposed on the surface and in drill holes. The base of the Revett is exposed adjacent to the project area. Beds in the upper part of the lower Revett are grouped into alternating sequences of quartzite-rich units and siltite-rich units. These sequences, each called a bed, are given alphabetical letter designators down section from the lower Revett/middle Revett contact. The labeling method is based on one devised by ASARCO geologists and reported by Hayes (1983). Beds A through the top of F have been drilled. Thicknesses range from 210 ft for an A bed intercept to less than 30 ft for an E bed intercept. The thickest stratigraphic interval of the upper part of the lower Revett penetrated by drilling is 459 ft.

The upper two members of the Revett are composed of contrasting lithologies. The middle member is a sequence of siltite, silty quartzite, and locally quartzite that averages about 600 ft thick. The upper member consists of thick-bedded quartzite and interbedded 0.5 to 10-foot-thick siltite beds. It averages about 215 ft thick.

The St. Regis and Wallace Formations, because they lack economic significance in the project area, have received scant attention.

Pelitic sediments in the Burke and Prichard Formations appear to have a higher sand and silt content than is common in other places in northwestern Montana (Mauk, 1983; Cressman, 1989). A grain-size increase in the Cabinet Mountains area could be the result of sediment deposition closer to a source area than elsewhere.

Structurally, the area is marked by both faults and folds (Figures 1 and 2). The major faults, high and low angle reverse structures, are related to west-directed thrusting. Only the largest, the Libby Lakes fault splays, are visible on the surface. However, decline mapping detects numerous reverse faults with major to minor offset. Another large fault, the Rock Lake fault, is a steeply-dipping, west side up, normal structure located on the western edge of the area (Figures 1 and 2). It is the most important fault as it is the western boundary of the Montanore mineralization. Based on stratigraphic evidence, the Rock Lake fault has about 2,500 ft of dip-slip movement. Fillipone and Yin (1991) indicate that the movement on the Rock Lake fault postdates the west-directed thrusting. The Snowshoe fault, a major structure along the eastern side of the Cabinet Mountains north of the project area, attenuates southward into a minor fault along an anticline hinge.

The most significant fold is the Rock Lake syncline. It is a plunging, asymmetrical fold overturned to the west and open to the north. The eastern limb is steeply overturned, and truncated on top by the Libby Lakes fault splays. The western, lower limb, which hosts the deposit, is fairly flat and terminates on the west against the Rock Lake fault. To the east of the Rock Lake syncline, in the upper Libby Creek and Ramsey Creek drainages, the rocks are distorted into broad anticlines and synclines. All the folds appear to have formed in response to west-directed thrust faulting.

The geology presented in this abstract differs somewhat from the geology as presented by Wells and others, (1981). The difference is due primarily to mapping on a larger scale (1:4800), and the recognition from decline mapping of the influence exerted by west-directed thrusting.

The deposit is hosted in the upper beds of the lower Revett Formation. Zoned Cu-Ag-Pb sulfide-bearing minerals occur as discrete, intergranular disseminations in a tabular layer. The layer consists of multiple, stratabound Cu-Ag-Pb zones. Total thickness of the entire, mineralized, tabular layer locally exceeds 460 ft, but commonly ranges in thickness from 250 to 325 ft. Two economic zones within this layer, the B zone and the smaller stratigraphically higher B1 zone, have been partially delineated and are bed-transgressive. These zones are separated by a continuous Pb zone. Several other poorly defined Cu-Ag zones, some with vertically adjacent Pb zones, have been detected up and down section from the B and B1 zones. True thickness of the B zone within the area of ore reserves as defined by mining parameters ranges from 13 to 60 ft and averages about 31 ft. The B1 zone true thickness ranges from 14 to 38 ft and averages about 30 ft. Mining limits include a room and pillar mining method using a 14 ft minimum vertical mining height, a 67 percent recovery rate, and a $(US)9.91 Net Smelter Return cut-off grade based on metal prices of $(US)5.00 ounces per ton silver and $(US)0.93 per pound of copper. The medial, uneconomic Pb zone ranges in thickness from 30 ft along the northeastern edge of the deposit to over 300 ft next to the Rock Lake fault. Mineralized rock is bounded by the Rock Lake fault on the southwest and the outcrop on the southeast. Its limits to the northeast and to the northwest are not determined.

Minerals in the ore consists of bornite, chalcocite, chalcopyrite, native silver, digenite, covellite, galena, pyrite and pyrrhotite. They formed primarily as pore-space filling and grain coatings and also as minor, secondary fracture-fillings. Sulfide concentrations along bedding planes and as heavy-mineral laminations are common. The sulfide grain size is directly proportional to the grain size of the host sediments. Bornite, chalcocite, and native silver are most abundant in the higher grade zones. In lower grade zones, chalcopyrite and pyrite concentrations increase, and bornite and chalcocite amounts decrease. Ore minerals are often accompanied by coarse grained quartz and/or manganoan ankerite (McNeil, A., unpublished company report, 1989). With few exceptions, galena does not co-exist with any economic Cu-Ag mineralization. Minerals associated with galena are chalcopyrite, pyrite and pyrrhotite.

Origin of the mineralization is unknown. In the Spar Lake model of Hayes and Einaudi (1986), and Hayes and others (1989), which is interpreted from study of the similar Spar Lake deposit located approximately 17 miles to the northwest, diagenetic ore solutions moved through permeable quartzite sand bodies, and metal deposition was localized by preore reactants This hypothesis might explain the Montanore deposit. Alternatively, Wodzicki (pers. commun., 1990) advances the possibility that reducing, sulfur-bearing brines from the underlying Prichard Formation may have ascended the Rock Lake fault and flowed into the Revett Formation. These brines then mixed with saline, oxidizing copper-silver rich fluids migrating through the Revett towards the margin of the Belt basin. This commingling of the two fluids in favorable beds proximal to the fault could result in the precipitation of metal sulfides.

Evaluation of the deposit by underground drilling should begin by early 1994. By late 1994, our understanding of the deposit should be significantly greater.

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GEOLOGY OF THE ROCK CREEK DEPOSIT, SANDERS COUNTY, MONTANA

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INTRODUCTION

The Rock Creek deposit is located about 16 miles southeast of ASARCO's Troy Mine, in the southwest portion of the Cabinet Mountains. The deposit was discovered in 1965 by geologists for Bear Creek Mining Company, who found mineralized outcrops near the south end of the deposit. In 1972 ASARCO acquired the deposit as part of the Spar Lake (Troy Mine) agreement. Since 1972, 121 diamond drill holes totalling 122,605 ft have been drilled, 103 of which utilized helicopter support. The Rock Creek project is currently undergoing environmental-impact studies and development (Balla, 1992).

The Rock Creek deposit is a stratabound-stratiform copper-silver deposit similar to the Troy deposit. The Rock Creek orebody is about 12,000 ft long, 6,000 ft wide, and averages 27 ft thick. Ore reserves are 143,760,000 tons with an average grade of 0.68% copper and 1.65 troy ounce per ton silver. The orebody is on average approximately 1,000 ft beneath the surface.

GEOLOGIC SETTING

The regional geology of the Rock Creek area has been described by Wells and others (1981). The deposit is located in the upper units of the lower Revett Formation, the oldest exposed formation in the area. The base of the Revett is not exposed in the project area, as it is covered by talus. Local detailed stratigraphy has been defined by ASARCO from drill holes. Progressing downward from the lower Revett-middle Revett contact, the lower Revett is subdivided into the following sub-units as shown below:

Unit	Predominate Rock Type	Thickness Range (ft)	Average Thickness (ft)
A	Quartzite	60-150	110
B	Siltite	29-138	85
C	Quartzite	40-130	90
D	Siltite	30- 95	70
E	Quartzite	10-130	?

The middle Revett varies from 150 ft to 425 ft in thickness, and averages approximately 300 ft. It consists of argillite, siltite and minor silty quartzite.

The upper Revett consists of alternating couplet beds of quartzite and siltite. The thickness of the upper Revett varies between 190 ft and 307 ft, and averages 300 ft.

The St. Regis Formation overlies the Revett Formation, crops out throughout the project area, and consists of green argillite, siltite and minor silty quartzite beds. The formation is approximately 550-600 ft thick.

The Wallace Formation crops out in only two small areas, the top of St. Paul Peak where the lowermost 75 ft of the Wallace forms a cap on the mountain, and in the area west of Moran Basin where the lowermost Wallace Formation is exposed.

The Rock Creek deposit is cut by two northwest-trending faults, the Copper Lake fault and the Moran fault, which divide the deposit into three structural blocks: the largest, the Chicago block, which contains about 57% of the ore reserves; the St. Paul block which contains about 35% of the ore reserves; and the North Basin block, which contains the remaining 8% of the ore reserves.

The Chicago block is a broad open anticline, the axis of which plunges very gently to the north; the anticline is truncated on the north by the Copper Lake fault and by erosion on the south side. The Copper Lake fault trends approximately N 40° W across the deposit. Displacement on the fault is variable, with the north side downdropped 200-500 ft. Displacement along the fault increases from north to south.

The St. Paul block is bounded on the south by the Copper Lake fault and on the north by the parallel Moran fault. The St. Paul block has been downdropped approximately 200-500 ft relative to the Chicago block. Based upon variations in thickness of individual stratigraphic units, there has been considerable strike-slip movement along both the Copper Lake and Moran faults. The distance across the graben is approximately 3,200 ft.

The North Basin block is the smallest of the three ore blocks. Erosion on the north and east sides has substantially reduced the size of the block.

MINERALIZATION

Stratabound-stratiform copper-silver mineralization of ore grade outcrops in two areas: the area south of Chicago Peak and St. Paul Peak, and the northern area, north of St. Paul Peak and east and north of Moran Basin. The stream sediment dispersion pattern from the outcrops has been described by Cazes (1981) and Cazes and others (1981).

The copper-silver mineralization at Rock Creek is similar to the mineralization at the Troy Mine which has been described by Hayes (1983), Hayes and Einaudi (1986), Hayes and Balla (1986), and Hayes and others (1989). The central bornite-chalcocite (digenite?) zone represents the ore zone. Sulfide minerals are disseminated in the interstitial spaces between sand grains in the quartzite. The bornite ore zone is surrounded by successive sulfide halo zones that contain chalcopyrite ± pyrite, galena ± sphalerite ± pyrite, and finally pyrite ± pyrrhotite. Silver is within the bornite zone.

At Rock Creek, sulfide minerals are at different stratigraphic positions within each of the three structural blocks. Within the Chicago block the mineralization is terminated on the north by the Copper Lake fault, and on the east by erosion. To the west, the limits of the mineralization have not yet been found, but decreases in grade; the grade also decreases to the south, before being cut out by erosion.

In the southern and western portions of the Chicago block the bornite-bearing ore zone occurs within the C, D, and E beds of the lower Revett. Northward and eastward, the bornite zone transgresses upward across strata until it occurs in the A, B, and C beds of the lower Revett adjacent to the Copper Lake fault. The thickest and highest grade copper-silver mineralization found as of 1993 is adjacent to the fault. Within this zone are multiple, stacked ore-grade deposits which are richer than the average grade of the Rock Creek deposit.

The St. Paul block represents the central northwest-trending graben between the Copper Lake fault and the Moran fault. Within the graben the Revett Formation has the form of a northwest-plunging anticline whose axis is parallel to the bounding faults.

The St. Paul block overall contains the highest grade of the three blocks. Within the block the bornite zone is also discordant to bedding. From the northwest, where it is in the D and E beds, the zone transgresses at a low angle across the strata to the southeast, where it occurs in the A unit beds. As the bornite ore zone approaches the Copper Lake fault it develops into multiple, stacked ore zones.

The North Basin block is bounded on the north and east by outcrop, and on the west and south by the northwest-trending Moran fault. Within the block the bornite ore zone occurs within the A and B units of the lower Revett. The sulfide halo surrounding the bornite ore zone rises stratigraphically from north to south, and the ore grade in drill holes increases from north to south.

ACKNOWLEDGEMENTS

The author gratefully acknowledges the excellent work of L. M. Appelgate on the Rock Creek deposit. He also thanks E. P. Bayley, Jr., and the ASARCO personnel who have worked on the project and advanced our knowledge of the project.

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STRUCTURE OF THE LEWIS THRUST PLATE BEHIND THE WATERTON-GLACIER SALIENT, MONTANA, SOUTHERN ALBERTA AND BRITISH COLUMBIA

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The Waterton-Glacier salient of the Lewis thrust fault (Fig. 1) is a major feature of the Lewis thrust system, and its existence and configuration may be partially the result of the presence of transverse basement features. Data from oil wells drilled in southeastern British Columbia provide excellent data for interpreting the basic structure of the Lewis thrust plate behind the Waterton-Glacier salient (Boberg, 1984, 1986).

The Shell MacDonald well (#1 on Fig. 1 and Fig. 2) was drilled to a total depth of 3,809 m (12,497 ft) through a normal stratigraphic sequence on the MacDonald dome in southeastern British Columbia. Shell MacDonald started at the surface in the Mississippian Mount Head Formation (Madison equivalent) and terminated in the Grinnell Formation of the Proterozoic Belt-Purcell Supergroup.

In the Sage Creek area, 15 km (10 mi) to the east of Shell MacDonald, Shell Sage Creek 2 (#2 on Fig. 1 and Fig. 2) drilled through the Flathead Fault into lower Proterozoic Belt-Purcell rocks, Waterton Formation equivalent or older; these strata were described by Fermor and Price (1983) as older than any other Belt-Purcell rocks exposed in the Clarke Range. The Tombstone thrust was penetrated at a depth of 1,392 m (4,567 ft) below which is the Tombstone thrust - Lewis thrust duplex zone, made up totally of lower Proterozoic, Waterton, or older, metasedimentary rocks. Below the Lewis thrust fault, at 1,675 m (5,373 ft), Shell Sage Creek 2 penetrated a total of 5 imbricate sub-Lewis thrust sheets which are part of a complex stack of thrust sheets formed between splays of the Lewis thrust fault as shown on Fig. 2. These multiple, sub-Lewis, imbricate thrust sheets have stacked up below the Lewis plate near the crest of the major Flathead ramp.

Stratigraphic and structural data from Shell MacDonald and from the Sage Creek area, indicate that the plane of the Lewis thrust fault is at least 2,740 m (9,000 ft) below the bottom of Shell MacDonald, assuming that the Lewis thrust fault below MacDonald dome is carried at the same stratigraphic horizon as at Sage Creek (Boberg 1984). This interpretation puts the plane of the Lewis thrust fault at a depth of about 6,550 m (21,500 ft) or a subsea elevation of -5,120 m (-16,800 ft) below MacDonald dome. Additionally, these data suggest that the Lewis thrust fault may have had but a single major ramp, the Flathead ramp, where the Lewis thrust fault has cut up the stratigraphic section from the basal detachment in the lower Belt-Purcell Supergroup to the top of the Mesozoic section, more than 9,000 m (30,000 ft) vertically in a horizontal distance of less than 15 km (10 mi). The change in character of the Lewis thrust fault from a bedding plane thrust with minimal stratigraphic throw to the west to a high angle thrust cutting more than 9,000 m (30,000 ft) upsection, forming the Flathead ramp, may have had its inception in a basement ridge or other structure which deflected the bedding plane thrust upsection. It is inferred that the basal décollement is essentially the contact between the Proterozoic Belt-Purcell Supergroup and the Archean basement and that the Lewis thrust fault has been deflected upsection from the basal décollement somewhere between the Rocky Mountain trench and the MacDonald Range.

The indications of all of the available data are that the Waterton-Glacier salient is the result of deep-seated detachment from the basal décollement of the plane of the Lewis thrust fault behind the salient. The pre-erosional thickness of the Lewis plate in the Waterton-Glacier Park region was probably in

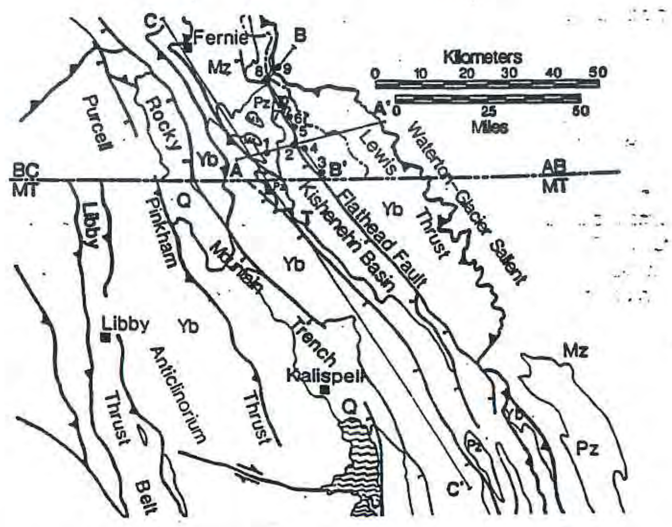

Figure 1. Index map of northwest Montana, southern Alberta and British Columbia showing the location of the Waterton-Glacier salient of the Lewis thrust fault and sections A-A', B-B'and C'C'. Q - Quaternary valley fill, T - Tertiary sedimentary rocks, Mz - Mesozoic rocks, Pz - Paleozoic Rocks and Yb - Belt-Purcell Supergroup rocks. Wells mentioned in text and on sections are shown with dark circles (•); 1, Shell MacDonald (BC, b-30-H/82-G-2); 2, Shell Sage Creek 2 (BC, c-55-E/82-G-1); 3, Shell Kishenena (BC, d-56-C/82-G-1); 4, Pacific Atlantic Flathead (BC, d-34-E/82-G-1); 5, Shell Flathead 3 (BC, b-35-L/82-G-1); 6, Shell Flathead 4 (BC, d-58-L/82-G-1); 7, Shell Flathead 2 (BC, d-12-A/82-G-7); 8, Shell North Kootenay Pass 2 (BC, b-58-H/82-G-7); 9, Shell North Kootenay Pass 1 (AB, 4-23-5-5). Well locations are given as standard NTS location format for British Columbia (BC) and as standard township and range legal subdivision format for Alberta (AB).

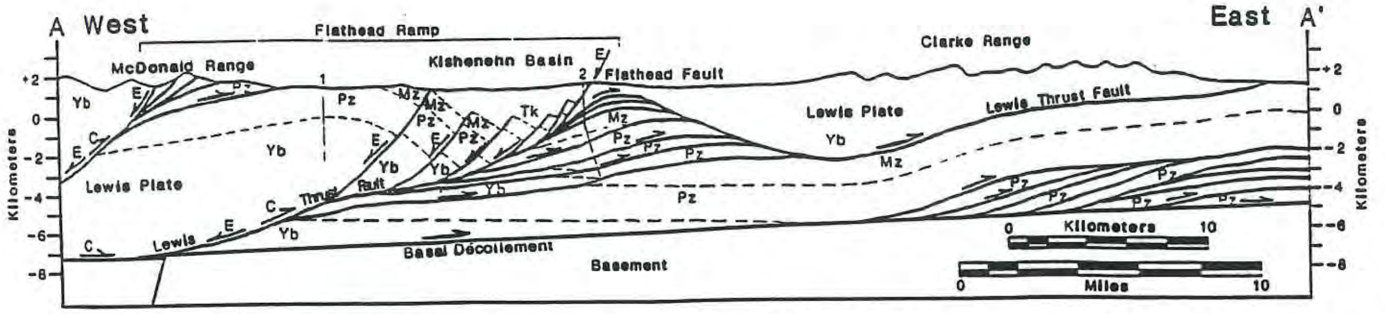

Figure 2. Structure section A-A' (see Fig. 1 for location and wells used in section). Tk - Tertiary Kishenehn Formation, Mz - Mesozoic rocks undivided, Pz - Paleozoic rocks undivided, Yb - Proterozoic Belt-Purcell Supergroup undivided. Letters "C" and "E" on fault directional arrows refer to movement on fault during compression and extension, respectively.

excess of 9,000 m (30,000 ft). Post thrusting extension during Tertiary time resulted in listric normal faulting (the Flathead Fault) back onto the plane of the Lewis thrust fault and the deposition of the Oligocene Kishenehn Formation in the Kishenehn Basin west of the fault. Section A-A' (Fig. 2) suggests that the throw on the Flathead fault is at least 6,000 m (20,000 ft) comparable to the minimum estimate of Constenius (1988) of 6,400 m (21,000 ft).

The sharp westward deflections in the trace of the Lewis thrust fault at the northern and southern ends of the Waterton-Glacier salient cut sharply upsection creating an apparent "tear fault" relationship, appearing to actually offset the Lewis thrust fault at the southern end. Constenius (1982, 1988) refers to the Lewis thrust fault as a "'scoop-shaped' thrust that systematically cuts downsection inward from the frontal and lateral plate margins". At the north end, Price (1965) reports that the westward deflection is the result of a transverse northwest facing monocline, and the plane of the thrust abruptly changes stratigraphic position from the Altyn Formation (lower Belt-Purcell) to the Siyeh Formation (Helena equivalent) a stratigraphic throw of about 1,000 m (3,000 ft). At the south end of the salient Mudge and Earhart (1980) report a similar westward deflection of the trace of the Lewis thrust fault and Kulik (1982) describes the fault as cutting up-section more than 1,800 m (6,000 ft). The presence of westerly trending folds and northeasterly trending normal and tear faults south of the Waterton-Glacier salient were described by Mudge and Earhart (1980) as possibly reflecting the northeasterly trend of a pre-thrust structure that may have influenced the configuration Lewis thrust fault.

The nature of the Flathead Fault also changes character both to the north and the south. Behind the Waterton-Glacier salient the Flathead Fault is a listric normal fault. To the north the Flathead Fault changes character to an apparent left-lateral tear fault along a westward deflection in the trace of the fault, roughly paralleling the westward deflection of the Lewis thrust fault at the northern end of the salient. To the south the Flathead Fault splays into additional faults but does not turn and change character to a tear fault as it does to the north. These relations may suggest that the apparent transverse basement feature to the south is of less magnitude or is less abrupt than that to the north.

Most sections illustrating the structure of the Lewis thrust fault are dip sections, parallel to the movement on the thrust, such as Fig. 2, section A-A'. These sections commonly show the basement contact as a planar feature with a gentle westerly dip. The plane of the thrust is also commonly shown as a planar feature with a gentle westerly dip except where the thrust ramps or is folded. Alternatively the basement very likely had some relief and that relief may have had some effect on the depth and stratigraphic level of thrusting.

Section B-B' (Fig. 3) is a strike section through eight oil exploration wells drilled in southeastern British Columbia. This section demonstrates the relief on the plane of the Lewis thrust fault above and near the crest of the Flathead ramp. The Lewis thrust fault climbs in elevation to the north to the Haig Brook and Cate Creek windows where the thrust is exposed at the surface. It then drops in elevation to the north while cutting up the stratigraphic section. In the 50 km (30 mi) of this section there is over 1,800 m (5,900 ft) of relief on the plane of the Lewis thrust fault. A comparable amount of relief probably occurs on the plane of the thrust west of the Flathead ramp where the thrust is at great depth and is most likely still part of the basal décollement at, or near, the base of the Belt-Purcell Supergroup (very likely the Proterozoic-Archean basement contact).

On the diagrammatic regional section, C-C' (Fig. 3) the Lewis thrust fault is shown as essentially a bedding plane thrust which is affected both to the north and the south by transverse structural features in the basement, although the magnitude of these features may be less than shown. It is inferred that the plane of section C-C' is only slightly east of the area where the Lewis thrust fault has splayed from the basal décollement and begun to cut upsection. The Lewis thrust fault appears higher in the section south of the southern transverse

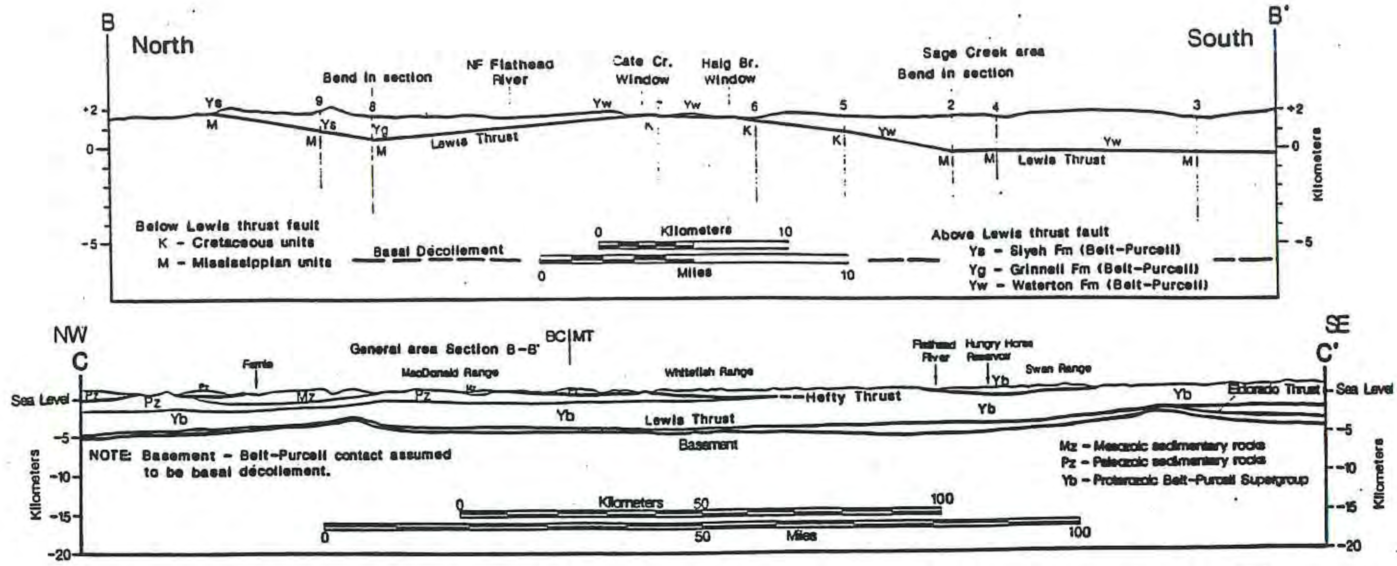

Figure 3. Structure section B-B' and diagrammatic structure section C-C' (see Fig. 1 for section locations and wells used in sections).

feature, and the Eldorado thrust has developed below it. North of the northern transverse feature the Lewis thrust fault appears to more closely follow the basement contact.

The northern transverse feature is shown as having a steeper southern flank which could possibly be the explanation for the abrupt change in the nature of the Flathead listric normal fault to an apparent left-lateral tear fault parallelling the westward deflection of the trace of the Lewis thrust fault. The southern transverse feature is inferred to have a relatively gentle northern flank possibly explaining the splitting of the Flathead Fault rather than turning into a lateral fault as it does to the north.

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OVERVIEW OF BELT GEOCHRONOLOGY AND PROBLEMS OF DATING THE BELT-PURCELL SUPERGROUP

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Despite new studies in the past decade, it remains difficult to derive a chronology of the Belt which accommodates all data. K-Ar, Rb-Sr, U-Pb, Pb-Pb and Sm-Nd systems can all show varying loss of radiogenic daughter product and/or inheritance from older crustal sources. Loss of the radiogenic isotope is most common in pre-Beltian rocks. Deformation and metamorphism of the basement rocks, in response to the Mesozoic Cordilleran orogeny, increase in intensity from east to west. To this effect must be added middle and late Proterozoic thermal events, and Tertiary crustal extension, which reset isotopic systems.

Along the east side of the Belt basin, K-Ar biotite ages of basement rocks fall between 1700 and 1750 Ma (figure 1, table 1). This range dates the closure of biotite to loss of argon following the Hudsonian arogeny. The temperature of closure was 300-350oC, the probable depth five to ten kilometers. Removal of this cover by regional uplift may have taken >100 million years. A shift of regional isostasy from positive to negative was necessary for the trangression of the Belt sea and basal clastic sedimentation. The maximum age of this basal sequence is probably no more than 1500 Ma. Older U-Pb ages of zircons from tuffs in the Yellowjacket Formation probably reflect inherited lead (K.V. Evans, written communication, 1992).

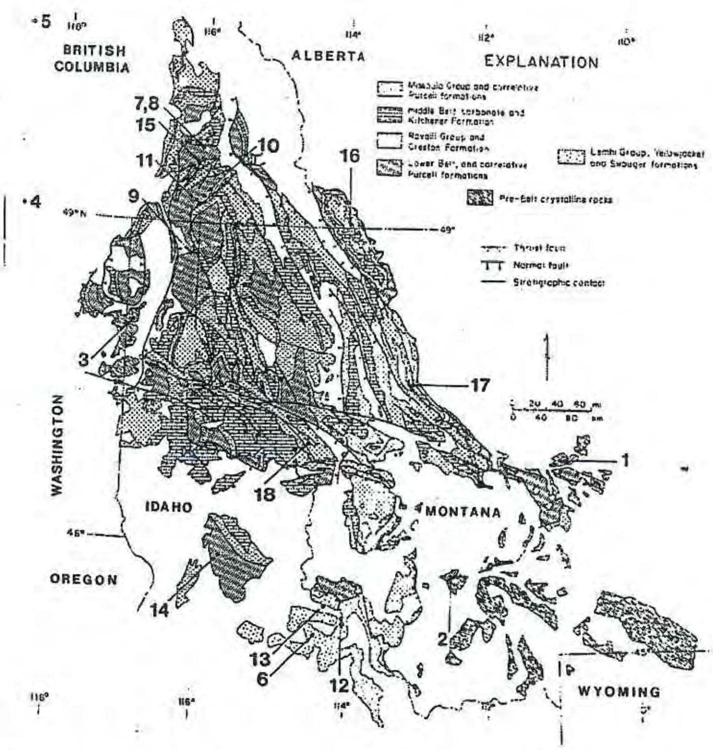

Figure 1. Location of samples critical to dating rocks of the Belt Supergroup. Map after Winston (1990). See table 1 for dates and references.

Other dateable rocks, the gabbro sills, have possibly assimilated some of the sedimentary host. This is especially likely if the sediments were clastic, water-saturated, and at shallow (~1 km) depths. Excess radiogenic argon in hornblende in the Lumberton sill (Hoy, 1989) is an indicator of this effect. Burwash and Wagner (1989) noted that central portions of the sill give higher Nd model ages than do the margins. Single-grain U-Pb Zircon dates of 1425-1710 Ma (table 1) from the granophyric phase of the Cooper Lake Sill led Ross and others (1992) to conclude the grains might have contained inherited Pb. By contrast, their single-grain U-Pb (sphene) analyses cluster on Concordia at 1350 Ma. D.W. Davis (oral communication, 1993) used concordant, single-grain zircon U-Pb analyses to date a gabbro sill near Kimberley at 1468±1.5 Ma.

In east-central Idaho and at Hellroaring Creek, B.C., a number of plutons of porphyritic granite are dated at approximately 370 Ma (figure 1, table 1, nos. 12-15). Host rocks of Yellowiacket or Aldridge Formation were metamorphosed to greenschist facies prior to intrusion (Evans and Zartman, 1990; Chamberlain and Doughty, 1993).

Reliable dates are lacking for Purcell lavas above the middle Belt carbonate. A sill in the Siyeh Formation at Lake Alderson, 500 m (1640 ft) below the lavas, gives a Rb-Sr isochron age of 1307±43 Ma. A metabentonite at the top of the Helena Formation contain zircons with inherited cores. Obradavich and others (1984) interpret the 1345 Ma U-Pb age to be maximum for middle Belt carbonate.

Upper Belt Rb-Sr ages for glauconite separates and wholerock isachrons range from 1170 to 900 Ma (Obradovich and others,1984). If valid, these numbers extend the range of Belt sedimentation 300 million years beyond other estimates. In view of widespread resetting of all other isotopic systems, these Rb-Sr values may also indicate a later event. Anderson and Parrish (1993) cite concordant U-Pb (sphene) dates of 1030 Ma from the Moyie sill as evidence of a Grenville-age metamorphic event in western North America. The history of the Belt-Purcell basin, as inferred from table 1 and figure 2, follows:

~1700 Ma- stabilization of crystalline basement;

~1500 Ma- transgressive sedimentation initiated;

1480 to 1430 Ma - extensional tectonics, basaltic sills, clastic sedimentation, stratiform ore bodies;

1370 Ma- anorogenic granitic magmatism;

>1310 Ma- widespread carbonate sedimentation;

~1310 Ma- basaltic volcanism with associated sills;

~1250 Ma- end of Belt sedimentation;

<1200 Ma- folding and regional metamorphism.

TABLE 1. Radiometric dates from the Belt-Purcell Basin and adjacent basement. B, biotite; H, hornblende; Z, zircon; WR, whole rock; tDM, depleted mantle model age; TCHUR, CHUR model age. [See Figure 1 for locations]

AREA	DATED UNITS/DATES	REFERENCE
PRE-BELTIAN		
1.Little Belt Mountains Little Rocky Mountains Western Canada Basin	Crystalline basement. K-Ar(B) isotopic closure during uplift, 1700-1750Ma; prior to deposition of basal Belt Neihart quartzite.	Billetti, 1969 Burwash and others, 1962
2.Highland Mountains, Montana	LaHood Formation down-faulted against Archean basement along a reactivated fault. K-Ar(B)~1700Ma.	Giletti, 1969
3.Priest River, Idaho	Augen gneiss in complex structure is probably pre-Belt basement. U-Pb(Z), mod. discordant, 1576±13Ma.	Evans and Fisher, 1986
4.Grand Forks, British Columbia	Gneiss horst, inferred autochthonous North American craton. Rb-Sr(WR) 1.70±0.4 Ba, Sm-Nd tDM 1.83±0.3 Ga.	Armstrong, and others, 1991
5.Thor Odin, British Columbia	Gneiss dome, inferred autochthonous North American craton. U-Pb(Z) 1934±6Ma.	Parkinson, 1991
LOWER BELTIAN		
6.Blackbird Mine, Idaho	Yellowjacket Formation. Mafic tuff with associated dikes. U-Pb(Z) 1670-1700 Ma. inherited Pb.	Hahn and Hughes, 1984 K.V. Evans (written comm., 1992)
7.Kimberley, B.C.	Gabbro sills. U-Pb(Z) 1468±1.5 Ma	D.W. Davis, (oral comm., 1993)
8.Sullivan Mine, British Columbia	Galenas from strataform deposits.	Godwin and Sinclair, 1982
	Pb-Pb age 1450-1500Ma using shale-hosted Pb growth curve.	Campbell and Ethier, 1983 1993)
	Pyrrhotites and galenas ~1350 Ma on Steacy and Kramer's curve.	Cumming (written comm., 1993)
	Chloritic footwall U-Pb(sphene) 1320±12 Ma.	Schandl and others, 1993
9.Crossport, Idaho	Granophyric differentiate of gabbro sill. U-Pb(Z) concordant 1433±10 Ma. Rb-Sr(WR) 12E35±165Ma.	Zartman, and others, 1982

10. Moyie Lake, British Columbia	Lumberton sill. Coarse-grained gabbro. U-Pb(Z) discordant but plot as linear array, 1445±10Ma.	Hoy, 1989
	Sm-Nd. TCHUR model ages: 1.43, 1.440 1.45, 1.50r 1.569 1.72 Ga. Three samples show crustal assimilation.	Burwash and Wagner, 1989
11. Cooper Lake, British Columbia	Granophyre from 90m thick sill. U-Pb(Z), single grains. 207Pb/ 206Pb, 1425-1457Ma (n=7); 1709Ma (n=1). U-Pb (sphene) 207Pb/206Pb, 1330-1412Ma (n=5).	Ross, and others, 1992
LOWER BELTIAN (?)		
12. Salmon River, Idaho	Rapikivi granite intrudes Yellow-jacket Fm. U-Pb(Z) 1365 Ma.	Chamberlain and Doughty, 1993
13. Shoup, Idaho	Foliated granite intruding Yellow-jacket Fm. U-Pb(Z) 1364±9 Ma.	Evans and Zartman, 1990
14. Red River, Idaho	Augen gneiss. U-Pb (Z) 1370 Ma, with inherited Pb.	Evans and Fischer, 1966
15. Hellroaring Creek, British Columbia	Pegmatitic muscovite-tourmaline granite; intrudes Moyie sill. U-Pb(Z) 1365±3 Ma.	J.Mortensen (oral comm., 1993)
MIDDLE BELTIAN (?)		
16. Lake Alderson, Albeta	Hornblende gabbro sill in Siyeh Formation 500m below Purcell lavas. Rb-Sr (9 WR, 2 apatite) 1307±43 Ma, I.R. 0.70658t0.00011, MSWD 1.26.	Burwash, in press
17. Sun River, Montana	Metabentonite at top of Middle Belt carbonate U-Pb(Z), $^{207}Pb/^{206}Pb$ 1345 Ma; zircons may have inherited cores.	Obradovich, and others, 1984
UPPER BELTIAN		
17. Sun River, Montana(2)	Shepard Formation; Rb/Sr(glauconite) 1170±20Ma. McNamara Formation; Rb/Sr (glauconite)1165t20Ma.	Obradovich, and others, 1984
18. Alberton, Montana	Pilcher and Garnet Range Formations, Rb/Sr(WR) ~900Ma.	Obradovich, and others, 1984.

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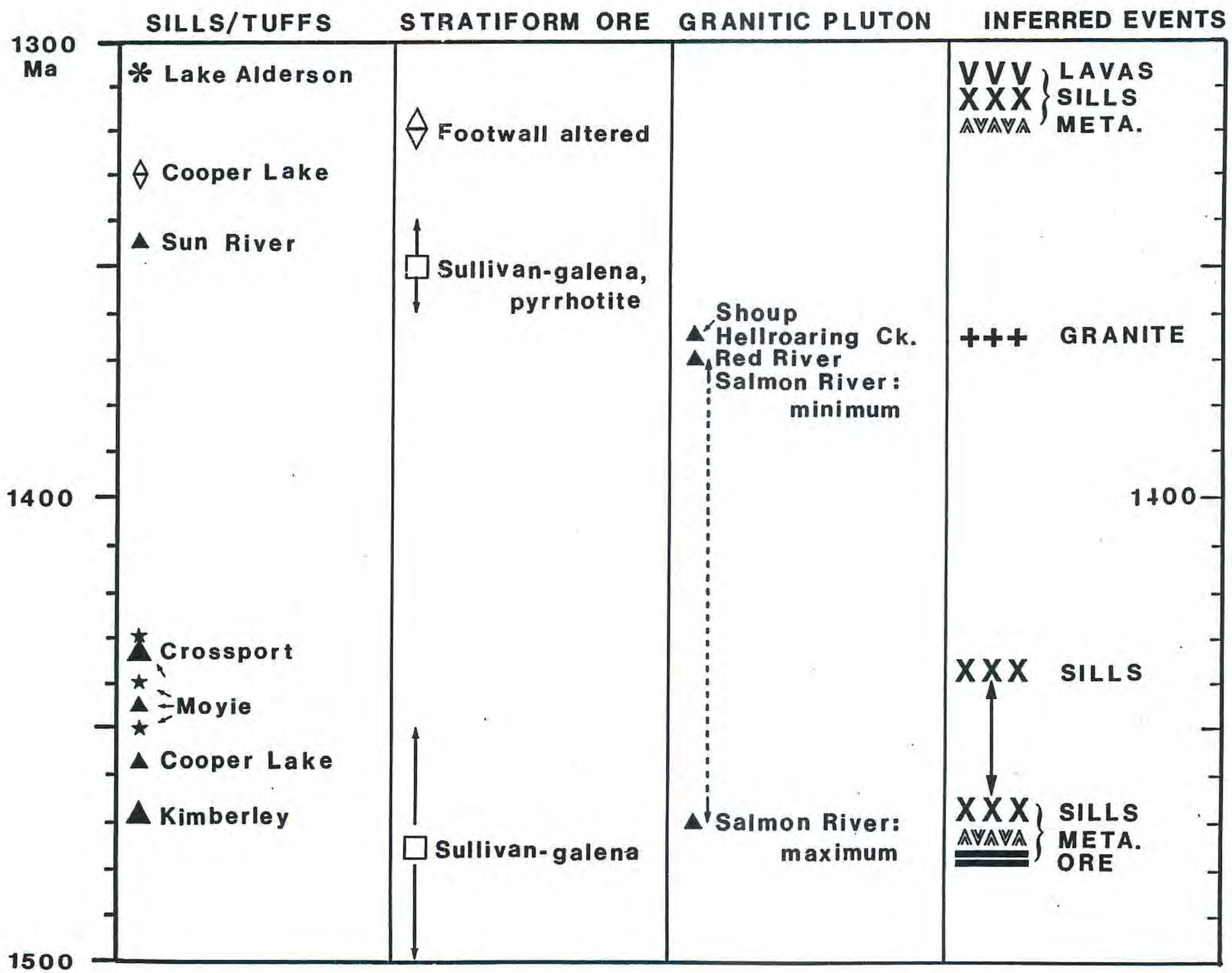
Figure 2. Radiometric ages in the 1300-1500 Ma interval. U-Pb: triangle=zircon; diamond=sphene (larger symbols indicate concordance: smaller symbols, slight discordance). Pb-Pb: square. Rb-Sr: asterisk. Sm-Nd: star.



GEOCHEMICAL AND SR AND ND ISOTOPE STUDIES OF PRE-BELT GNEISSES IN NORTHERN IDAHO AND NORTHEASTERN WASHINGTON

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Basement gneisses in the Sandpoint 1° x 2° quadrangle of northern Idaho and northeastern Washington (Fig. 1) were described by Hobbs and others (1965) as Belt equivalent, and by Aadland and Bennett (1979) as pre-Belt. These gneisses, whose outcrop areas are shown in Figure 1, range in metamorphic grade from upper amphibolite to granulite facies, contrasting with the greenschist facies of the Belt Supergroup. The most common rock type is biotite-sillimanite-gneiss, in which both fibrolite and rod-shaped crystals of sillimanite replace biotite. Kyanite is also present in some samples, with sillimanite replacing kyanite. Leucocratic and biotite-rich gneisses are also common, as are concordant tabular lenses of dark green to black amphibolite. In order to gain some information about the protoliths of these gneisses, representative samples were collected from each of the four areas A, B, C, and D shown on Figure 1, and 23 of these samples were chosen for geochemical analysyis, five each from areas A and D, seven from B, and six from C. Major and trace element analyses were obtained by XRF at Durham University. CIPW norms were calculated using IGRET (FeO/Fe_2O_3 = 9 for all samples).

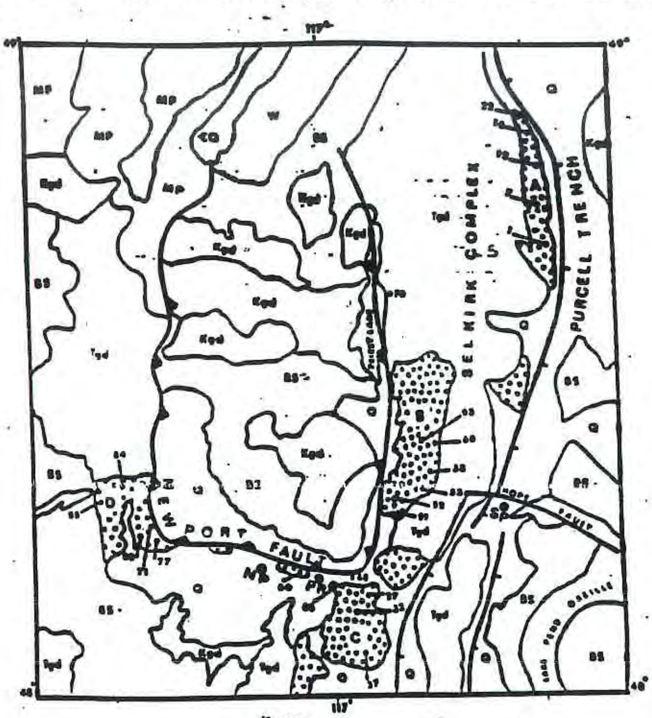

Figure 1. Geologic map of the Sandpoint quadrangle (after Aadland and Bennett, 1979). Pre-Belt gneiss is patterned and the main outcrop areas labelled A, B, C, and D. Sample locations are indicated by number. BS, Belt Supergroup; W, Windermere; CQ, Cambrian Quartzite; MF, Metaline Formation; MP, Maitlen Phyllite, Kgd, Cretaceous plutons; Tgd, Tertiary plutons; Q, Quaternary. Towns are marked by solid circles: NP, Newport; PR, Priest River; SP, Sandpoint. Heavy lines are faults, hachured where normal and barbed where thrust.

The normative Q-Ab-Or plot (Fig. 2) shows that samples 7, 12, and 22 (from area A), 53 (from area B), 32 and 37 (from area C), and 80, 84, and 85 (from area D) have high normative quartz. These samples also tend to have high normative orthoclase and corundum (except sample 32 which is quartzite). This combination of high normative quartz, orthoclase, and corundum suggests that these gneisses are of sedimentary origin. The samples with lower normative quartz have geochemical compositions that are more typical of igneous rocks and are therefor interpreted as orthogneisses. Samples 1 (area A), 63 (area B), 27 and 68 (area C) and 77 (area D) have compositions similar to granite, and also plot near the ternary minimum on the Q-Ab-Or plot. Samples 60 and 70 (area B), and 71 (area D) have the composition of granodiorite, and also plot closer to the albite corner on the Q-Ab-Or plot. Samples 51 and 52 (area B) and 45 (area C) contain no normative quartz and have the composition of diorite. Samples 16 (area A) 58 (area B) and 66 (area C) are high in normative anorthite and have the composition of basalt.

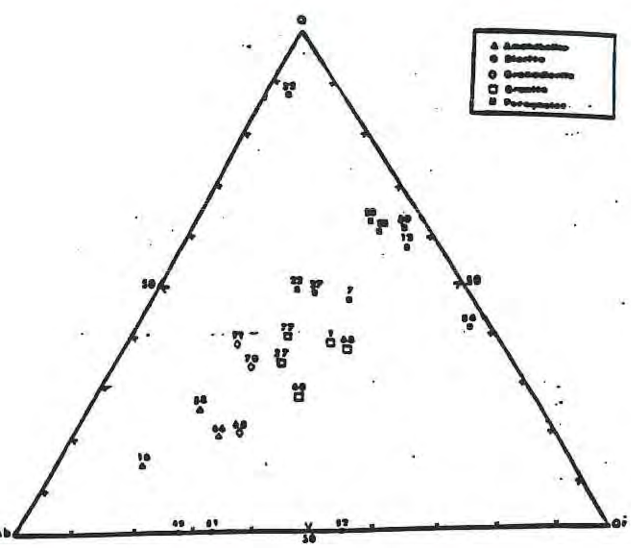

Figure 2. Triangular diagram showing normative Q-Ab-Or, Pre-Belt basement gneisses, Sandpoint 1° x 2° quadrangle.

The samples interpreted as paragneiss plot in a fairly compact group on Figure 3 (CaO against Y). They have low CaO concentrations and high but variable Y concentrations, whereas those interpreted as orthogneiss have much more variable compositions. The orthogneiss samples plot in positions typical of calc-alkaline igneous rocks (Lambert and Holland, 1974). From diorite, through granodiorite, to granite, the samples have progressively lower CaO contents and generally low Y concentrations, suggesting fractionation of Y-acceptor minerals such as hornblende and garnet. This indicates a calc-alkaline affinity and suggests that the igneous protoliths of these gneisses were calc-alkaline.

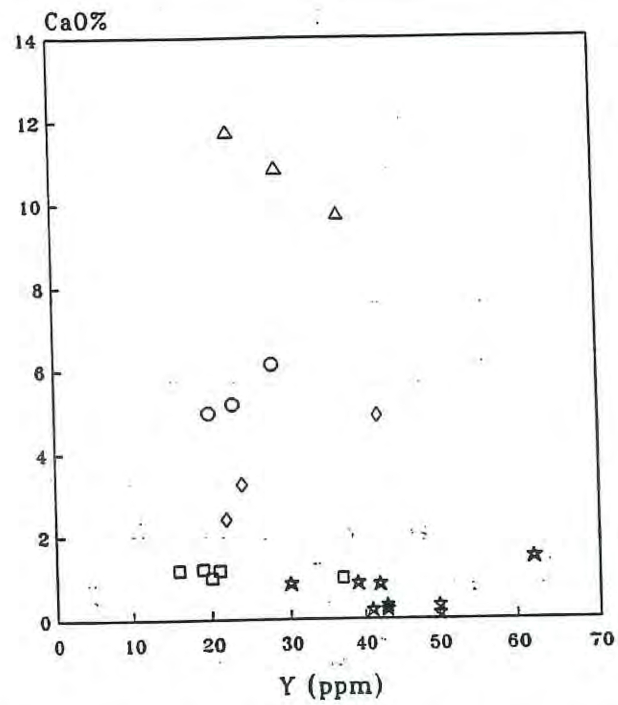

Figure 3. Plot of CaO vs Y, Pre-Belt gneisses, Sandpoint 1° x 2° quadrangle. Symbols as in figure 2.

Further evidence of the calc-alkaline nature of the orthogneiss is seen in Figure 4, where Nb is plotted against Y (Pearce and others, 1984). Here, the orthogneiss samples plot in the volcanic arc and syncollisional field, usually an indication of calc-alkaline affinity. As in Figure 3, there is a reasonable separation of the orthogneiss samples from the paragneiss samples, the latter mostly grouping together in the "within plate" field.

Sr and Nd isotopic analyses were made in order to gain some information concerning the pre-metamorphic history of the gneisses. Sixteen of the gneiss samples were analyzed for $^{87}Sr/^{86}Sr$ as well as for Rb and Sr concentrations. Thirteen were analyzed for $^{143}Nd/^{144}Nd$ and nine for Sm and Nd concentrations. A metamorphic age cannot be determined from the Rb and Sr data, which show a high degree of scatter on a Rb-Sr isochron plot. However, an estimate for a

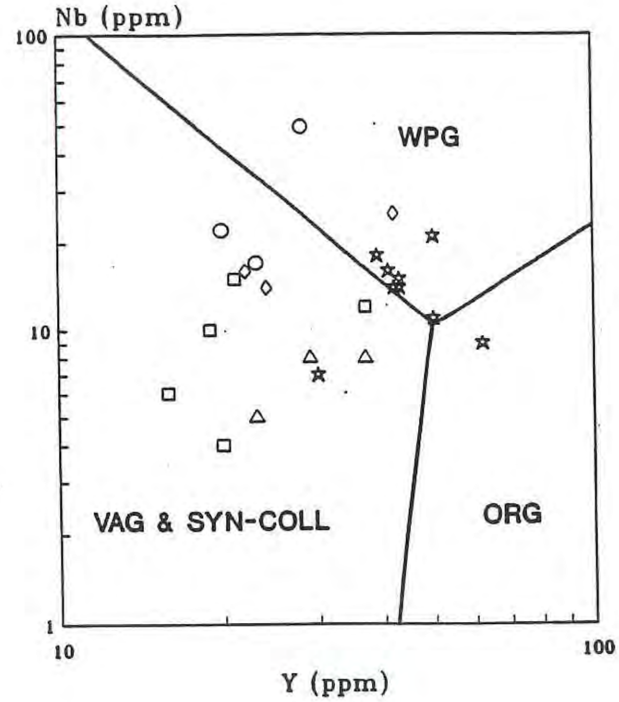

Figure 4. Plot of Nb vs Y, Pre-Belt gneisses, Sandpoint 1° x 2° quadrangle. WP, Within Plate field; VA and SYN-COLL, Volcanic Arc and Syn-collisional field; OR, Orogenic field. Symbols as in figure 2.

crustal residence age can be obtained by using the average whole-rock values of $^{87}Sr/^{86}Sr = 0.73430$, Rb = 122 ppm, and Sr = 460 ppm. Assuming an initial $^{87}Sr/^{86}Sr$ ratio of 0.705, a crustal residence age of 2600 Ma is obtained.

Nine Nd T_{DM} crustal residence ages range from 1400 Ma to 2430 Ma, with an average of approximately 2000 Ma (using a Bulk Earth value of 0.51264 today and De Paolo's "least depleted mantle" growth curve (De Paolo, 1988). The youngest Nd crustal residence age for the pre-Belt gneisses is close to the probable upper limit for the age of the Belt Supergroup metasediments. (Zartman and others (1982) dated the Crossport C sill in northern Idaho at 1433 Ma, and Le Couteur (1979) obtained a K-Ar age of 1436 Ma on a biotite from the Sullivan Mine. Hoy (1989) dated the Lumberton sill in British Columbia at 1445±11 Ma, which he interpreted as the age of sedimentation of the middle Aldridge Formation (equivalent to the Prichard Formation in the United States), because of evidence suggesting that the sediments were still wet at the time of intrusion.) However, these youngest pre-Belt crustal residence ages may have resulted from magma mixing during later intrusion, and so the true crustal residence ages of these gneisses are probably closer to the maximum Nd model ages of around 2400 Ma.

An epsilon Sr-Nd diagram is shown in Figure 5, where the unweighted average of all thirteen data points is Epsilon Nd = -14.7 and Epsilon Sr = 425. The negative Epsilon Nd and positive Epsilon Sr values suggest derivation from the continental crust, which is consistent with the geochemical evidence, and further suggests that the orthogneisses may be derived from S-type intrusives.

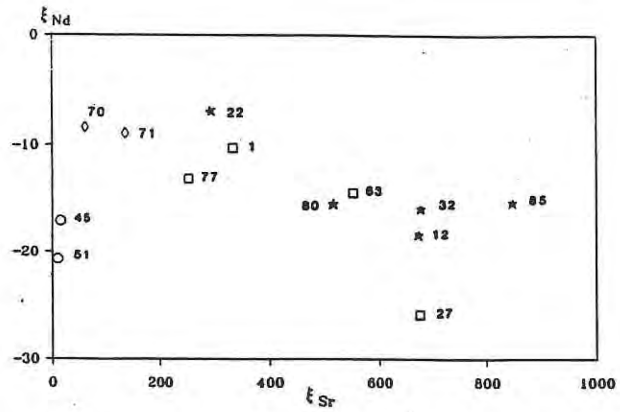

Figure 5. Epsilon Nd vs Epsilon Sr, Pre-Belt gneisses, Sandpoint 1° x 2° quadrangle. Symbols as in figure 2.

In conclusion, the pre-Belt gneisses of the Sandpoint quadrangle are of upper amphibolite to granulite facies, and consist of the same range of rock types in all four areas in which they are exposed, in the western wall of the Purcell Trench in the northern part of the quadrangle, as well as east, southeast, and southwest of the Newport fault, in the southern part of the quadrangle. In each area the predominant rock type is biotite-sillimanite-(kyanite)-paragneiss, but orthogneisses are also present, ranging in chemical composition from basalt to granite, with probable calc-alkaline igneous protoliths. Sr and Nd crustal residence ages range from 1400 Ma to 2600 Ma, and are interpreted as minima.

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PROVENANCE OF THE BONNER FORMATION (MIDDLE PROTEROZOIC BELT SUPERGROUP), MONTANA AND THE BUFFALO HUMP FORMATION (DEER TRAIL GROUP), WASHINGTON: PETROGRAPHIC EVIDENCES FOR MULTIPLE SOURCE TERRAINS

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INTRODUCTION

The Bonner Formation and its Canadian counterpart, the Philips Formation, forms a large arenite wedge, 200-500m thick, that is exposed over a wide area stretching from southwestern Montana to eastern Washington, Alberta and British Columbia (Fig. 1). It occupies the middle part of the Missoula Group of Middle Proterozoic Belt Supergroup and has been interpreted as a vast alluvial apron complex that headed south and west of the present Beaverhead Mountains and extended northward to eastern Washington and Canada (Quattlebaum, 1980; Winston, 1977, 1978, 1986; Winston and others, 1986; Stanley and Sinclair, 1989). Mostly on the basis of stratigraphy, a number of Belt workers (Fenton and Fenton, 1937; Ross, 1963; Harrison and Grimes, 1970; Harrison, 1972; Winston, 1984; Winston and others, 1986 and others), have proposed several source terrains for sediments of the Belt Supergroup, ranging from Archean rocks of Laurentia, east of the Belt Basin, to the Dillon Block of the Wyoming Province, south of the Belt Basin, to rocks that lay west of the Belt Basin that were rifted away in the Late Proterozoic (Sears and Price, 1978; Ross and others, 1992). Frost and Winston (1987) concluded from Sr/Nd analysis that 1.9-2.0 Ga modal crustal residence dates for mixed fine grained sediments combined with westward thickening sand wedges indicate that most Belt sediments were derived from a western, mostly Early Proterozoic, source terrain, not the Archean rocks of Laurentia or the Wyoming Province to the east or south. Ross and Parrish (1991) confirmed Frost and Winston's findings by reporting a mixture of Archean and Early Proterozoic zircons from Bonner Formation. They have more recently (Ross and others, 1992) interpreted most of the Bonner zircons to have come from Early Proterozoic crust, west of the Belt basin, mixed with fewer Archean zircons.

The Bonner Formation has also been considered a possible lithologic correlative of the Buffalo Hump Formation of Deer Trail Group of northeastern Washington (Evans, 1987; Miller and Whipple, 1989; Winston and Link, 1993). Based on detrital zircons and monazites from both the formations Ross and others (1991) inferred that the Bonner arenite was mostly derived from an Early Proterozoic magmatic belt but some is reworked Archean sediments. On the other hand, Buffalo Hump Formation contains zircons of 1.1 Ga, considerably younger than the proposed age of the Bonner Formation. Evans (1987) suggested an unconformity at the base of quartz sandstone beds within the Buffalo Hump Formation that separated the lower argillaceous part from the upper quartzose part.

In this study I have analyzed the Bonner Formation and the Buffalo Hump Formation petrographically in order to determine source terrains for each and their possible relationship. To date samples from Wise River and Flint Creek sections of the southwestern part of the Belt Basin, Nayak and Libby sections of the northern part of the Belt Basin, and from one section of the Buffalo Hump Formation of Stensger Mountain quad-rangle, eastern Washington, have been analyzed. Mineralogy and texture of framework detrital grains have been studied for petrographic comparison of the southern Bonner, the northern Bonner and the Buffalo Hump Formation. This study shows that the detritus of the southern Bonner probably came from metasedimentary, granitic and/or meta-morphic source rocks southwest of the Belt Basin, the northern Bonner probably came from volcanic (andesite ?), granitic and/or meta-morphic source rocks northwest or west of Belt Basin. The quartzose upper part of Buffalo Hump Formation was probably derived from deformed sedimentary (quartz arenite ?), granitic and metamorphic source rocks west of the Belt Basin, a source different from that of the Bonner Formation.

DATA AND ANALYSIS

Samples from four measured sections of the Bonner Formation and one section of the Buffalo Hump Formation (Evans, 1987) have been analyzed (Fig. 1). Thin sections were studied and the detrital grains are grouped into eight categories. Since this study is ongoing, point counting and population percentages are yet to be finished. The following data and analysis are based on identification of minerals and texture of detrital framework grains.

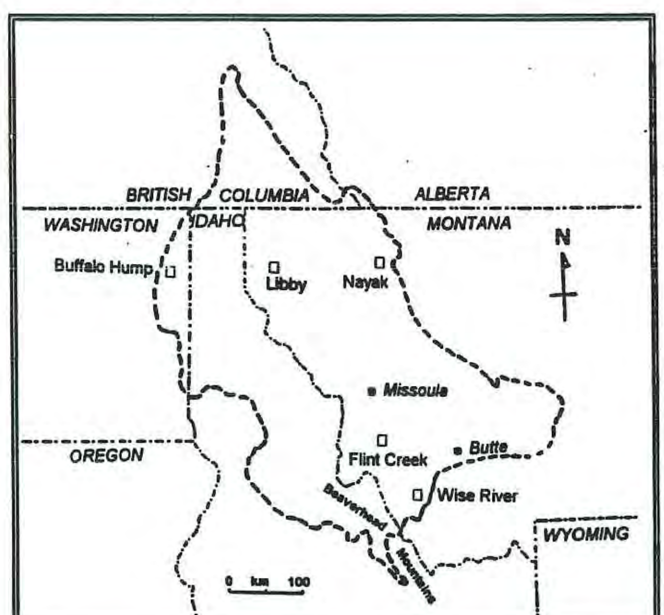

Figure 1. Map showing locations of four sections of the Bonner Formation (Belt Supergroup), and one section of the Buffalo Hump Formation (Deer Trail Group). Also shown is the boundary of Belt Basin.

Southern Sections of Bonner Formation

The Wise River Section of the Bonner Formation is composed of very poorly sorted, medium- to very coarse-grained arenite with pebble to cobble conglomerate beds in the middle part of the section. The detritus can be divided into six major types: 1) Poorly sorted, subrounded to well rounded, medium-grained monocrystalline quartz grains and pebbles; 2) Three distinctly different populations of poorly sorted, subrounded to well rounded, medium-grained to granule-size polycrystalline quartz grains; a) undulose, sutured subcrystals (stretched metamorphic quartz of Krynine (1946)), b) straight to slightly undulose, elongate subcrystals with straight boundaries (schistose metamorphic quartz of Krynine (1946)), and c) straight, equant subcrystals with simple boundaries (recrystallized metamorphic quartz of Krynine (1946)); 3) fresh, angular to subrounded, medium- to fine-grained microcline; 4) fresh to slightly altered, angular to subrounded, medium- to course-grained orthoclase; 5) rounded to subrounded, medium- to fine-grained chert; and 6) medium to fine-grained muscovite. Further, in recrystallized metamorphic quartz three categories have been identified; i) micro-mosaic; equant interlocked crystals ranging from 20 to 30 microns in diameter, ii) Macro-mosaic; semiequant interlocked crystals ranging from 90 to 200 microns in diameter, and iii) Mega-mosaic; semiequant, interlocked crystals ranging from 200 to 500 microns in diameter.

The Flint Creek section of the Bonner Formation is composed of poorly sorted, subangular to well rounded medium- to very coarse-grained arenite and pebble conglomerate. The detrital mineralogy and texture is the same as that of Wise River section except that grains are relatively smaller, schistose metamorphic quartz grains are absent, and the amount of microcline is diminished.

Northern Sections of Bonner Formation

The Nayak section of the northern Bonner Formation is composed of poorly sorted, very angular to subrounded, medium- to very fine-grained arenite. The detrital grains are of seven types: 1) poorly sorted, rounded to well rounded fine- to coarse- grained monocrystalline quartz; 2) Two distinctly different populations of poorly sorted, subangular to well rounded, fine- to coarse-grained polycrystalline quartz; a) recrystallized metamorphic quartz and b) stretched metamorphic quartz; 3) poorly sorted, subrounded to angular, very fine- to coarse-grained orthoclase; 4) poorly sorted, subrounded to angular, very fine- to coarse-grained microcline; 5) poorly sorted, subrounded to rounded, very fine- to fine-grained plagioclase; 6) poorly sorted fine- to medium-grained muscovite; and 7) trace amount of rounded, medium- to coarse-grained chert.

The Libby section of the Bonner Quartzite is composed of poorly sorted, subangular to subrounded, very fine- to medium-grained arenite with subordinate well rounded coarse-grained arenite. The detrital grains belong to six major types: 1) poorly sorted, subangular to rounded, very fine- to coarse-grained monocrystalline quartz; 2) poorly sorted, angular to rounded, fine- to coarse-grained recrystallized metamorphic quartz of macro-mosaic type; 3) moderately sorted, subangular to subrounded, fine- to medium-grained orthoclase; 4) well sorted, subangular to subrounded very fine-grained plagioclase; 5) fine-grained muscovite; and 6) trace amounts of rounded to subrounded, fine-grained chert.

Buffalo Hump Formation

In contrast to the Bonner Formation, the Buffalo Hump Formation is composed of beds of poorly sorted, angular to well rounded, fine- to very coarse-grained quartz arenite with subordinate granule size quartz grains that alternate with beds of argillite. The detrital grains can be divided into four major types. They are: 1) poorly sorted, angular to well rounded, very fine-grained to granule size monocrystalline quartz; 2) two different types of poorly sorted, subrounded to well rounded, medium-grained to granule size polycrystalline quartz; a) recrystallized metamorphic quartz of micro-, macro-, and mega-mosaic types, and b) stretched metamorphic quartz; 3) poorly sorted, angular to subangular, medium- to coarse-grained metamorphic rock fragments; and 4) medium- to fine-grained muscovite grains. Thus the Buffalo Hump Formation differs from the Bonner Formation in at least two ways; A) Buffalo Hump Formation contains no feldspar whereas Bonner Formation is rich in orthoclase and microcline in the south and orthoclase, microcline and plagioclase in the north; B) Buffalo Hump is rich in metamorphic rock fragments whereas Bonner has no metamorphic rock fragments in either part.

CONCLUSIONS

Preliminary conclusions include: (1) the southern and the northern parts of the Bonner Formation had different source terrains. (2) The southern Bonner, which contains orthoclase and microcline, was derived from at least two different types of source rocks; a) a distant metamorphosed quartz arenite which provided far-travelled rounded second cycle quartzite pebbles and sand grains, and b) a proximal granitic and/or metamorphic source which contributed short-travelled angular orthoclase and microcline. (3) The northern part of the Bonner Formation, which contains plagioclase in addition to microcline and orthoclase, also received sediments from at least two different source rocks; a) a distant granitic and or metamorphic source which provided far-travelled, rounded quartz, and b) a proximal volcanic (andesitic ?) and or granitic source which contributed angular plagioclase and potassium feldspar. (4) Sediments of the northern part of Bonner Quartzite were probably derived from west or northwest of the Belt Basin and were deposited before deposition of Buffalo Hump Formation. Thus, rather than being one large uniform arenite wedge, Bonner Formation is probably a composite wedge, which received sediments from two different source terrains, one from west or southwest and the other from west or northwest. (5) A western source suggests that tectonism was more active on the west side of the Belt Basin than east. (6) Petrographically, quartzitic beds of the Buffalo Hump Formation do not appear to correlate with the Bonner Formation as was suggested by Miller and Whipple (1989). (7) The detritus of the Buffalo Hump Formation, which is devoid of recognizable feldspar, were dominantly derived from a distant metamorphic source terrain, possibly located north northwest of the Belt Basin. (8) Therefore, Buffalo Hump Formation appears to be a separate lithologic unit, different from the Bonner Formation.

Parts of the Lawson Creek Formation and part of the Swauger Formation (Ross, 1947; Ruppel and others, 1975) of east central Idaho are also correlated lithologically with the Bonner Formation (Hobbs, 1980; Winston and Link, 1993). A goal of this study is to clarify these correlations.

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SOME OBSERVATIONS REGARDING FORMATION CORRELATIONS AND REGIONAL PALEOGEOGRAPHY OF THE SOUTHERN HELENA EMBAYMENT

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INTRODUCTION

Belt Supergroup rocks exposed in the Whitehall, Mt. area are near the southern margin of the present-day expression of the Helena embayment. Most of the 5,000 to 8,000 foot-thick Proterozoic section here is interpreted as a progradational basin plane/submarine fan/slope complex (Foster and others, this volume). The depositional environments were defined using the facies association scheme presented in Nelson and Nilson (1984). Several interesting regional implications were recognized as a result of these studies, limited regional reconnaissance across the southern Helena embayment, and comparing this work with published literature. A location map showing the areas discussed below is also given in Foster and others (this volume).

FORMATION CORRELATIONS

Attempts to apply the lithologically-based formation nomenclature from the central Helena embayment to the Proterozoic rocks in the Whitehall area have created some confusion. Greyson Formation has been applied to both shale-rich slope/shelf facies associations above the 3,000 foot-thick submarine fan complex, as well as to the basin plain shales beneath it (Alexander, 1955; McMannis, 1963). Similarly, Newland Formation has been applied to calcareous units both in the upper part of the fan complex (Hawley and others, 1982; McMannis, 1963), as well as to the calcareous basin plain facies association beneath it (McMannis, 1963).

It is not clear which central Helena embayment formations correlate with the Proterozoic rocks in the Whitehall area. Correlations based upon general lithologic similarities are questionable in units with such rapid and repetitive lithologic changes. In addition, these rocks appear to have been derived from a southern source, and do not interfinger with Greyson, Newland, or any central embayment formations in the area examined. The Proterozoic rocks in the Whitehall area have long been correlated with the lower Belt Supergroup. This is reasonable based upon their depositional setting, position within the basin, and style of associated mineralization, but is by no means absolute.

The Proterozoic rocks in the Whitehall area are overlain by Cambrian sedimentary rocks. It can therefore be argued that some of the Whitehall area Proterozoic could be younger than lower Belt, and perhaps correlative with middle or even upper Belt units. Although lower Belt is not an unreasonable assignment for the Whitehall Proterozoic rocks, confident correlation will require additional study.

REGIONAL PALEOGEOGRAPHY

Previous studies and depositional environment interpretations of the LaHood Formation are summarized by Schmitt (1988). We interpret most of the LaHood in the Whitehall area as basin plane and various submarine fan facies associations. The fan was deposited in deep water as evidenced by the lack of depositional structures indicating wave base effects or subaerial exposure, by the preservation of synsedimentary sulphide horizons and turbidite deposits, and by its stratigraphic position between basin plain and shelf deposits. Alluvial fan deposits which overlie the submarine fan in the southern part of the Whitehall area have also been previously identified as LaHood Formation.

The submarine fan appears to have prograded in a north to northwesterly direction based upon paleocurrent and paleoslope measurements. Approximately 120 measurements were taken from the various facies associations. The aerial distribution and lithofacies changes within the fan complex also support a north to northwesterly progradation.

Cressman (1989) describes two shallowing-upward basin fill sequences in the lower Belt Prichard Formation of the main Belt basin. The upper sequence is described as a large submarine fan/slope/shelf complex that prograded to the north-northwest, and occupied much of the Belt basin. Finch and Baldwin (1984) believe that the Prichard Formation contains several such northwesterly prograding fans. The north to northwesterly prograding fan/slope/shelf deposits in the Whitehall area could be part of the same deposystem. If so, much of the Belt basin would have been occupied by one or more large prograding fan complexes during lower Belt time.

Much of the LaHood Formation outside of the Whitehall area also appears to be composed of submarine fan deposits. The LaHood Formation is situated along the southern edge of the Helena embayment, and extends from the Highland Mountains on the west to the Bridger Range on the east. Reconnaissance examination of the LaHood Formation in the Highland Mountains west of Whitehall by the authors suggests that it also contains submarine fan deposits. Bonnet (1979) interpreted the LaHood east of Whitehall in the Horseshoe Hills and Bridger Range as part of a submarine fan. The southern Horseshoe Hills were also briefly examined, and we concur with the submarine fan interpretation. In addition, four of five paleocurrent measurements indicated a north to northwesterly progradation.

It therefore appears that sediment may have been supplied to the Helena embayment by north to northwesterly prograding submarine fan deposits distributed all along its southern margin. It is not known whether this deposystem consisted of numerous individual fans, or perhaps one large fan complex derived from a line source all along the Willow Creek fault. As discussed above, these deposits may have fed or joined submarine fan deposits in the main Belt basin. Submarine fan deposits also occur in the Proterozoic Yellowjacket Formation in central Idaho. Hughes (1982) suggests that the axis of this depositional basin plunged northwesterly. Perhaps these fan deposits also fed into the main Belt basin.

The Belt basin appears to have extended south of the Willow Creek fault, where it comprised a carbonate shelf. Calcareous shales and micrites and calcarenite beds composed of detrital carbonate occur in the submarine fan deposits of the Whitehall area. Some of these calcarenites have also been described by Coppinger (1981) and Hawley and others (1982). A carbonate-dominant source terrain was required to produce such beds. Archean rocks south of the Willow Creek fault contain only a few isolated marble beds (Vitaliano and Cordua, 1979), and cannot be considered a source. However, a carbonate shelf bordering the submarine fan complex could have provided a carbonate source sufficient to produce these calcarenite and fine-grained carbonate beds.

A number of large undated bodies of carbonate within the submarine fan and slope facies associations are exposed north of the Willow Creek fault in the Whitehall area. They have been interpreted as basement topographic highs (Lowell, 1956; McMannis, 1963) and as tectonic slivers (see geologic map in Schmidt, 1975; Schmidt and others, 1988). Several of these anomalous carbonate bodies occur along the Willow Creek fault in the Mayflower Canyon area. Along their margins, they grade into carbonate-clast conglomerates in Proterozoic submarine fan/slope deposits. Their isolated occurrence and gradational contacts indicate that they are large slump blocks or olistoliths, as originally suggested by Hawley (1973) and Hawley and others, (1982). We further suggest that they may be slumped blocks of the Proterozoic carbonate shelf which extended south of the Willow Creek fault. Other isolated bodies of Archean metamorphic rocks along the Willow Creek fault may also be olistoliths.

Proximal submarine fan deposits north of the Willow Creek fault in the LaHood Canyon area contain abundant carbonate clasts. Berg (1991) states that most of the clasts are fine-grained calcitic marble, in contrast with the coarse-grained dolomitic marble which typifies the Archean. Again, a Proterozoic-age shelf source for these clasts could explain why their texture and composition is distinct from the Archean carbonates. The thin shelf deposits would have been removed by subsequent erosion of the Tobacco Root Mountains, which border the Willow Creek fault to the south. The evidence for the carbonate shelf is admittedly suggestive, but certainly warrants further study.

Detrital carbonate-rich intervals are common near the top of the LaHood Formation (i.e. in the upper part of submarine fan deposits) in the Whitehall area. Detrital upper LaHood carbonates have also been observed by the authors in the Highland Mountains and southern Horseshoe Hills, where they locally comprise over 20 percent of the section. Bonnett (1979) described the detrital carbonates in the Horseshoe Hills as well as in the nearby Bridger Range. We suggest that these calcareous rocks may also have been derived from a carbonate shelf. If so, a single large carbonate shelf would have bordered the entire southern margin of the Helena embayment. Progradation and erosion of the shelf would account for the incorporation of carbonate detritus into the upper submarine fan sediments across such a large area.

SUMMARY

Much of the LaHood Formation in the Whitehall area was deposited by submarine sediment gravity flows as originally suggested by Hawley (1973). In particular, it is a prograding submarine fan complex which is divisible into mappable outer, middle, and inner fan deposits. LaHood Formation in the southern part of the area contains some alluvial deposits which overlie the fan complex. The submarine fan prograded in a north to northwesterly direction. It appears to have been one of a series of fan deposits which were distributed along the southern margin of the Helena embayment. These may have been part of a considerably larger group of fans which could have occupied much of the Belt basin during lower Belt time.

The Belt basin may have extended south of the Willow Creek fault, as previously suggested by Reynolds (1984) and Schmidt and Garihan (1986). Specifically, we suggest that a carbonate shelf likely extended south of the fault. Erosion of the shelf would explain the detrital calcarenite and fine-grained calcareous beds, clasts of fine-grained calcitic marble, and large slump blocks or olistoliths of carbonate rock in the Proterozoic basinal deposits of the Whitehall area. Scattered exposures of Archean metamorphic rocks along the northern side of the Willow Creek fault may also be olistoliths. The entire southern margin of the Helena embayment may have been bordered by this carbonate shelf. Progradation and erosion of such a shelf would have produced the carbonate-rich intervals which are common in the submarine fan deposits of the upper LaHood Formation from the Highland Mountains to the Bridger Range.

Previous workers in the Whitehall area have assigned basin plain deposits below the fan complex and slope/shelf deposits above it, which are separated by 3,000 feet of strata, to the Greyson Formation due to the abundance of siltstone and shale. Similarly, calcareous rocks in the basin plain and upper part of the fan complex have previously been designated Newland Formation due to the presence of carbonate.

Because the Proterozoic rocks in the Whitehall area were derived from a southerly source and are not exposed interfingering with Greyson, Newland, or any other central embayment formations, they cannot be reliably correlated. Bonnet (1979) and Hawley and others (1982) were also unable to correlate LaHood Formation with the finer grained central embayment formations to the north. It is conceivable that the Whitehall area Proterozoic rocks could in part be younger than lower Belt.

The possibility of a carbonate source along the southern basin makes correlation of calcareous rocks in the southern Helena embayment with the Newland Formation of the central embayment questionable. Such correlations could be erroneous in the absence of other evidence. This illustrates the potential pitfalls of attempting to correlate basin margin with central basin deposits using general lithologic similarities alone.

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PALEODEPOSITIONAL SETTING AND SYNSEDIMENTARY MINERALIZATION IN BELT SUPERGROUP ROCKS OF THE WHITEHALL, MONTANA AREA

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INTRODUCTION

Belt Supergroup rocks exposed in the Whitehall area have been previously identified as the LaHood, Newland, and Greyson Formations (Figure 1). Paleodepositional environment studies were recently conducted by Nilsen and Chadwick at the request of the Golden Sunlight Mine. Much of the 5,000 to 8,000 foot-thick Proterozoic section here is interpreted as a progradational basin plain/submarine fan/slope complex. Shelf and alluvial deposits overlie the complex in the northern and southern part of the area, respectively.

The Cretaceous breccia pipe-hosted gold deposit at the Golden Sunlight Mine is superimposed upon uneconomic Proterozoic-age sulphide mineral concentrations. Considerable core drilling in the mine area for ore estimation and slope stability purposes has afforded an opportunity to evaluate the character and depositional setting of the sedimentary rocks and associated sulphides. The term synsedimentary is used in this paper to define all sulphides which formed prior to or during initial lithogenesis of the host sedimentary rocks.

PALEODEPOSITIONAL SETTING

Basin plain deposits are poorly exposed because they are composed predominantly of easily weathered shale. They consist of carbonaceous black shales, with interbeds of laterally continuous turbidite sandstone that do not display vertical organization or cyclicity. The basin plain deposits are overlain in ascending order by outer, middle, and inner fan deposits. The submarine fan is the mixed sediment variety, and includes sediment ranging in size from clay to sand.

Outer fan deposits contain thicker sandstone beds than the basin plain. These sandstone lobes are less laterally continuous than the basin plain sandstones, but are not channelized. The lobes form thickening- and coarsening-upward sequences, and are interbedded with black shales. Middle fan deposits contain massive sandstone channels that thin and become finer up-section. Overbank and levee deposits of interbedded shale and sandstone occur marginally to the channels.

Inner fan channel deposits contain abundant amalgamated sandstone and conglomerate channels but little intervening shale, resulting in resistant outcrops. Cobbles and boulders are common in proximal inner fan channels, as can be seen along Montana Highway 2 in LaHood Canyon. Grain size of inner fan channels decreases rapidly in a northerly direction. Inner fan overbank deposits contain considerable shale and minor sandstone. A section of calcareous inner fan overbank deposits occurs on Bull Mountain. It is approximately 300 feet thick, and contains calcarenite beds composed of detrital carbonate, calcareous shales, and micrite.

Submarine canyon and slope deposits overlie the submarine fan deposits. They are finer grained, consisting of shales and siltstones with a few sandstone beds. The slope deposits are divided into upper and lower slope in the Golden Sunlight pit where they are well exposed.

Lower slope rocks are coarser grained (siltier) than upper slope rocks, and contain fining- and thinning-upward cycles as much as 75 feet in thickness. The lower slope rocks display considerable soft sediment deformation, including beds with plastically deformed intraclasts. These are thought to have been the result of submarine slumping and associated accumulation of debris flows on the lower slope. The upper slope rocks consist of even, plane-parallel, graded laminae of silt to mud. They contain numerous low-angle depositional discordances, which are thought to represent headwall scarps that were infilled after slumps were shed off of the upper slope.

Shelf deposits overlie the slope in the northern part of the area, on Bull Mountain. They consist of well-stratified shale, siltstone, and sandstone. The presence of hummocky cross stratification and erosional surfaces are thought to reflect reworking by waves during storm events.

Alluvial fan deposits overlie slope deposits in the southern part of the area, near the Willow Creek fault. They consist primarily of pebbly sandstone interbedded with minor amounts of shale. The sandstone beds are typically massive and poorly sorted, with coarse matrix-supported angular clasts, and are thought to be subaerial debris flow deposits. Mudcracks in interbedded shales also support a subaerial depositional environment. These alluvial fan deposits appear to become finer and show increased stratification, rounding, and sorting to the north. This transition may represent a change from proximal (debris flow-dominated) to distal (streamflow-dominated) alluvial fan deposits.

SYNSEDIMENTARY MINERALIZATION

Inner fan, submarine canyon, and slope deposits are exposed in the Golden Sunlight open pit. These rocks host synsedimentary massive, semi-massive, and disseminated sulphides. Massive sulphides are defined as those rocks with greater than 50 volume percent sulphide, whereas semi-massive sulphides contain 15 to 50 volume percent sulphides.

Massive to semi-massive sulphides are best developed in soft sediment deformed submarine canyon deposits. These sulphidic rocks are as much as 16 meters thick, and consist of millimeter-thick laminae of fine-grained pyrite and quartz/feldspar silt, alternating with laminae of pyrite and green mica. Pyrite content ranges from 15 to over 50 volume percent. In addition to stratiform sulphide laminae, soft sediment folded sulphide beds, plastically deformed sulphide intraclasts and matrix, and paragenetically early sulphide cements are present. These textures indicate that the sulphides formed prior to lithification of the sediment, but it is not known whether they are syngenetic or diagenetic.

Multielement geochemical analyses were obtained on 13 small (30 to 60 gram) core samples of this pyritic rock. Only samples with very minor or no visible epigenetic fracturing were analyzed in an attempt to approximate the synsedimentary element signature. They contained moderate concentrations of Ba (220-1300 ppm), F (1200-4200 ppm), and Te (0.1-8 ppm). Low concentrations of Au (30-500 ppb), As (50-350 ppm), and Cu (60-2000 ppm) are also present. The samples were notably low in Pb (<81 ppm), Zn (<44 ppm), and Co (<87 ppm). This synsedimentary system therefore appears to have had a copper-gold affinity.

A distinctive lithology overlies the massive to semi-massive sulphides, and has a similar thickness and restricted aerial distribution. It is a dark, dense rock which consists of massive to planar interlaminated siltstone and shale. It consists of quartz, phlogopite, pyrite, and minor plagioclase (G.A. Desborough written commun., 1991). Irregular to conformable lenticular blobs and laminae of microcrystalline quartz are locally present. Pyrite content averages 5-15 percent by volume, and occurs as relatively coarse crystalline laminae, blebs, and conformable disseminations. This rock appears to have been a chemical sediment which formed during the Proterozoic hydrothermal event.

Disseminated sulphides are common in upper slope deposits which overlie the chemical sediment described above. They occur throughout hundreds of feet of section in the mine area. The upper slope deposits consist of predominantly planar laminated centimeter-scale fining-upward depositional units. They contain relatively clean planar laminated to ripple cross-laminated siltstone overlain by micrograded mudstone. Interstitial cement containing pyrite, ferroan carbonate, and minor chalcopyrite is concentrated in the basal siltstone beds. This stratiform pyrite occurs as complex blebs and fine disseminations, and is coated by supergene chalcocite and covellite in oxidized near-surface exposures. The pyrite itself appears to be cupiferous, because the trace concentrations of chalcopyrite are not voluminous enough to account for the abundant supergene copper minerals. Copper grades average 100 ppm throughout the slope facies, and locally exceed 300 ppm. The sulphides are thought to be diagenetic due to their occurrence as an interstitial cement.

Authigenic orthoclase and albite are associated with all of the above-mentioned styles of synsedimentary mineralization. In the massive to semi-massive sulphides and overlying chemical sediment, the feldspars occur as an interstitial cement within individual laminae and intraclast-bearing beds. It is particularly common in the micrograded mud laminae of the slope deposits. Authigenic orthoclase has also been identified in lower Belt strata at York, Montana where it is associated with gold (Baitis, 1988; Thorson and others, this volume; Whipple and Morrison, this volume). Authigenic feldspars have also been reported associated with copper mineralization in Proterozoic basin-margin facies in Morocco (Leblanc and Billaud, 1990).

Analyses were performed on the feldspar by G.A. Desborough (written commun., 1991). It consists of very fine-grained (less than 10 micron) orthoclase and lesser amounts of albite. The crystalline morphology of the feldspar grains and their lack of rounding and abrasion are not consistent with a detrital origin, and they are therefore thought to be authigenic. One SEM photo shows orthoclase growing around a detrital rutile grain, suggesting a diagenetic origin. The authigenic orthoclase differs from orthoclase in felsic igneous rocks in the mine, which is more ordered toward microcline.

A series of high-angle northeast-trending structures is present in the mine. The massive to semi-massive sulphides, overlying chemical sediment, and authigenic feldspar all appear to be confined to this trend, and to thicken toward its axis. Soft sediment deformation and beds containing plastically deformed intraclasts are also common in these rocks. This indicates that the structural zone was active during Proterozoic sedimentation, and focused hydrothermal activity. The zone may have further enhanced sulphide deposition by creating a reduced environment and/or a topographic low. The Golden Sunlight gold deposit formed along this same trend during the Late Cretaceous, and was superimposed upon the synsedimentary mineralization. The trend is thought to be related to the Great Falls Tectonic Zone (Foster and Chadwick, 1990).

The multielement signature of the synsedimentary mineralization discussed above is very similar to the signature of the Cretaceous gold system. It is possible that some metals were recycled and epigenetically upgraded from the metalliferous Proterozoic rocks into the Cretaceous hydrothermal system which was district wide in scale and effect. Regardless of the source of the metals, the Golden Sunlight Mine illustrates the important role of periodically reactivated basement structure over Proterozoic, Laramide, and younger metallization.

SUMMARY

Studies of the Proterozoic sedimentary rocks at Golden Sunlight have been an important part of the geological research program. This work has helped identify favorable hosts, local and regional structures, and has generated new exploration targets. The ability to recognize various Proterozoic clasts in the Cretaceous breccia pipe has shed light on the mechanics of pipe formation. The work has also been extremely useful from a practical operational standpoint for slope stability and grade control.

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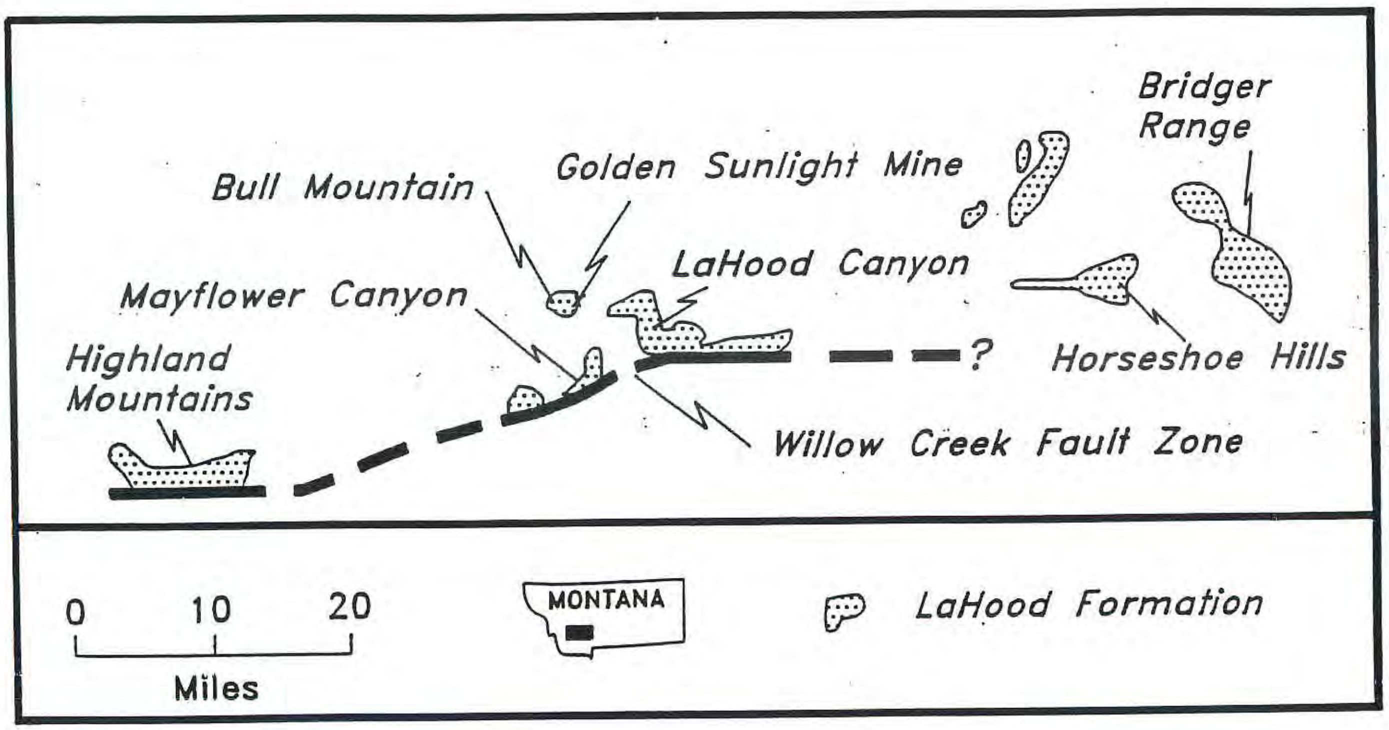

Figure 1: Location of the Golden Sunlight Mine with respect to exposures of the LaHood Formation (modified from McMannis, 1963).

PRELIMINARY GEOLOGIC SECTION TO BASEMENT FROM CHEWELAH, WASHINGTON,

TO GLACIER NATIONAL PARK, MONTANA

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The geologic terrane crossed by the section from Chewelah to northern Glacier National Park can be divided crudely into three structural zones, each with its own complexities. In common are thrust faults resulting from the continental override that began in Jurassic time and lasted through the Cretaceous (Monger and Price, 1979) and listric normal extension faults of Tertiary age documented by numerous workers along the western margin of the North American continent. Unique to the terrane west of the Purcell trench is the abundance of Cretaceous and Tertiary intrusive rocks that disrupt surface continuity of stratigraphic packages and of many structural patterns (Fig. 1, section A-A´). In the middle part of section A-A´, between the Purcell trench and the Rocky Mountain trench, is a thick sequence of Belt Supergroup strata; listric thrust faults are as common as ramp-and-flat thrusts and listric normal extension faults (at least 200) are characteristic and too closely spaced to show at the scale of Figure 1. East of the Rocky Mountain trench, the remnant Belt section is relatively thin above the Lewis thrust that carries Belt strata eastward over Paleozoic and Mesozoic rocks; ramp-and-flat thrusts and duplexes are common and extensional faults sparse (Fig. 1, sections A-A´ and B-B´).

Several sources of geologic and geophysical data were used to constrain interpretations shown in the sections. Surface geology is based on 1:250,000-scale geologic maps of the Sandpoint (Miller and Burmester, written commun., 1992), Kalispell (Harrison and others, 1992), and Cut Bank (Harrison and Whipple, in prep.) 1°x2° quadrangles. Slight variations of the geologic interpretations of Harms and Price (1992) for the area around the Newport fault are incorporated here. Seismic control is from data and interpretations by Potter and others (1986) for Washington and Idaho, variations of seismic interpretations by Yoos and others (1991) for the central area, and from Fritts and Klipping (1987a, 1987b) for the area east of the Rocky Mountain trench. Unpublished gravity maps of the Kalispell and Sandpoint quadrangles compiled by my colleague M.D. Kleinkopf (U.S. Geological Survey, written commun., 1992) were used to help identify the basement.

Several concepts are built into the interpretations shown in the sections. The basal surface of detachment is shown as a line, but it is probably a wide shear zone in previously stretched crust (Price, 1984; Cressman, 1989) on which the Belt basin formed. Precambrian crustal attenuation, which probably destroyed or homogenized the distribution of crustal magnetite, is also supported by the disappearance of west-southwest-trending magnetic stripes and anomalies that extend from the Superior Province of Canada to the edge of Belt terrane (Committee for a Magnetic Anomaly Map of North America, 1987) into a magnetically quiet zone. Crustal attenuation also is shown by periodic intrusion into Belt rocks of basic sills and extrusion of lava (Harrison, 1972). Deep seismic reflections (Potter and others, 1986) show many subhorizontal reflections in this layered or gneissic basement, and these multiple reflectors make choosing the base of the Belt strata difficult. The Belt sequence and underlying gneissic basement were broadly folded before the basal Paleozoic unit (Middle Cambrian Flathead Quartzite) was deposited (Harrison and others, 1992, Harrison and Cressman, 1993). The Belt sequence was also faulted in pre-Flathead time as shown by a pre-Flathead unconformity that rests on various strata from uppermost to lower Belt, and has thousands of feet of pre-Flathead displacement on fault blocks within a few miles of each other. This pre-Flathead folding and faulting means that (1) all folds are not necessarily related to the subsequent Cretaceous thrusting and folding, and subsurface thrust forms (blind thrusts, ramp folds, and so forth) cannot necessarily be based on concepts of tectonics formulated from terranes in which strata were flat before thrusting, and (2) a Cambrian "surface" at a low angle of unconformity does not extend uniformly across the Belt basin and cannot be used to reconstruct shape, size, and depth of the Belt basin.

The western part of section A-A´ shows unique features of thrusting and extension. The Jumpoff Joe thrust fault juxtaposes the Deer Trail Group, a partial correlative of the Middle Proterozoic Belt Supergroup (Miller and Whipple, 1989), as well as unconformably overlying panels of Late Proterozoic Windermere Supergroup and Paleozoic strata, against Belt strata. East of the Jumpoff Joe thrust fault, large Cretaceous plutons intruded the Belt and underlying basement to form the Priest River terrane, which consists of Belt, metamorphosed Belt, and basement rocks in a tectonic and plutonic culmination. Eocene extension in the complex (Harms and Price, 1992) is reflected by the flat-lying Newport fault and by Tertiary plutons that intruded a dilatent zone. The extension lowered a panel of Belt and other rocks above the Newport fault onto the underlying Priest River terrane. In my interpretation, the Newport fault is unique and does not connect with either the Jumpoff Joe thrust or Purcell trench faults (Fig. 1).

The section from the Purcell trench east to the Rocky Mountain trench shows the effects of thrusting in previously folded strata. The Cretaceous Moyie thrust fault, whose root must have joined the basal detachment before being destroyed by a major Cretaceous intrusion (Fig. 1), has a translation eastward of about 30 miles and rides up the flank of the Proterozoic Sylvanite anticline. A splay of the Moyie clipped off a piece of the anticline and places younger rocks on older rocks along the west side of the anticline. Farther east, the Libby thrust belt is the remnant of a Proterozoic syncline now mostly destroyed during the thrusting that jammed the Sylvanite anticline against the west-facing limb of the Purcell anticlinorium. Complicating the structural picture is a Cretaceous ramp fold above the Pinkham thrust fault. Not obvious from the section but clear on the geologic map is the fact that the Cretaceous ramp fold refolds part of the Proterozoic Purcell anticlinorium. Tertiary extension of the region is shown symbolically by a few of the tens of listric normal faults and by backsliding on some of the Cretaceous thrust faults. Also shown is the gradual decrease in the number of sills in the Prichard Formation eastward from major intrusive centers in the west (Cressman, 1989, Fig. 46). Interpretation of the presence of basement rocks as a core in the Sylvanite anticline and beneath the Purcell anticlinorium is based in part on detailed analysis of major gravity highs beneath all or parts of those structures (M.D. Kleinkopf, oral commun., 1990).

East of the Rocky Mountain trench, a wedge of Belt strata was carried eastward about 50 miles on the Lewis thrust fault. Flat thrust faults connected by ramps and duplexes characterize much of this eastern terrane where a thin, almost flat, Belt section is above the Lewis thrust fault that cuts gently upward through Belt strata as it approaches the topographic surface. Duplex structures not present west of the Rocky Mountain trench have the basal surface of detachment as a floor and a folded Lewis thrust fault as a ceiling. These duplex structures, as shown on sections A-A′ and in more detail on section B-B′, are at the depths and widths identified by Fritts and Klipping (1987a, b) from seismic data. Tertiary extension on listric normal faults and from backsliding on Cretaceous thrust faults decreased rapidly eastward and is rare in the thin Belt wedge of Glacier National Park. Most of the Tertiary extension occurred just west of Glacier National Park where thick sedimentary rocks of Eocene and Oligocene age fill an extensional structure (the Kishenehn basin) above the Lewis thrust fault (section A-A′).

Retrodeformable sections at 1:250,000 scale were used to estimate total tectonic shortening and later extension along the line of section A-A′. The estimates are minimums because (1) the full zone of both thrusting and extension extends westward beyond the section and (2) an unknown amount of extensional backsliding on original thrust faults leaves doubts about the exact amount of both shortening and extension. The section indicates a minimum of 105 miles of eastward thrusting, mostly on the Moyie and Lewis thrust faults but not including a probable large throw on the Jumpoff Joe thrust fault. Extension is a minimum of about 50 miles, the bulk of which is represented by dilation (Harms and Price, 1992) in the area of the Newport fault.

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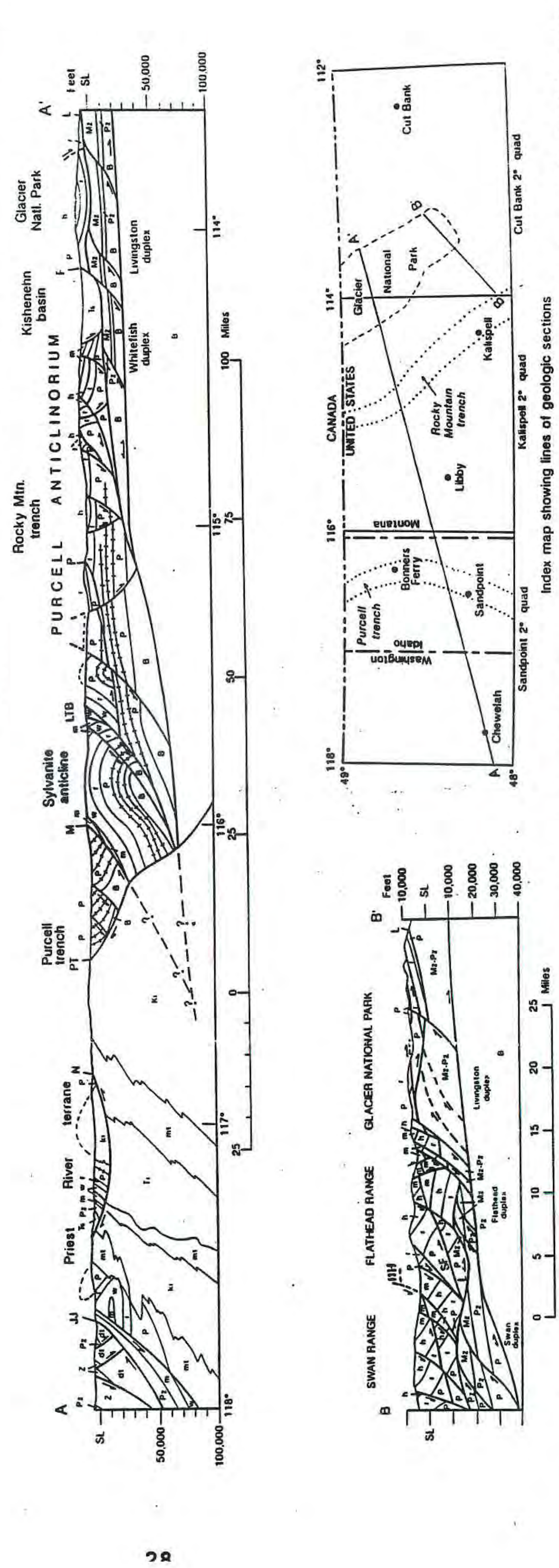

Figure 1. Geologic cross section from Chewelah, Washington, to northern Glacier National Park, Montana, with a supplemental section across the Swan and Flathead Ranges to southern Glacier National Park.

Ts, Tertiary sedimentary rocks; Ti, Tertiary intrusive rocks; Ki, Cretaceous intrusive rocks; Mz, Mesozoic sedimentary rocks; Pz, Paleozoic sedimentary rocks; Z, Late Proterozoic Windermere Supergroup; dt, Middle Proterozoic Deer Trail Group; m, w, h, r, p, units of Middle Proterozoic Belt Supergroup (Missoula Group, Wallace Formation, Helena Formation, Ravalli Group, and Prichard Formation and equivalents, respectively; highly generalized mafic sills in Prichard Formation shown by railroad track symbol); mt, metamorphic rocks of Priest River terrane; B, basement rocks undivided. Major faults: JJ, Jumpoff Joe thrust fault; N, Newport fault; PT, Purcell trench fault; M, Moyie thrust fault; LTB, Libby thrust belt; P, Pinkham thrust fault; F, Flathead fault; L, Lewis thrust fault. Double-headed arrows indicate backsliding on Cretaceous thrust faults during Tertiary extension.

PALEONTOLOGY OF THE BELT SUPERGROUP

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Strata of the Belt Supergroup contain a wide variety of fossils: *i)* microfossils occur in mudstone, chert, and carbonates; *ii)* megafossils occur as carbonaceous compressions, as impressions, as well as in other modes of preservation; and *iii)* evidence of microbial communities is preserved in stromatolites and microbial-mat impressions. In addition, a variety of structures, which once were considered to be of biogenic origin but now are considered to be nonbiogenic, occur in Belt strata. Examples of these structures are discussed herein. More thorough descriptions and discussions are presented in Horodyski (1989a; 1989b; 1993a; 1993b).

Microfossils in mudstone Mudstones in the Chamberlain Shale, Little Belt Mountains, Montana, contain a variety of organic-walled microfossils, which have been studied in maceration and in bedding-parallel thin sections (Horodyski, 1980). They include tubular filaments 0.5 to 13 μm wide, which have been referred to, or described as, *Archaeotrichion* spp., *Siphonophycus* spp., *Siphonophycus crassiusculum*, and *Siphonophycus beltensis*. The narrower filaments could be fossils of bacteria or cyanobacteria, while the filaments probably have cyanobacterial affinities. Also present, but rare, are filaments ranging from 17 to 40 μm wide and probably having cyanobacterial affinities. In addition, sphaeromorphic acritarchs ranging from 12 to 440 μm across occur in these mudstones, and some have been assigned to *Leiosphaeridia* and *Satka*. These acritarchs could represent cyanobacterial unicells and the outer envelope of colonial coccoid cyanobacteria, but others possibly represent the preserved vegetative or encystment stage of eukaryotic algae. This assemblage has some similarities to microfossils occurring in the Middle Proterozoic Roper Group, Australia (Peat et al., 1978); however, because microfossils rarely have been described from Middle Proterozoic siliciclastic strata, their biostratigraphic usefulness has not been established. Microfossils in siliciclastic strata of the Belt Supergroup also have been described from the Appekunny Formation (Horodyski, 1981) and Libby Formation (Kidder and Awramik, 1990).

Microfossils in chert Exquisitely preserved organic-walled microfossils occur in some Proterozoic cherts. However, although chert (including partially silicified stromatolites) occurs in several Belt formations, they have not yielded diverse, well preserved fossil assemblages. Silicification may have been a relatively late diagenetic event, occurring subsequent to considerable degradation of organic matter. Or, if silicification were early, conditions may not have been hypersaline, resulting in rapid and complete degradation.

Microfossils known from Belt cherts have morphologies that are consistent with cyanobacterial affinities. They have very little biostratigraphic or paleoecologic significance. However, the Newland cherts are of potential paleobiologic interest because they originated in offshore and apparently not hypersaline waters. There is the possibility that many Middle Proterozoic microfossil-bearing cherts originated in hypersaline settings, which obviously would preserve fossils of halophilic organisms (Horodyski and Donaldson, 1983). Therefore, microfossils in cherts of the Newland Limestone could provide us with a better indication of the nature of Middle Proterozoic life in normal marine waters (note that fossils in offshore mudstones also provide such information).

Microfossils in carbonates Microfossils are extremely rare in Middle Proterozoic carbonates. Neomorphism and the formation of microstylolitic seams generally obliterates organic-walled microstructures. One of the rare occurrences of Middle Proterozoic microfossils in limestone is in calcitic stromatolites in the Snowslip Formation, Glacier National Park, Montana (Horodyski, 1975, 1983, 1989a). They do not occur as organic-walled microstructures. Rather, they are defined by submicrometer- and micrometer-sized hematite particles, which probably represent oxidized iron sulfide. Strings of hematite particles and tubular cylinders 2.5 to 10 μm wide and composed of hematite particles occur in these stromatolites. Their morphology is consistent with organisms having bacterial or cyanobacterial affinities. Also occurring in these stromatolites are branched microstromatolites, which are 20 to 500 μm high, 10 to 200 μm wide, and composed of submicrometer- to micrometer-sized hematite particles (and locally sheet silicates or relatively clear calcite). These microstromatolites evidently were formed by rather small organisms, which probably were bacteria (this is based upon the thickness of laminae and the width of the structures). They were referred to *Frutexites* (not *Frutextites*).

Megafossils More than 90 years ago, Charles D. Walcott (1899, 1914a) described a variety of megascopic structures from the Belt Supergroup as metazoan body fossils and trace fossils. Some of these structures currently are considered to be pseudofossils, others are considered to be fossils, and the remainder are possible fossils. Of the specimens collected by Walcott, the coiled carbonaceous compression now known as *Grypania spiralis* is the best candidate for being a fossil of a megascopic organism. Evidently, only eleven specimens were collected by Walcott from the Greyson Shale at a locality he described as near the mouth of Deep Creek Canyon, Big Belt Mountains. Horodyski has obtained several hundred specimens from several localities that are very close to the mouth of this canyon. The ribbons comprising these coils are 0.5 to 1.7 mm wide, and most of the coils are 8 to 20 mm across. Some of the wider ribbons exhibit an indistinct mottled structure; however, it is unclear whether this is an original or a taphonomic feature. Similar material occurs in 1400 Ma strata in China (Du Rulin et al., 1986). Considering that animals rarely are preserved as carbonaceous compressions in Phanerozoic rocks, whereas algae and higher plants commonly are preserved in this manner, *Grypania* most likely represents a metaphyte (Walter et al., 1990) in the sense that metaphytes are multicellular photosynthetic organisms. However, there may not be a direct phylogenetic linkage between *Grypania* and Phanerozoic algae. *Grypania* could represent an evolutionary dead end.

In addition to *Grypania*, a variety of carbonaceous compressions were obtained at the Greyson localities. They include: sheet-like structures; millimeter-wide straight ribbons, which are more carbonaceous than *Grypania* and have closely spaced transverse markings; millimeter- to centimeter-wide disks; and submillimeter-wide, irregular carbonaceous ribbons having a bulbous end. All of these could represent megafossils having algal affinities.

The dolomite/shale tongue in the upper Chamberlain Shale, Little Belt Mountains, contains abundant, irregularly shaped, millimeter- to centimeter-wide carbonaceous compressions. They were described as *Beltina danai* and interpreted as fragmentary remains of a crustacean by Walcott (1899). They currently are considered to represent fragments of a microbial mat or possibly fragments of a sheet-like or ribbon-like megascopic alga (Horodyski, 1989a, 1993a, 1993b). At this same locality are rare, 1- to 5-cm-wide carbonaceous ribbons, which provisionally are interpreted as fossils of a megascopic alga.

Bedding surfaces of mudstones and muddy sandstones in the lower and middle Appekunny Formation, Glacier National Park, contain a variety of megascopic structures that may be biogenic. Most prominent are structures that resemble a string of beads (Horodyski, 1982a, 1989a, 1993a, 1993b). Each string is 1 to 7 cm long and consists of 3 to 30 beads, which range from 1 to 4 mm across and generally are separated by 1 to 3 mm. A prominent feature of these structures is the uniformity, within a string, of the size and spacing of the beads. Geometrically similar structures have been described from the 1.4 Ga Bangemall Group, Western Australia (Grey and Williams, 1990). The Montana and Australian material possibly represent fossils of a megascopic organism having algal affinities.

Also occurring in the Appekunny Formation in Glacier National Park are a variety of bedding-plane markings that possibly are biogenic. Among these are shallow oval depressions, which are 1 to 1.5 cm long and bordered by a submillimeter-high ridge. They may be fossils of a megascopic organism or a colony of microorganisms. Also present are 8-cm-wide coiled impressions, which could represent fossils of a megascopic organism--possibly an alga.

The wide variety of megafossils and possible megafossils that occur in the Belt Supergroup indicate that megascopic life was present and possibly abundant during Middle Proterozoic time. The Middle Proterozoic may have been a time of experimentation in megascopy. However, these organisms may not have a direct phylogenetic linkage with Proterozoic metaphytes or metazoans. They could represent evolutionary failures--a scenario that is supported by the scarcity of megafossils in Late Proterozoic strata.

<u>Stromatolites and microbial-mat impressions</u> Stromatolites are prominent in many Belt units, particularly in eastern exposures. They are particularly prominent in Glacier National Park, where they have been described by Horodyski (1975, 1976a, 1976b, 1977, 1983, 1985a, 1989b, and papers referenced therein). Also, see Horodyski (1993b) for further references to Belt stromatolites. Wrinkled bedding planes in red mudstones of the Snowslip Formation were interpreted to be the result of the desiccation or wrinkling of microbial mats in a tidal flat, supratidal flat, or coastal ponds (Horodyski, 1982b). If intertidal or supratidal, they would be of interest because they would demonstrate that microbial life could exist upon exposed surfaces during Proterozoic time. This could provide an indication of the degree to which an ultraviolet-light-filtering shield was developed in the atmosphere.

<u>Pseudofossils</u> A variety of pseudofossils (nonbiogenic structures that could be misinterpreted as biogenic) occur in Belt strata. Among the most interesting of these are "molar-tooth" structure and Walcott's *Newlandia* and associated structures. "Molar-tooth" structure is a microscopic structure named by Bauerman (1889), who described it from exposures near the Kootenai River as "resembling the markings on the molar tooth of an elephant." It is most prominent in the Siyeh, Helena, and Wallace formations, but it also occurs in the Snowslip, Shepard, and Chamberlain formations. "Molar-tooth" structure most commonly consists of horizontal and vertical sheets, which range from 0.2 to more than 5 mm thick and from 1 cm to several decimeters long, and ellipsoids generally a few millimeters across. In the Siyeh (Helena) Formation in Glacier National Park it typically is composed of pure, 5-20 μm sized, equant calcite crystals. They are interpreted as being abiogenic (Horodyski, 1976, 1983) and that they represent a variety of open-space structures including: fluid-evasion structures; desiccation cracks; syneresis cracks; primary voids; voids produced by gas generated by decomposition of organic matter; and voids produced by the decay of organic matter. "Molar-tooth" calcite also occurs filling displacive skeletal halite crystals in the Siyeh (Helena) Formation in Glacier National Park (Horodyski, 1989a). A neomorphic or replacement origin for the calcite is likely considering that crystal size is uniform regardless of distance from a bounding surface (Horodyski, 1989a).

Walcott (1914) described geometrically regular megascopic structures in the Newland Limestone near White Sulfur Springs as fossils of colonies of blue-green algae. They were sent to him by M. Collen, a Montana rancher. They include: the laminated structures *Newlandia frondosa, Newlandia lamellosa, Newlandia major*, and *Kinneyia simulans*; the semispheroidal laminated structure *Newlandia concentrica*; the tubular structures *Copperia tubiformis* and *Greysonia basaltica*; and the irregularly shaped tubular structure *Camasia spongia*. Gutstadt (1975) considered them to be "largely if not entirely of diagenetic or concretionary origin." Horodyski (1989a, 1993b) examined Walcott's specimens as well as specimens in outcrops near the Zieg Ranch and in the lower part of Deep Creek Canyon, and he interprets all of these structures to be of inorganic origin, being produced, in part, by fluid evasion and dissolution. At the Zieg Ranch locality, the distribution of some of these structures clearly is spatially related to crests of overlying megaripples. It appears that they originated by a combination of fluid evasion and dissolution (Horodyski, 1993b).

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MESOZOIC-CENOZOIC STRUCTURAL EVENTS AFFECTING THE BELT BASIN BETWEEN THE IDAHO BATHOLITH AND THE LEWIS AND CLARK LINE - IMPLICATIONS ON IDENTIFYING SYNDEPOSITIONAL BELT-AGE STRUCTURES

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INTRODUCTION

Structurally, the region of the Belt Basin between the Bitterroot lobe of the Idaho batholith (Cretaceous) and the Lewis and Clark Line is dominated by numerous strike-slip faults which trend approximately N70°W. The most prominent of these structures are the Osburn, Placer Creek, St. Joe, and Kelly Forks fault zones. In this region, the position of the Lewis and Clark line is interpreted to coincide with the trace of the Osburn fault zone which marks the abrupt change in trends of Mesozoic to Cenozoic fold and thrust faults from northerly to westerly. Recent detailed mapping and structural analysis in complexly deformed, lower and middle parts of the Belt Supergroup in the Moose Creek Buttes area and southeast part of the Coeur d'Alene Mining District, northern Idaho demonstrate a similar history of two episodes of complex major strike-slip movement and associated rotation of pre-existing fold and fault structures in late-Mesozoic and early Cenozoic time (Fig. 1). Differences in structural history and deformational style between the two areas reflect (1) proximity to the region of high heat flow of the Late Jurassic to Early Cretaceous magmatic arc, (2) forceful intrusion of Middle to Late Cretaceous or Eocene plutons, and (3) deformation mechanism(s) (translation or rotation and bending) associated with strike-slip faulting. The extreme complexity resulting from the overlap of superposed Mesozoic to Cenozoic crustal shortening and strike-slip displacement along the fault zones obscures recognition of possible basement structures related to formation of the Belt Basin. Models, which propose that these structures influenced depositional patterns and mark the main extensional axis in the western part of the Belt Basin (eg. Winston, 1986), need further evaluation, first requiring palinspastic reconstruction of Mesozoic and Cenozoic features.

MOOSE CREEK BUTTES AREA

In the Moose Creek Buttes area, positioned in the north-central border zone of the Idaho batholith-Bitterroot lobe, a 4.5 km.-thick allocthonous amphibolite facies (M_1) plate of Prichard Formation, Ravalli Group, Empire Formation, and lower part of the Wallace Formation is thrust (D_1) over greenschist facies (M_1) Wallace Formation and lower Missoula Group. Synkinematic intru

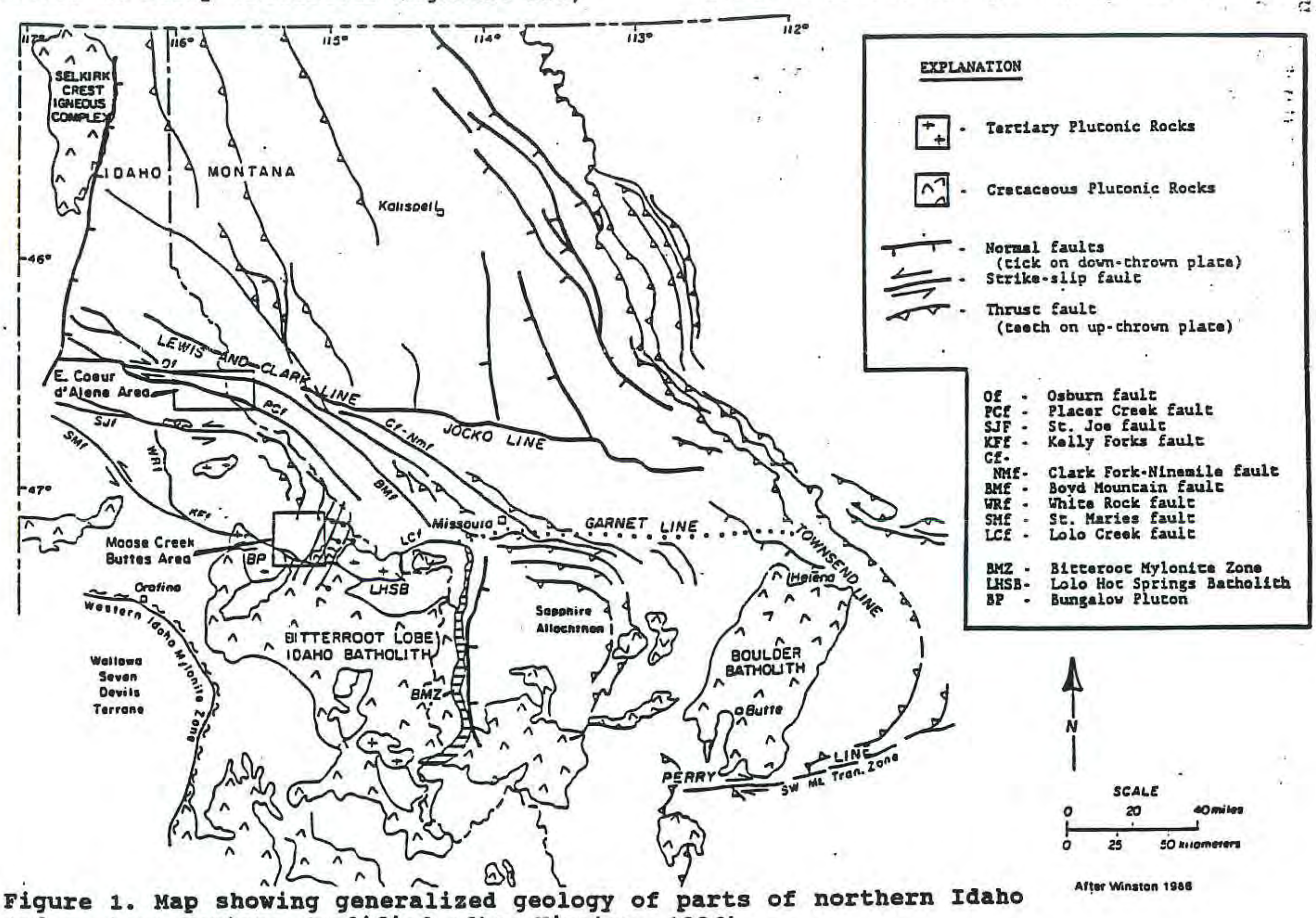

Figure 1. Map showing generalized geology of parts of northern Idaho and western Montana (modified after Winston, 1986).

sions of tonalite and anatectic granite into the thrust faults accompanied the D_1 deformation. The nature and configuration of the D_1 structures, M_1 metamorphic zones and synkinematic intrusions are typical of deep-seated, nappe/thrust tectonics developed by a fault-propagation mechanism in rocks positioned within a high heat flow regime (30°-50°C/km.) proximal to a developing magmatic arc. D_1 was probably caused by eastward thrusting with contractional tectonics along the subducting plate boundary of the North American Cordillera during the Late Jurassic to Early Cretaceous(?). The D_1 thrust faults and related folds have undergone extreme crustal shortenings (20%-80%) due to superposed deformations (D_2-D_4) and metamorphisms (M_2-M_4) attributed to forceful intrusions (ballooning) of Idaho batholith plutons during the Middle to Late Cretaceous. D_2 and D_3 deformed the D_1 nappe/thrust complex, like a stratigraphic pile, into upright/gently to steeply plunging fold systems which tighten toward respective causitive plutons. Steep plunges are present where D_3 or D_4 fold sytems were superposed upon steeply inclined D_2 fold limbs. Subsequent strong left-lateral faulting (D_5) (south to the east) along the Kelly Forks fault zone displaced the Idaho batholith >50 km to the east of related D_2-D_4 deformation and M_2-M_4 metamorphic aureoles. At the regional scale, wrench folding with the left-lateral D_5 movement along the Kelly Forks and St. Joe faults reflects 40°-50° of counter-clockwise rotation and distinct sigmoidal bending of D_1-D_4 structures as a result of D_5 eastward to northeastward compression direction. To accommodate part of the crustal shortening with this rotation, supracrustal-style northeast to east-directed imbricate thrusts developed at the east terminus of the St. Joe fault (Reid and others, 1981). The Kelly Forks fault zone was forcefully intruded by Eocene plutons including the Bungalow pluton and Lolo Hot Springs batholith, causing local crustal shortening (D_6) and shallow-level contact metamorphism. The north border zone of the Bungalow pluton was then displaced 10 km to the east by right-lateral movement on the rejuvenated Kelly Forks fault zone. Farther west, this displacement appears to be completely transferred along north-northwest-trending block(?) faults as shown by Hietanen (1968).

SOUTHEAST COEUR D'ALENE DISTRICT

Data collected as part of copper-silver exploration by Anaconda Minerals Company in middle parts of the Belt Supergroup in the southeast part of the Coeur d'Alene district between the Placer Creek and Osburn faults demonstrates a similar sequence of structural events. The earliest episode of deformation (D_1) generated north-trending, upright and gently plunging folds with an associated axial-plane cleavage under low-grade (sub-greenschist facies) metamorphic conditions. The style of D_1 deformation is typical of a fault-bend mechanism in a sub-greenschist facies environment positioned far from the high-heat flow regime of a magmatic arc and associated late-Jurassic to Cretaceous contractional tectonics. Major Coeur d'Alene veins likely formed at this time by metamorphic fluid migration into fault zones (high strain) flanking the axial trace of the D_1 Burke-Big Creek-Champion-Atlas anticline (Leach and others, 1988). A Cretaceous age for the Coeur d'Alene-type veins suggested by their structural setting is also supported by the recent work of Fleck and others, 1991; and Eaton and others, 1993). The D_1 folds and veins have been complexly segmented, rotated and folded along faults of the Lewis Clark line. The earliest post-D_1 movement along the Lewis and Clark line was left-lateral (south to the east) and produced a strike-slip fault duplex (plan view) between the Osburn and Placer Creek faults. The duplex consists of juxtaposed structural blocks (horses) within which pre-existing D_1 folds, cleavage, and vein structures have been rotated counter-clockwise into west-northwest to northeast trends. Hence, on the regional scale, the Lewis and Clark line marks the north boundary of >100 km of cumulative left-lateral displacement between the Kelly Forks and Osburn fault zones. To the east, much of this displacement was transferred along zones of northeast to east-directed crustal shortening (thrusting). The arrangement of horses and the position of thrust faults suggests that the duplexing process formed by (1) rotation of D_1 structures south of the Lewis and Clark line to nearly an east-west orientation, followed by (2) development of an eastward (forward(?)) propagating duplex. The second episode of strike-slip movement produced right-lateral displacement by either rejuvenation of old faults or by generation of new faults which cut across the left-lateral duplex. D_1 structures along the north margin of the Osburn fault were rotated clockwise and folded with about 24 km. of right-lateral offset (Hobbs and others, 1965). For many faults, right-lateral displacement was transferred to north-northwest-trending normal faults. Such movement suggests southeast-directed compression which overlapped extensional tectonism of the Eocene.

REGIONAL EFFECTS

On a regional scale, the two periods of strike-slip faulting caused profound re-orientation and offset of large-scale folds, related thrust faults, metamorphic zones and plutons produced by continentward (eastward) crustal shortening and magmatic arc plutonism that accompanied Late Jurassic to Early Cretaceous collision tectonism along the subducting plate boundary of the North American Cordillera. The earliest left-lateral strike-slip movements and associated areas of thrusting remain the most prominent features of these two superposed structural events. The arcuate pattern of deformed and offset structural features bounded by the Lewis and Clark line and central Montana transverse zone closely resembles that produced by indentation collision tectonics (eg. Himalayas). Whether this pattern is the result of (1) an indentor and second collision event, and/or (2) pre-existing basement structures in the Belt Basin providing a zone of weakness to accommodate a greater amount of crustal shortening remains unclear (Strayer, 1993). Regardless, the structural pattern requires that, south of the Lewis and Clark line, parts of the Belt Basin have been transported a great distance eastward in increments of 40-50 kms and accommodated significant strains across major strike-slip fault zones.

IMPLICATIONS

A comparison of the stratigraphic sequence and lithologic character of lower to middle Belt Supergroup units within the various fault blocks and north of the Lewis and Clark line suggests there are no significant differences. For the structural block between the St. Joe and Kelly Forks faults, the thicknesses of units of the middle to upper Belt Supergroup appear to be appreciably greater, than in structural blocks between or north of the Placer Creek and Osburn faults. This suggests that, prior to crustal shortening, the Belt Basin once extended a great distance to the south and southwest. The problem is how much of the thickening resulted from down-to-the-south Belt-age faulting and how much from juxtaposition of thicker western slabs moved eastward along the southern boundary of the Osburn fault.

Considering the magnitude and complexity of Phanerozoic movement history, the affect of the strike-slip faults on depositional patterns in the Belt Supergroup as syndepositional growth faults is obscure in the western part of the Belt Basin. Straightforward comparisons of stratigraphic thicknesses, facies relationships, directional depositional features, and spatial distribution of slump folds and breccia have been used to identify syndepositional faults and to infer their orientation and relative sense of displacement. Without palinspastic restoration of these features to their original relative positions and true orientation, the significance of the postulated Garnet, Jocko, and Perry lines is difficult to demonstrate in the western part of Belt Basin.

An alternative interpretation of the structure of Coeur d'Alene Mining District shows that, with restoration of D_1 folds, the Atlas Mine zone of soft sediment deformation and slump breccias in the St. Regis Formation would likely have had a pre-strike-slip orientation between N and N20°W. Comparison of the thicknesses based on complete drill intersections through the lower member of the Wallace Formation further indicates that the east block is down-thrown into a second-order(?) basin. On a more-regional scale, soft-sediment deformation localities and the "Wallace slope break", in the region south of the Lewis and Clark line, could be restored to a position far to the west-southwest into approximately a north-south alignment. These restorations may indicate that a north-northwest-oriented axis of intracratonic extension was important during formation of the western part of the Belt Basin.

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THE VENT COMPLEX OF THE SULLIVAN SEDIMENT-HOSTED Zn-Pb DEPOSIT, BRITISH COLUMBIA

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The vent complex of the Sullivan stratiform Zn-Pb deposit includes 1) massive pyrrhotite and pyrite-chlorite replacement bodies within the bedded ores; 2) underlying tourmalinite pipe consisting of breccia, altered and fragmental strata and disseminated or veinlet sulphides; and 3) albite-chlorite altered sedimentary rocks in the hangingwall (Fig. 1). Work to date on the vent complex includes geologic description (Hamilton and others, 1982), a study of the character of footwall veins (McAdam, 1978), whole-rock geochemistry (Shaw and others, 1993a, 1993b) and unpublished studies of alteration (summarized in Shaw and Hodgson, 1986).

Within the vent complex, the western massive sulphide portion of the deposit is underlain by an extensive, in places intensely developed, pyrrhotite ± quartz-Fe carbonate stringer network in tourmalinite (Fig. 2A; Leitch and Turner, 1992). This network probably was the major fluid upflow zone during formation of the sulphide body. The network ranges from wispy, irregular veinlets that appear to have been emplaced in unindurated sediments at relatively early stages, to planar veins with increasingly abundant quartz and carbonate that apparently formed within more indurated rocks at later stages in the development of the feeder zone. A crescent-shaped zone around the margins of the tourmalinite pipe is characterized by sphalerite and galena in the network veinlets (Fig. 2A), with associated tourmaline-destructive muscovite alteration (Fig. 2B); this may represent the site of late-stage fluid flow after sealing of the main central conduit by hydrothermal minerals (Fig. 3). Chlorite-dominant veins and alteration envelopes in the footwall and albite-chlorite-pyrite alteration in the hangingwall (Figs. 2B, 3) may be younger, related to hydrothermal flow during and after emplacement of mafic dykes and sills in the immediate footwall of the western part of the deposit (Turner and Leitch, 1992).

Quartz, and to a lesser extent sphalerite, carbonate, and cassiterite in the footwall network of pyrrhotite veins, contain abundant pseudosecondary or secondary fluid inclusions (Leitch, 1992a). Inclusions are not seen in wallrock detrital quartz grains, and the latter do not appear to be significantly recrystallized by post-ore greenschist facies metamorphism at 350-400°C (McMechan and Price, 1982; Ethier and others, 1976) and 2 kb (Edmunds, 1977). Fluids trapped in the inclusions are probably samples of the mineralizing fluids, in places compositionally diluted and/or thermally reset by metamorphism. The mineralizing fluids are saline 15-35 wt% $NaCl-CaCl_2$-?$MgCl_2$ brines, in Type 1 halite-saturated and Type 2 unsaturated inclusions; homogenization temperatures (Th) range from 200-300°C, but may reflect metamorphism. Dilution of these brines to <5 wt% NaCl, with mainly low but variable CO_2+CH_4, in several generations of secondary inclusions with Th = 200-350°C, is likely a metamorphic overprint. Low-T secondary inclusions (Th = 90-150°C) contain 3-20 wt% NaCl.

At the Sullivan deposit, bedded sulphides are facies equivalent to and separated from a lens of massive sulphides by an arcuate transition zone (TZ) above and around the eastern margin of the major footwall tourmalinite-pyrrhotite vent zone (Hamilton and others, 1982). Several minor minerals (cassiterite, stannite, freibergite, arsenopyrite, bismuthian boulangerite and jamesonite, bournonite, gudmundite, ?bismuthinite, Bi-Sb alloy) containing some or all the semi-metals As, Sb and Bi, or Sn or Ag, appear to be concentrated in the TZ in late-stage veins associated with muscovite alteration (Leitch, 1992b). Zoning of the trace metals appears, at least in part, to reflect this distribution of minor minerals; As and Sn are concentrated inboard, and Sb outboard, of the TZ (Fig. 4; Ransom and others, 1985); Sn shows a relation to massive pyrrhotite and As to the pyritic core (N. del Bel Belluz, pers. comm., 1993), but Ag may be largely controlled by the distribution of galena (Freeze, 1966).

Muscovite alteration is common in and around the Sullivan orebody and the small Stemwinder and North Star deposits 2-4 km to the south, forming extensive zones that crosscut stratigraphy and envelope the deposits up to 50 m above and below. Stratabound muscovite-rich rocks envelope the mineralized horizon as far away as Concentrator Hill, 5 km southeast of the deposit. Although muscovite is common in the greenschist facies rocks regionally, there appears to be an increase in 0.1 mm muscovite to the 30-50% range at the expense of first feldspar and then biotite near the deposits, coincident with an increase in sulphide.

Electron microprobe analyses show two varieties of chlorite in the Sullivan deposit: (1) a widely distributed Mg-rich variety, and (2) a peripheral Fe-rich variety present mainly in the bedded ores. Preliminary evaluation of microprobe data suggests zoning of mineral compositions about the vent zone that give several vectors of potential use to exploration: Fe in type (1) chlorites is highest in the center of the deposit, with Fe:(Fe+Mg) ratios to 0.49, but for type (2) Fe increases toward the fringe of the deposit, to ratios of 0.99. There is also a compositional zonation in carbonate, garnet and chlorite to higher Mn contents toward the fringes of the deposit. Plagioclase remaining in tourmalinized rocks is anorthite (up to An_{92}) compared to andesine (An_{38}) in regionally unaltered rocks, and albite (An_{0-2}) in albite-chlorite altered hangingwall or footwall rocks of the vent zone. Regionally distributed clinozoisite becomes more Fe-rich (epidote) near the deposits. The Fe/(Fe+Mg) ratio of tourmaline in the vent zone is about 0.42 in most of the fine felted tourmalinite, but ranges up to 0.70 in bluish schorl recrystallized by Moyie gabbro intrusions, to as low as 0.25 in amber dravite recrystallized by albite-chlorite alteration (cf. Ethier and Campbell, 1977). Traces of axinite, another borosilicate, have been discovered at the fringes of the tourmalinite zone in the footwall.

An extensive network of pyrrhotite veinlets, and scattered veins in places up to several meters thick, in the footwall tourmalinite pipe appear to be the main feeders to the Sullivan deposit (Leitch and Turner, 1992). The major veins are mineralogically similar to the Stemwinder vein. The high ratio of sulphide to gangue in the feeder veins and veinlets directly underlying the Sullivan deposit suggests a rapid mixing and quenching of metal-rich, possibly sulphide-bearing, fluids with seawater and a minimal input of detrital material to the basin during sulphide precipitation and deposition. An analogous situation is the Atlantis II Deep in the Red Sea, where a brine pool develops because high-salinity hydrothermal fluids exhaled into a bathymetric depression do not mix significantly with overlying seawater (Ramboz et al, 1988).

A brine pool model for formation of the bedded ores in the Sullivan deposit (Fig. 3) is supported by fluid inclusion data and by the presence of local coticule-type lithologies (R.L. Barnett, unpub. data) and rare occurrences of magnetite iron formation in the bedded ores. The extremely finely laminated nature of the major portion of the bedded ores, in which details of stratigraphy can be followed for up to 2 km (Hamilton and others, 1982), could reflect plume fallout within a brine pool or in an open marine setting. If the Type 1 and 2 fluids represent the mineralizing fluids, the observed 15-27 wt% salinities at Sullivan are like those in anhydrite veins underlying the Red Sea brine pools, and similar to the salinity of the brine pools themselves (13.5-25.6%: Ramboz and others, 1988). The source of the high salinities in the Sullivan fluid inclusions could be from evaporites.

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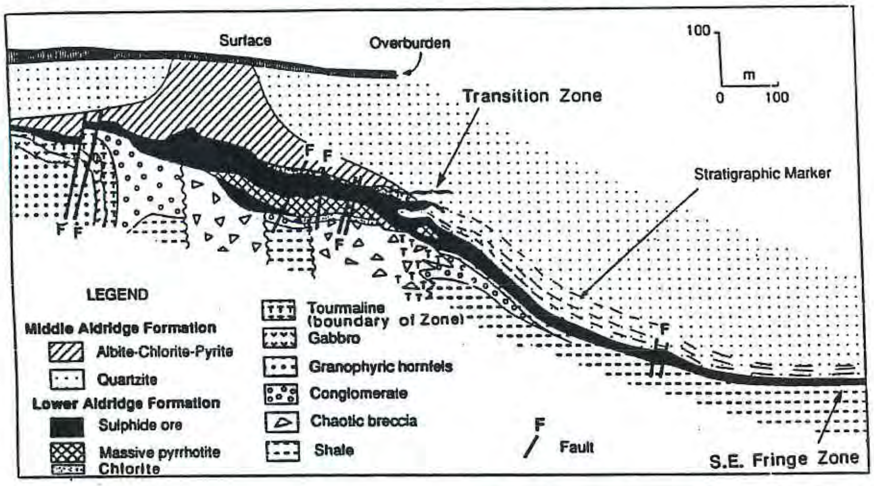
Figure 1. Geologic cross-section of the Sullivan deposit (modified from Hamilton and others, 1982).



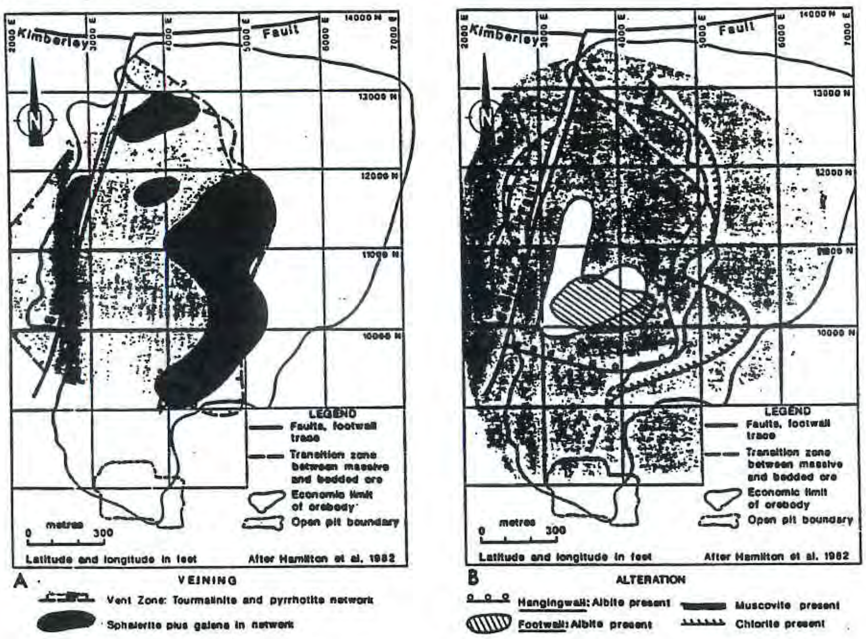


Figure 2. Plan views of the Sullivan deposit showing distribution of A) pyrrhotite-rich and sphalerite-galena-rich footwall mineralization, and B) albite-chlorite-pyrite alteration separately for hangingwall and footwall, and muscovite (shaded) and chlorite (inside hachured line) alteration from footwall through hangingwall (from Leitch and Turner, 1992).

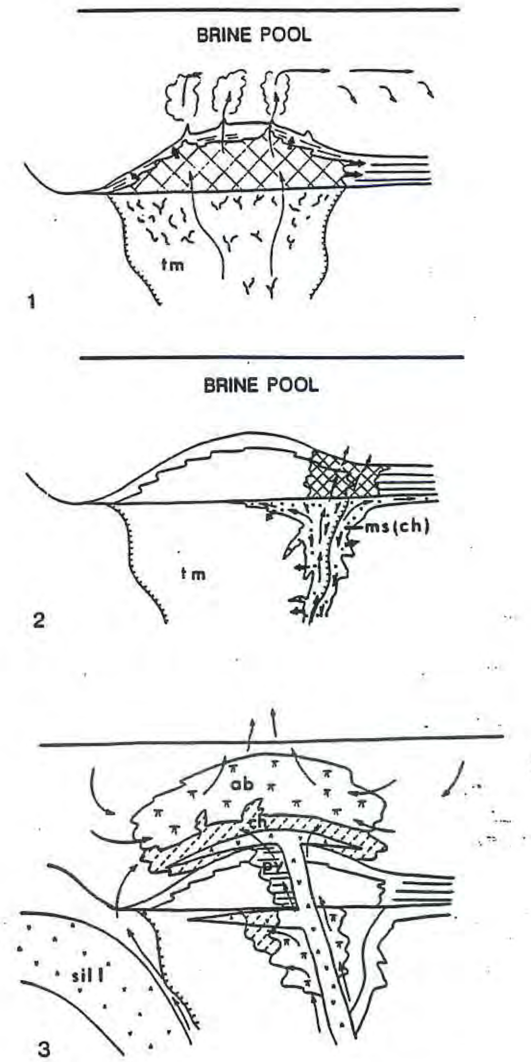


Figure 3. Postulated evolution of the hydrothermal system, Sullivan deposit (from Leitch and Turner, 1992). (1) Main stage hydrothermal flow results in sulphide sedimentation in a brine pool, with progressive replacement of sulphides by massive pyrrhotite (crosshatched), and underlying pyrrhotite network/tourmalinite (tm) alteration. (2) Late-stage hydrothermal flow is concentrated at the periphery of the main vent zone giving rise to pyrrhotite-sphalerite-galena-sulphosalt veinlets and replacements (crosshatched) in the transition zone, accompanied by muscovite and ?later chlorite. (3) Post-mineral fluid flow set up by magma body feeding Moyie sills/dykes and focused by vertical structures results in albite(ab)-chlorite(ch)-pyrite(py) alteration.

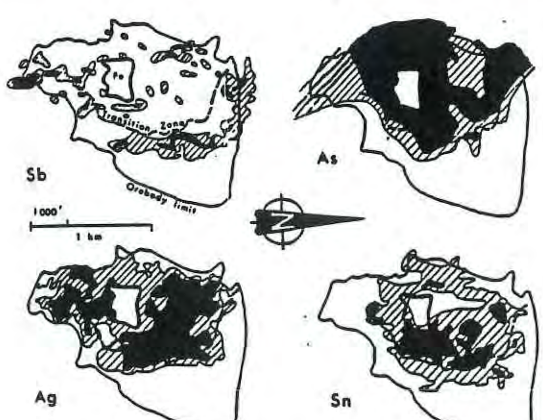


Figure 4. Distribution of minor metals in the Sullivan deposit (after Ransom and others, 1985 and Freeze, 1966). White, hachured, and dark areas indicate increasing levels of each element; Fe=central pyrite-chlorite-calcite zone.

STRUCTURE SECTIONS ACROSS SOUTH-CENTRAL BELT TERRANE, WEST-CENTRAL MONTANA.

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Structures in the south-central part of the outcrop belt of the Middle Proterozoic Belt Supergroup (Belt terrane, Fig. 1) are dominantly Late Cretaceous thrust faults, folds, and strike-slip faults; Tertiary normal faults are locally prominent, and they offset Cretaceous structures. Some deformation occurred prior to deposition of Middle Cambrian rocks, as indicated by the local truncation of some Belt units beneath the basal Paleozoic (Middle Cambrian) unconformity. Late Cretaceous to early Tertiary deformation and plutonism was more intense and at many places overprinted and obscured evidence for earlier deformation. Consequently, much of south-central Belt terrane is now characterized principally by tectonic and plutonic features of Late Cretaceous to early Tertiary age (Fig. 1). The principal tectonic features are the fold and thrust terranes of the Sapphire thrust plate, the Grasshopper thrust plate and its frontal fold and thrust zone, the Montana disturbed belt, a terrane of open folds, and steep faults of the northwest-trending Lewis and Clark line. The main plutonic masses are the Idaho and Boulder batholiths and related stocks. Most of these tectonic and plutonic features are crossed by the two structure sections in Figure 1 that are based mainly on the geologic map of the Butte 1° x 2° quadrangle (Wallace and others, 1987). The Grasshopper plate and its frontal zone (Ruppel and Lopez, 1984) are south of the cross section lines and are not discussed.

The Sapphire thrust plate is divided into two structural terranes: (1) a thrust-sheet terrane in the western to interior parts of the plate, and (2) a frontal imbricate terrane in the northern and eastern parts of the plate (Lidke and Wallace, 1988). The thrust-sheet terrane is characterized by stacked and folded thrust sheets consisting mostly of rocks of the Missoula Group and the middle Belt carbonate (Helena and Wallace Formations) of the Belt Supergroup. The frontal imbricate terrane is characterized by tight folds and numerous imbricate, listric thrusts in Paleozoic and Mesozoic rocks.

In the western part of the Sapphire thrust plate (section A-A'), stacked thrust sheets composed of rocks of the middle Belt carbonate and the Missoula Group are intruded by the eastern part of the Idaho batholith. The roots of these thrust sheets apparently were destroyed by the Idaho batholith. Evidence discussed by Hyndman (1980) indicates that the western part of Sapphire thrust plate moved across the region now occupied by the Idaho batholith. Certain structural and stratigraphic relations in the thrust-sheet stack suggest that some of the stacking resulted from out-of-sequence thrust faults that cut and restacked previously formed thrust sheets (Lidke and Wallace, 1988).

In the central part of section A-A', the thrust-sheet terrane is faulted against the frontal imbricate terrane along a thrust that places Missoula Group rocks above tightly folded and thrust-faulted Paleozoic and Mesozoic rocks. The frontal imbricate terrane apparently rides a décollement in the upper part of the Missoula Group. Paleozoic rocks along the leading thrust of the frontal terrane generally are faulted on broad anticlinally folded Missoula Group rocks present northeast of the Sapphire thrust plate in a terrane of open folds. About 10 mi south of section A-A', however, the leading thrust cuts across the west limb of the anticline and climbs section on the east limb to place Paleozoic rocks of the frontal imbricate terrane on broadly folded Paleozoic and Mesozoic rocks in the terrane of open folds that are a different facies and thickness than those of the frontal terrane. These relations across the leading thrust suggest that the frontal terrane was transported eastward several miles or more and cut and overrode preexisting folds in the terrane of open folds.

The eastern part of section A-A' crosses broad folds of the terrane of open folds and most of the Montana disturbed belt. The broad folds appear to deform Belt Supergroup and Paleozoic rocks equally; however, either the Belt Supergroup sequence was also very broadly folded or it was tilted westward prior to deposition of Middle Cambrian rocks, because an unconformity at the base of the Middle Cambrian rocks (base of Pz unit on cross sections) gradually truncates the Belt Supergroup sequence eastward across this area. In construction of section A-A', the unconformity was projected above the topographic profile and was used to broadly constrain stratigraphic relations in the lower plate of the disturbed belt. In the disturbed belt, broadly folded Ravalli Group rocks are thrust eastward on tightly folded Paleozoic and Mesozoic rocks. Lower Belt rocks (Prichard Formation equivalents) may be present at fairly shallow depth but are not exposed here or in nearby areas. Directly north of section A-A', thrust-faulted and folded Paleozoic and Mesozoic rocks of the lower plate overlie crystalline basement rocks (Mudge and others, 1982). Relations of the Montana disturbed belt to the terrane of open folds and to the Sapphire thrust plate are not known, but the open folds and Sapphire thrust plate appear to structurally overlie the projected basal décollement of the disturbed belt. In section A-A', the broad folds are interpreted as fault bend folds; lower Belt and basement rocks are inferred at depth in the lower plate of the disturbed belt and in the bottom of the broad folds; and, thrust faults of the disturbed belt are projected, as a basal décollement surface, beneath the broad folds and beneath the frontal part of the Sapphire thrust plate.

The eastern part of section A-A' also crosses high-angle, northwest-trending faults of the Lewis and Clark line and some of these faults have been intermittently active from Middle Proterozoic to Holocene time (Wallace and others, 1990). Although controversy continues about displacement directions and times of displacement along specific faults, a general consensus is that the major offsets were lateral and that most postdate thrust faulting and folding.

Section B-B' shows the thrust-sheet terrane of the Sapphire plate where thrust sheets of Missoula Group and middle Belt carbonate rocks are tightly folded but appear to be continuous with less folded sheets in section A-A'. In the thrust-sheet terrane (section B-B') middle Belt carbonate rocks are thrust on Paleozoic rocks along the folded Georgetown thrust (GT, Fig. 1). In turn the Paleozoic rocks below the Georgetown thrust are faulted on lowermost Missoula Group and middle Belt carbonate rocks along a fault that is also folded. A still deeper thrust, exposed as remnants between Cretaceous and Tertiary stocks, forms a basal thrust beneath the folded Belt and Paleozoic thrust sheets. Near the east end of the basal thrust, the overlying thrust sheets are folded into a recumbent anticline. Farther east (at the west edge of the frontal-imbricate terrane) tightly folded and faulted Paleozoic and Mesozoic rocks are mostly buried beneath Tertiary sedimentary and volcanic rocks and intruded by the Boulder batholith.

Thrust-fault and fold relations of the western, eastern, and southern parts of the Sapphire thrust plate have been obliterated or modified by Late Cretaceous and early Tertiary plutons, and locally have been modified by normal faults. In places, such as the area intruded by the plutonic suite in the Anaconda Range, the plutons appear to have been emplaced beneath a thrust-and-fold structural culmination and appear to have stoped older rocks and structures of the area. Schmidt and others (1990) have presented evidence that suggests the Boulder batholith was emplaced by pull-apart along an old fault zone. A number of mechanisms, possibly including crustal melting from tectonic loading, may have played a role in the formation and emplacement of the plutons. After emplacement, isostatic uplift of the plutonic regions and related normal faulting further modified structural and plutonic relations and probably also initiated the formation of Tertiary valleys, such as the one filled by Tertiary rocks near the center of section B-B'.

Relations shown in section A-A' suggest that the basal décollement of the Montana disturbed belt projects beneath the Sapphire thrust plate and may deform basement rocks. The décollement probably would join the roots of the Sapphire thrust plate in the region now occupied by the Idaho batholith. Relations shown in section B-B' are more complex, but suggest that any preexisting basement rocks or basal décollement would have been stoped by Cretaceous and Tertiary plutons in this area.

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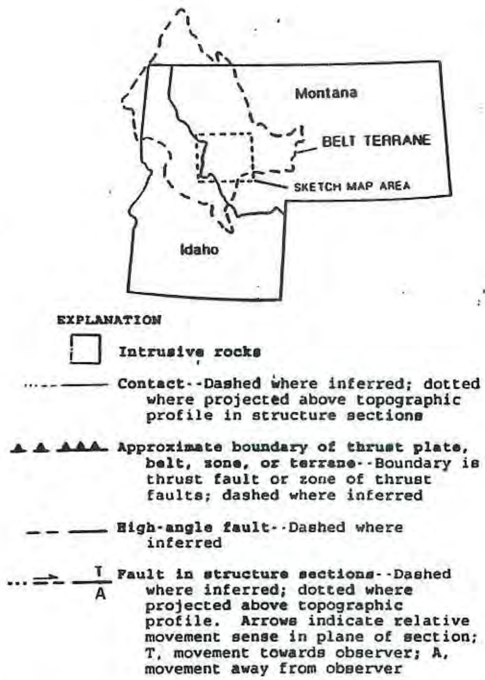

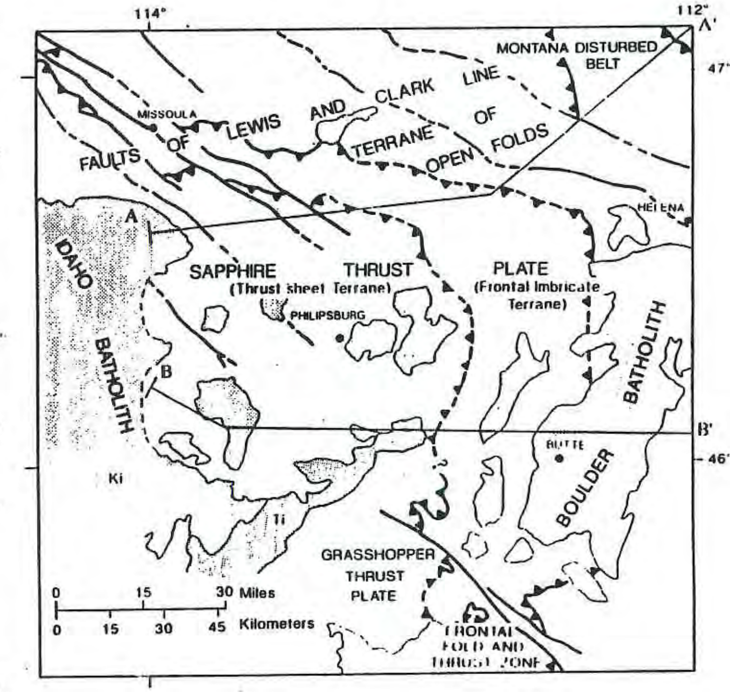

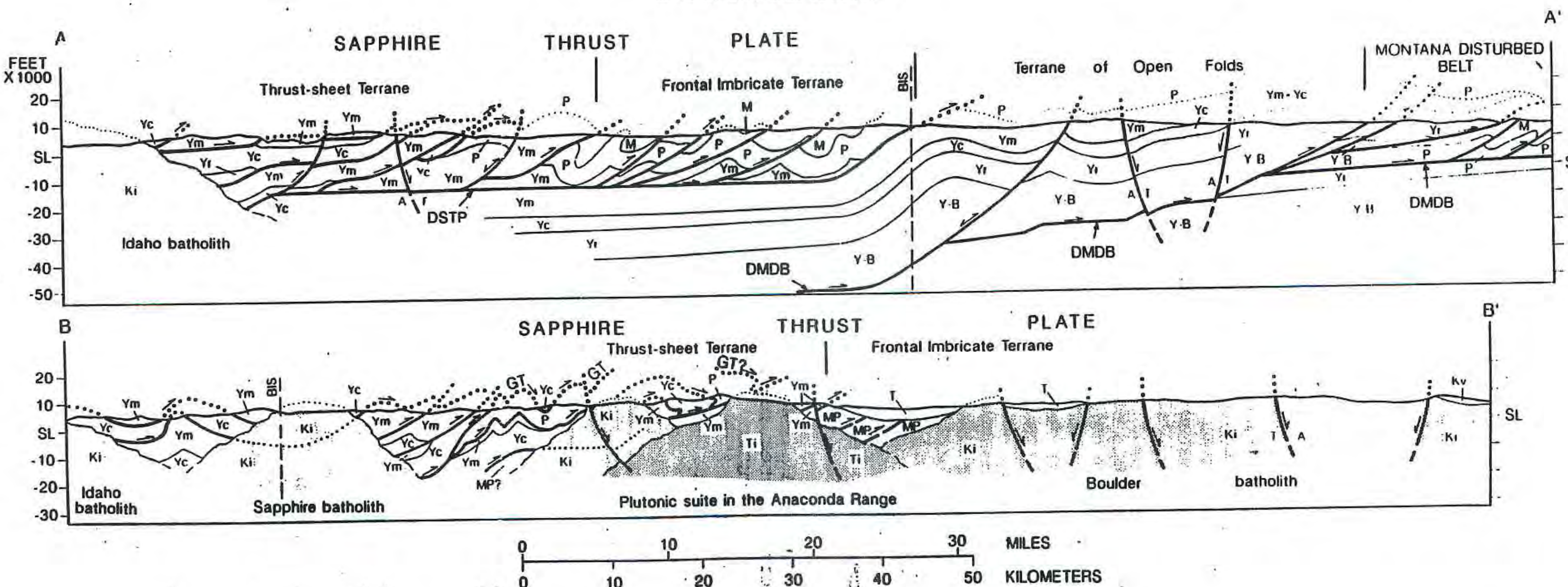

Figure 1. Index map, tectonic sketch map, and interpretive structure sections A-A' and B-B' across south-central Belt terrane, Montana. T, Tertiary sedimentary and volcanic rocks; Ti, Tertiary intrusive rocks; Kv, Cretaceous volcanic rocks; Ki, Cretaceous intrusive rocks; M, Mesozoic sedimentary rocks; MP, Mesozoic and Paleozoic sedimentary rocks undivided; P, Paleozoic sedimentary rocks; Ym, Missoula Group; Yc, middle Belt carbonate rocks (Helena and Wallace Formations); Yr, Ravalli Group; Y-B, lower Belt Supergroup rocks and (or) crystalline-basement rocks; GT, Georgetown thrust (section B-B'); DMDB, projected décollement of Montana disturbed belt; DSTP, projected décollement of Sapphire thrust plate; BIS, bend in section.

GEOLOGY OF THE SOUTHERN WHITEFISH RANGE, MONTANA

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INTRODUCTION

The Whitefish Range in northwestern Montana lies west of Glacier National Park between two fault-bounded valleys, the Rocky Mountain trench on the west and the Kishenehn basin on the east (Figure 1). The southern part of the range consists of allochthonous Middle Proterozoic Belt (Purcell) Supergroup rocks. Although there are Phanerozoic rocks which crop out in the northern part of the range (Harrison and others, 1992), such rocks are not present in the southern Whitefish Range. The oldest rocks exposed in the southern Whitefish Range are within the upper part of the Prichard Formation, located in the northwest corner of the mapped area (Figure 2). The Prichard is overlain by a relatively complete and uninterrupted stratigraphic succession through the McNamara Formation eastward across the range. Several major northwest-southeast trending faults cut the area which may be related to duplex structures associated with the Lewis thrust fault.

STRATIGRAPHY

Lithostratigraphic units within the southern Whitefish Range correlate well with those described in Glacier National Park (Whipple 1992; Whipple and others, 1984). Salient features of stratigraphic units are described below. The Mount Shields and Snowslip Formations have been divided into several informal members by Whipple and others (1984). These units have been identified in the southern Whitefish Range where they were mapped separately and are described below. They are, however, mapped together in figure 2 for clarity.

UNDIFFERENTIATED QUATERNARY DEPOSITS (QU): includes alluvium, colluvium, and glacial deposits.

MCNAMARA FORMATION (YM): Gray-green, fining-upward siltite/ argillite couplets (terminology of Winston, 1986) containing distinctive sili-cified mudchips and medium- to coarse-grained calcareous discontinuous layers. A thin red-bed unit is present in the lower part. Only the lowermost 800 feet (250 m) is exposed.

BONNER QUARTZITE (YBO): Coarse- to fine-grained pale red to brick red feldspathic arenite and siltite beds which commonly occur as fining-upward units (up to 3 meters thick). Cross-bedding, ripple marks, and large channels are typical of the coarse-grained layers. The unit is approximately 800 feet (245 m) thick, with the lower 100 feet (30 m) consisting of medium- to coarse-grained gray-green to green siltite and feldspathic arenite. The lower section is also commonly cross-stratified and channelled.

MOUNT SHIELDS FORMATION (YMS): Rocks of the Mount Shields Formation have an exposed thickness of about 2,650 feet (810 m). Informal member designations after Whipple 1992; Whipple and others 1984.

Members 4 and 5 - mapped together in the southern Whitefish Range due to the poor exposure of uppermost 30 feet (9 m) which includes thinly laminated, dark argillite of member 5. Member 4 is composed of gray-green siltite and argillite, locally dolomitic. Base of member 4 is placed at the first bed which contains carbonate cement. Total thickness of members 4 and 5 is approximately 530 feet (160 m).

Member 3 - Purple to brick-red siltite and argillite fining-upward couplets containing distinctive salt casts (up to 2 cm across). Salt cast abundance decreases upward in the unit. The base of the unit is placed at the first bed above the stromatolites of member 2. Approximately 1,475 feet (450 m) thick.

Members 1 and 2 - In the southern Whitefish Range, the two members are indistinguishable from one another. Thinly laminated maroon to pale-purple argillite, brick red siltite, and very fine-grained arenite. Arenite content increases upward. The top of the unit is characterized by oolitic calcareous layers which contain small (5-25 cm high) stromatolite beds. Total thickness is approximately 550 feet (165 m).

SHEPARD FORMATION (YSH): Light-gray to gray-green dolomitic and pyritic siltite and argillite, weathers light brown. Includes thin beds of dolomite and limestone. Rock types typical of the Shepard Formation in surrounding areas, containing oolites, onkolites, and stromatolites, were not found in the study area due to the limited exposure and perhaps absence of such features. The base of the unit is placed at the bottom of the lowermost carbonate bed above the Snowslip Formation. Maximum thickness is 750 feet (230 m).

PURCELL LAVA (YPB): Black- to dark green- weathering amygdaloidal and vesicular basaltic lava flows with red sedimentary interflow and brecciated units, entirely contained within the upper part of the Snowslip Formation as described by Whipple and others (1984). Plagioclase phenocrysts up to 2 inches long are present. The unit thins from approximately 150 feet (45 m) in the northwest to 10 feet (3 m) in the southeast.

SNOWSLIP FORMATION (YSN): Member 6 - Light-green tabular silt beds and gray-green even fining-upward couplets of siltite and argillite. Maximum thickness is 115 feet (35 m), not including the Purcell Basalt which it completely encloses or overlies in the map area.

Members 1 through 5 - Informal members after Whipple and others (1984). Grey-green to green and pale-purple to brick-red fining-upward couples and couplets (terminology of Winston, 1986). Unit is undivided due to inadequate exposure. Twenty to fifty centimeters thick arenite beds increase in abundance upward in the sequence. One to three meter fining-upward units have been recognized, especially in the uppermost part of the unit: Thin discontinuous stromatolite layers (up to 1 meter thick) are common in the lower portion of the section. Mudchip intraclasts, ripple marks, shrinkage cracks, and fluid-escape structures are locally abundant throughout the unit. Total thickness of members 1-5 is approximately 1,700 feet (510 m).

HELENA FORMATION (YH): The Helena Formation is divided into three parts. The upper part consists of laminated oolitic limestone and black argillite interbedded with laminated dolomite and very fine arenite that is equivalent to the middle Wallace Formation (Whipple and others, 1984). The base is a 100 foot (30 m) thick stromatolitic unit composed primarily of Baicalia stromatolites. The thick middle part of the unit is dominated by beds of the molar-tooth dolomite. The lower part of the unit consists of interlaminated gray calcareous siltite and green argillite at the base, molar-tooth dolomite in the middle, and thickly bedded gray limestone at the top. The base of the Helena Formation is placed at the first discernable dolomite bed above the Empire Formation. The maximum thickness of the Helena Formation is 3000 feet (915 m).

EMPIRE FORMATION (YE): Gray-green pinch and swell fining-upward couplets, tabular green siltite and argillite beds and tabular white to pale green and pale red coarse siltite beds. Most layers are at least somewhat calcareous or dolomitic. Pyrite (up to 3 cm across) is common in several beds within the unit. The upper part of the formation consists of green laminated siltite and argillite disrupted by fluid-escape structures and large (up to 25 cm across) horizontal calcareous pods which commonly weather to give the outcrop a "vuggy" appearance. These calcareous pods along with the dolomite beds found in the basal Helena Formation are the culmination of the increasing carbonate content upward in the unit. The Empire Formation is approximately 800 feet (250 m) thick.

GRINNELL FORMATION (YGL): The Grinnell Formation is mapped in the southern Whitefish Range due to similarities with correlative rocks in Glacier National Park as described by Whipple (1992). This unit has also been mapped as Spokane Formation by Harrison and others (1992) who arbitrarily limit use of the term Grinnell Formation to Glacier National Park. The upper 395 feet (120 m) of the unit consists of pale purple to reddish gray siltite interbedded with argillite, dark green siltite, and pale green argillite. Layers of gray to white fine-grained quartz arenite are common. Desiccation cracks, fluid-escape structures, mudchip breccia, and armored mudballs are common. The lower part, comprising most of the formation thickness, consists of green gray siltite and pale purple argillite in even fining-upward couplets (terminology of Winston, 1986). Mudchip breccia and disrupted laminae are common. Carbonate cement is locally abundant, especially in the lowermost

295 feet (90 m). Total thickness is approximately 3,750 ft (1,140 m).

BURKE (?) FORMATION (YB(?)): Laminated to tabular beds of olive green, gray, and buff siltite and fine-grained arenite. Limonite staining, possibly after magnetite, is very common. Contact with the underlying Prichard Formation is interpreted as gradational. Total thickness is approximately 850 ft (260 m).

PRICHARD FORMATION (YP): Only the upper part of the Prichard Formation is present and it consists of interlaminated, rust-weathering, black argillite and gray siltite. Some small-scale cross-lamination is present. Several manganiferous limestone pods and layers contain small stromatolites. The base is not exposed. The maximum exposed thickness in the map area is approximately 800 ft (245 m).

Stromatolite zones in the upper part of the Helena Formation, Snowslip Formation member 2, and Mount Shields Formation member 2 provide excellent marker beds for stratigraphic correlation of Belt rocks in the Whitefish Range and adjacent areas (Hanson, 1988; Lundblad, 1988; Reese, 1988; Whipple, 1992). These horizons have also been identified in Glacier National Park.

The Middle Proterozoic stra-tigraphic sequence is punctuated by the Purcell Lava, which is located within the upper Snowslip Formation. Maximum total thickness of the Purcell Lava in the Southern Whitefish Range is approximately 150 feet (45 meters) in the northernmost exposure in figure 2, but it thins southeastward to approximately 10 feet (3 meters), where exposure ends. The Purcell Lava changes stratigraphic position within the upper pat of the Snowslip Formation across the mapped area. In the northwest, it is located completely within member 5 of the Snowslip Formation. Near the center of the map the Purcell Lava lies between members 5 and 6, and in the southeast it is entirely within Snowslip member 6.

Major Structures

The southern part of the Whitefish Range is part of a large, open, north-northwest-trending monocline which may be related to the development of the Whitefish Duplex structure associated with the Lewis Thrust Fault. Several major northwest-trending faults are inferred on the basis of repeated units, attitude changes, offset geologic contacts, increased cleavage intensity, and rapid thickness changes. Some of these faults may have accommodated both Laramide compression and later (post-Laramide) extension (Whipple and Harrison, 1987).

Northwest-trending normal faults bound the Rocky Mountain Trench along and near the western edge of the mapped area and bound the North Fork Flathead River valley along and near the eastern edge of the mapped area (Whipple and Harrison, 1987; Harrison and others, 1992; Whipple 1992). The Whitefish Divide, which includes Werner Peak (see Figure 2), is nearly coincident with a northwest-trending, open, chevron-type anticlinal fold which plunges gently to the southeast.

The Whitefish Divide anticline may be the result of ramping along the fault which is directly adjacent to the east. This fault has been identified as the southern part of the Whitefish Thrust Fault (Harrison and others, 1992). The trace of this fault was not directly observed in the field, but evidence for its existence includes stratigraphic thickness anomalies in the Empire and Grinnell formations, minor but abrupt changes in attitudes across the fault trace, and a sharp increase in cleavage near the fault trace. This fold, and to some extent the corresponding fault, diffuse into several open, southeast-plunging folds in the southern part of the mapped area near the Big Mountain ski area.

Two major normal faults cut the eastern part of the mapped area. The South Fork Fault bounds the east side of the Flathead River valley and has down-to-the-west offset of approximately 5000 feet (1500 m) (Whipple, 1992). This fault juxtaposes the Mount Shields Formation and the Snowslip Formation. An antithetic normal fault bounds the western margin of the Flathead River valley. A synthetic large-displacement normal fault is located to the west of the South Fork Fault. Juxtaposition of Bonner Quartzite and Shepard Formation strata indicate that this fault has offset of approximately 3300 feet (1000 m). Exposures of quartz-cemented breccia occur along the inferred trace of the fault north of its exposure trace.

Minor folds and faults, particularly in the western part of the range, have northwest-southeast trends similar to the trends of larger structures. Well-developed joints and spaced cleavage are also present and are generally northwest-trending and steeply northeast-dipping features. The main Laramide detachment beneath the Whitefish Range is inferred at a depth of approximately 20,000 feet (6,100 meters) based on seismic data, with the possible presence of Paleozoic and Mesozoic rocks beneath the southern Whitefish Range within the Whitefish Duplex structure (Fritts and Klipping, 1987a, 1987b).

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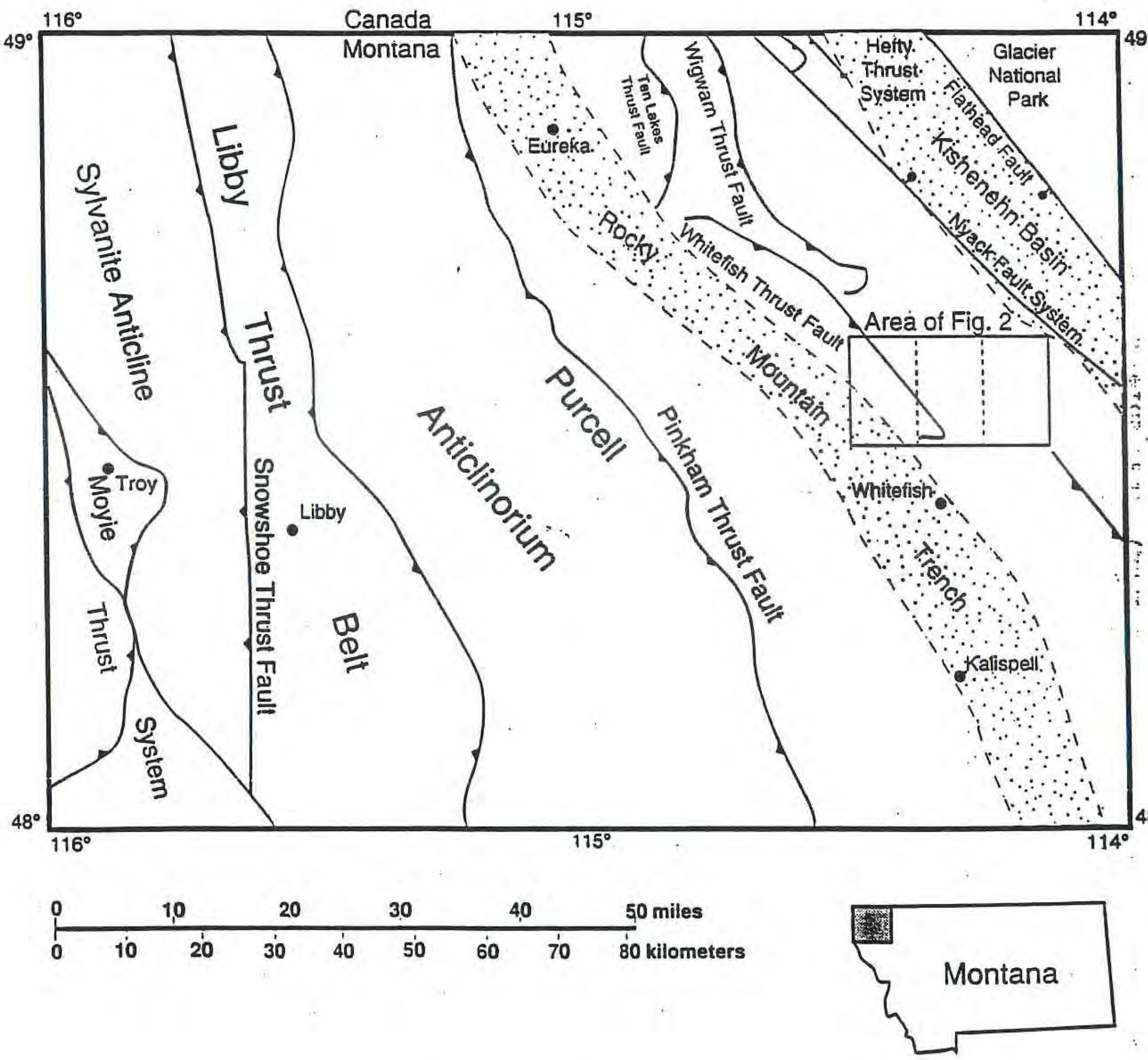

Figure 1. Regional Geologic and Index Map of Northwestern Montana.
Generalized structure map of northwestern Montana showing the location and regional structural setting of the southern Whitefish Range (modified from Harrison and others, 1992). Major fold axes (not shown) generally trend north-south in this region. The outlined area delineates the map in figure 2 and the Werner Peak (west), Skookoleel Creek (center), and Huckleberry Mountain (east) 7.5 minute quadrangles originally mapped at 1:24,000.

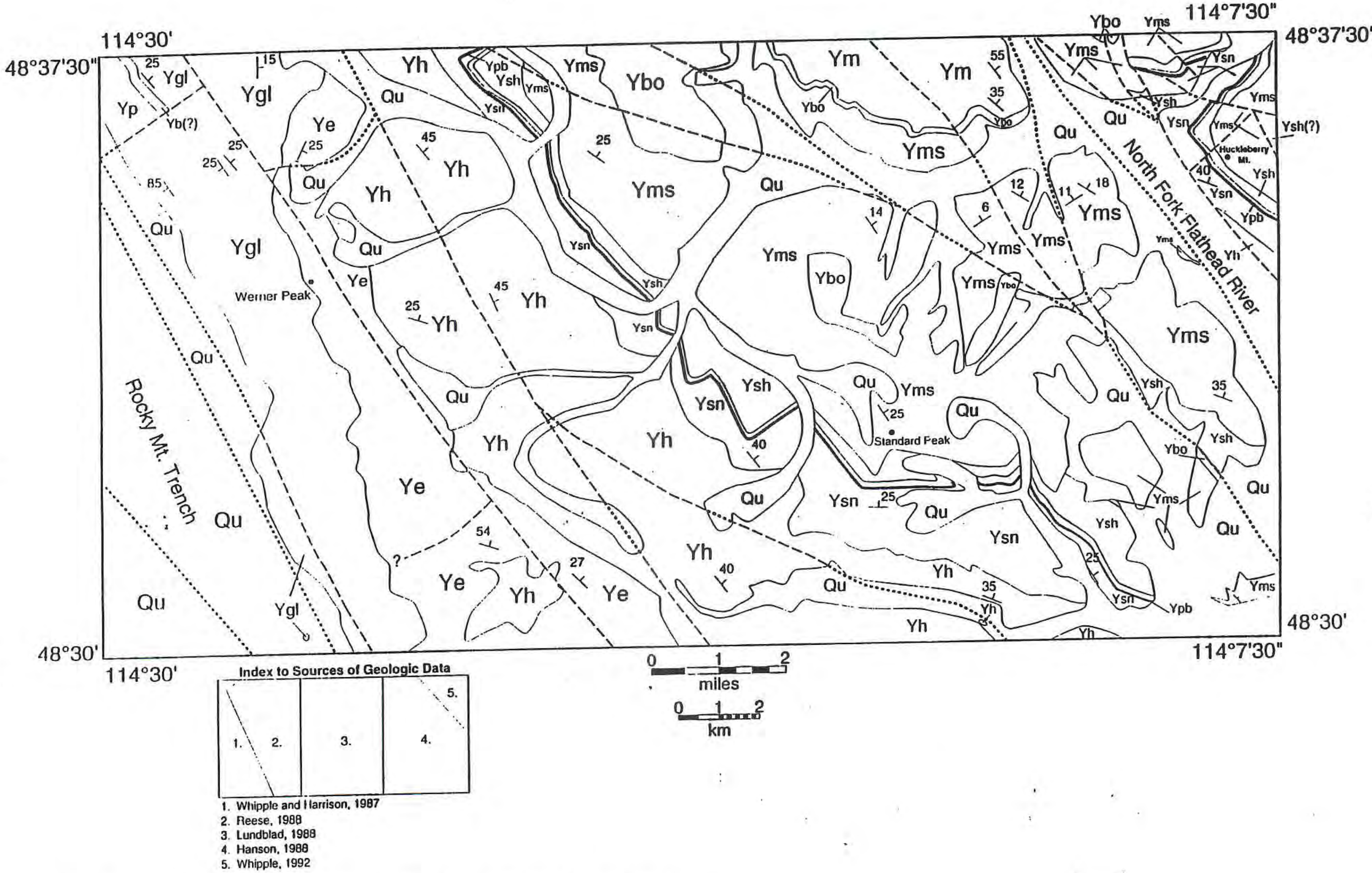

Figure 2. Geologic Map of the Southern Whitefish Range.

Yp=Prichard Fm., Yb(?)=Burke (?) Fm., Ygl=Grinnell Fm., Ye=Empire Fm., Yh=Helena Fm., Ysn= Snowslip Fm.(undivided), Ypb=Purcell Lava, Ysh=Shepard Fm., Yms=Mount Shields Fm. (undivided), Ybo=Bonner Quartzite, Ym=McNamara Fm., u=Quaternary deposits (undifferentiated). Dashed lines represent faults, dotted where covered. Representative strike and dips are shown

CROSS-SECTION OF THE ROCKY MOUNTAIN THRUST BELT FROM CHOTEAU TO PLAINS, MONTANA: IMPLICATIONS FOR THE GEOMETRY OF THE EASTERN MARGIN OF THE BELT BASIN

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INTRODUCTION

A geologic cross-section of the Rocky Mountain thrust belt between Choteau and Plains, Montana, (fig. 1) implies that the eastern margin of the Middle Proterozoic Belt basin was translated approximately 115 km northeast on the Hoadley, Lewis, and associated thrust faults of the Montana Disturbed Belt. The thrust mass was extended by 20 to 25 km along post-thrusting normal faults.

CROSS SECTION

The cross-section was constructed at 1:100,000 or greater scales, compiled at 1:250,000, and further reduced for this publication. It is constrained by surface geologic mapping, stratigraphic and structural data, and by seismic reflection lines and borehole information. The western part of the section is constrained by a number of seismic reflection lines that cross the Flathead Indian Reservation. These lines clearly show prominent, continuous reflectors that can be tied into surface outcrops of mafic sills in the Prichard Formation, and show the basement-cover contact as a structural discontinuity at a depth of several km (Sheriff, 1986). The eastern part of the line is in the Montana Disturbed Belt, and is modified from a cross-section presented by Mudge and others (1982), which was constructed with the use of detailed geologic mapping, surface stratigraphic and structural control, and borehole data.

STRUCTURAL ELEMENTS

From northeast to southwest, the key structural elements of the cross-section are:

1)Montana Disturbed Belt: This set of imbricated, west-dipping thrust plates of Paleozoic and Mesozoic rocks has a minimum cumulative horizontal displacement of 25 km. The eastern half involves mostly Upper Paleozoic and Mesozoic strata, while the western half includes Lower Paleozoic rocks.

2)Lewis Thrust Plate: Along this section, the Lewis thrust is as much as 3 km thick, and has a minimum width of 40 km. It carries a thin section of Belt Supergroup rocks. The leading edge is truncated by erosion, but plunge projections at the north and south ends of the Montana Disturbed Belt suggested that the plate may have continued 20 to 30 km farther east, forming the roof of a thrust duplex. It's original width may have been 70 km. The cross-section shows the west end of the Lewis Plate at the position of the Flathead fault. This interpretation is constrained

by the ramp/flat geometry of the overlaying Hoadley Plate.

3)Hoadley plate: This massive thrust plate contains the Continental Divide syncline and Purcell anticlinourium. The Continental Divide syncline is due to the ramp/flat geometry of the eastern plate of the Hoadley plate. The west limb of the syncline over-lies a hanging wall ramp through the lower Belt Supergroup., while the east limb overlies a hanging wall flat in the Missoula Group. The cross-section shows the Hoadley plate displaced about 12 km over the Lewis plate. The minimum total northeastward horizontal displacement of the Hoadley plate, given by summing the minimum displacement of the Montana Disturbed Belt, width of the Lewis plate, and overlap of the Hoadley over the Lewis Plate is 77 to 107 km.

The Purcell anticlinorium resulted from the tectonic inversion of the Belt basin. The Belt Super-group rapidly thickens westward in the Hoadley plate, across the west limb of the Continental Divide syncline and Swan Range, from about 3 km to greater than 20 km. All units thicken, but most of the increase shown on the section line is due to thickening of the lower Belt from a few hundred meters to many kilometers. The east dip of the Belt Supergroup is accentuated by rotation along the Mission, Swan, and Flathead faults. The west part of the Purcell anticlinorium contains two large kink-folds, the Camas Prairie and Plains anticlines. The east limbs appear due to hanging wall ramps in the lower part of the Prichard Formation. The west limb of the Plains anticline appears to drape a major footwall ramp along the western margin of the Purcell anticlinorium. A large downward step in the level of decollement is necessary to accommodate the thick stratigraphic section that is preserved in the Lewis plate in the Coeur d'Alene Mountains. The cross-section suggests that the Prichard Formation is 15 km thick as it descends the ramp.

CROSS-SECTION RESTORATION

The cross-section was graphically restored by reversing the displacement on the extensional faults, restoring the hanging wall truncation of the Hoadley plate to match the footwall truncation on the Lewis plate, then adjusting the Belt Supergroup to a Cambrian sea level datum. The effect is to invert the Purcell anticlinorium, restoring the Belt to a deep basin configuration with a steep eastern margin (fig. 1B). The hinge zone of the

basin now forms the Swan Range and western limb of the Continental Divide syncline. A shelf region bordered the hinge on the east, and now forms the Lewis plate. The deep basin was to the west of the modern central Swan Range.

How far removed the Belt rocks are from their depositional sites is determined by matching the geometry of the allocthonous and autochthonous rocks along the section line. The allochthonous eastern basin margin matches the footwall step at Plains, indicating 115 km of thrust displacement. This estimate is consistent with the minimum 77 to 107 km horizontal displacement for the Lewis plate estimated above.

The restored cross-section shows a deepening of the basin approximately 70 km west of the basin margin. The palinspastic position of this step corresponds to the present site of the northern margin of the Idaho batholith. The batholith occupies a regional synform that may represent the draping of allochthonous rocks.

REGIONAL CONSIDERATIONS

A key to estimating the displacement of the Belt basin is to measure the map distance, in the transport direction, between the Continental Divide syncline and the first major west-dipping homocline in the interior of the thrust system. The distance between these features decreases systematically to the southeast (Sears, 1988); they merge near Helena. The geometry of the system implies approximately 30 degrees of clockwise thrust rotation of this part of the Belt basin, about a vertical pole of rotation near Helena. If so, the Belt rocks become more allochthonous northward into Canada.

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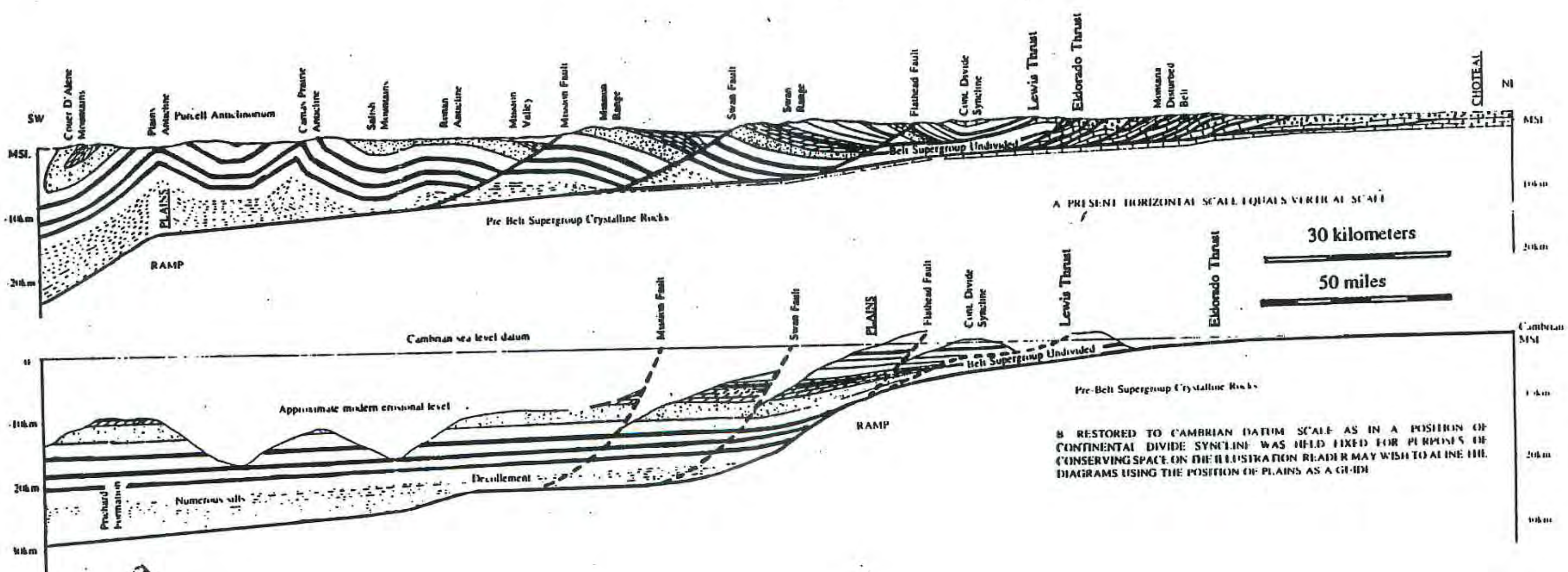

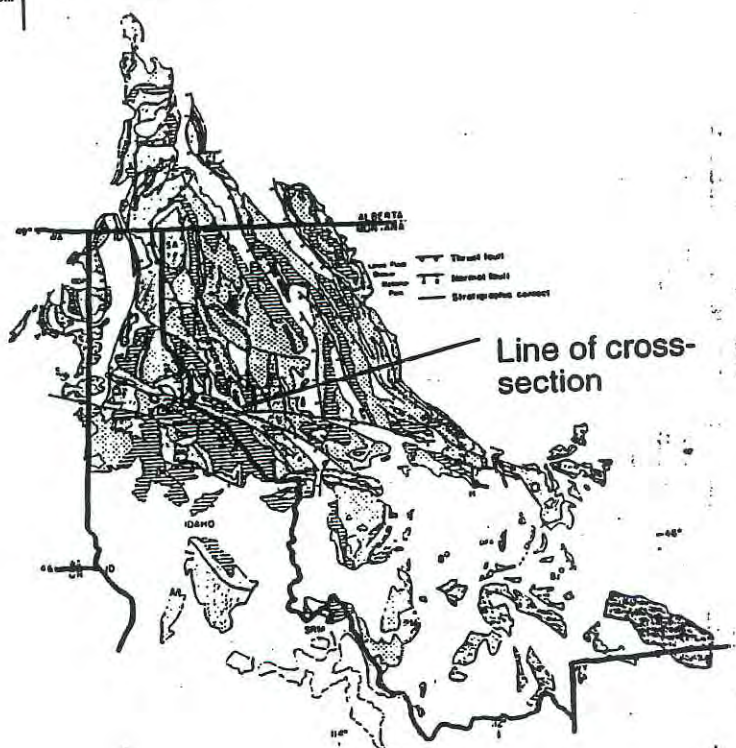

Figure 1 and 1b. Geologic cross-section and restored cross-section, along a line from Choteau to Plains. Dashes = lower part of Prichard Formation; railroad track pattern = mafic sills; small dots = Ravalli Group; slanted bricks = middle Belt carbonate; straight bricks = Paleozoic rocks; large dots = Mesozoic rocks.

MODELS FOR TOURMALINITE FORMATION IN THE MIDDLE PROTEROZOIC BELT AND PURCELL SUPERGROUPS AND THEIR EXPLORATION SIGNIFICANCE

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Tourmalinites are distinctive stratabound rocks that contain 20 percent or more tourmaline by volume (Slack, 1982; Slack and others, 1984). They occur in many Proterozoic and early Paleozoic terranes, and are well documented in association with a variety of stratabound mineral deposits including those containing Cu, Pb, Zn, Ag, Au, W, Co, and U (see Slack and others, 1993, for references). Since the pioneering work of Ethier and Campbell (1977) on tourmaline-rich rocks and tourmalinites in the Middle Proterozoic Belt and Purcell Supergroups, exploration efforts have focused on these unusual rocks as prospecting guides.

Nearly all known Belt-Purcell tourmalinites are in the Aldridge and Prichard Formations (e.g., Ethier and Campbell, 1977; Beaty and others, 1988). Two major types are recognized, discordant and stratiform. The discordant tourmalinites form bodies several meters to hundreds of meters in diameter that clearly crosscut stratigraphy (but nevertheless are broadly stratabound). Examples include the tourmalinized feeder zone of the stratiform Sullivan Pb-Zn-Ag deposit, British Columbia (Shaw and Hodgson, 1980; Hamilton and others, 1982), and discordant tourmalinites at prospects such as Goatfell, British Columbia, and Morning Glory, Montana. Some of these discordant tourmalinites are associated with Moyie-type gabbroic intrusions, coarse fragmental rocks or conglomerates, and stratabound albite-rich rocks or albitites.

The stratiform tourmalinites are centimeters to meters in thickness and locally may be traced along strike for tens to hundreds of meters. Many of these tourmalinites show primary sedimentary structures including fine-scale laminations, graded beds, cross laminations, flame structures, and tourmalinized rip-up clasts in a tourmaline-free matrix. A few localities (e.g., Fork, Montana) contain thin (1-2 mm) veinlets of quartz + tourmaline + ilmenite ± pyrite that form unequivocal feeders for associated stratiform tourmalinites. Stratiform (bedded) tourmalinites are known at the Neg, St. Joe, Can Am, and Estella localities, British Columbia, and at the Fork and related occurrences in the Perma-Plains area, Montana. Some of the tourmalinites in the Sullivan mine area also are stratiform (Shaw and Hodgson, 1986). The St. Joe and Fork tourmalinites are unusual in being associated with highly carbonaceous argillites.

Tourmalinites in the Belt and Purcell Supergroups vary greatly in texture from aphanitic and cherty to somewhat coarse and granoblastic. Most are dark gray to black in hand specimen, although pale to dark brown tourmalinites occur in a few areas (e.g., Sullivan mine and Fors prospect, British Columbia). Some of the fine-grained tourmalinites have been worked by Native Americans into high-quality projectile points. The cherty tourmalinites have a distinctive grain size of approximately 5-10 μm (Ethier and Campbell, 1977), apparently reflecting primary hydrothermal precipitation. Other tourmalinites in the region may be much coarser (up to 5 mm), mainly as a function of recrystallization associated with younger hydrothermal alteration and/or contact metamorphism.

The mineralogy of the tourmalinites is dominated by subequal amounts of quartz and tourmaline, together with accessory phases such as zircon, sphene, rutile, allanite, and ilmenite. Some Belt-Purcell tourmalinites also contain substantial pyrrhotite, garnet, feldspar, epidote, chlorite, biotite, and/or muscovite. The garnets are particularly distinctive, forming generally small (<3 mm) euhedral grains. In the Sullivan-North Star trend, such garnets locally make up 5-10 percent of the tourmalinites, and as much as 30 percent in tourmalinites from the immediate stratigraphic footwall of the Sullivan deposit (J.F. Slack and D.R. Shaw, unpub. data). Garnet-rich tourmalinites also occur at Mt. Mahon, British Columbia, and Morning Glory, Montana. Reconnaissance microprobe analyses by the author of similar garnets from Sullivan and from other Belt-Purcell tourmalinites (e.g., Morning Glory) document a major spessartine (Mn) component.

Models that have been proposed for the formation of tourmalinites include diagenesis of boron-rich colloids or gels, syngenetic precipitation from submarine-hydrothermal fluids, diagenetic modification of evaporitic borates, and syndepositional replacement of sediments and volcanics. A major constraint on any viable model is the limited solubility of aluminum in aqueous fluids at low to moderate temperatures and intermediate pH, conditions that are believed applicable to the formation of most, if not all, Belt-Purcell tourmalinites. Tourmaline typically contains 28-35 weight percent alumina (Al_2O_3), and it seems probable that the formation of most such tourmalinites was controlled by the availability of aluminum in detrital clays or feldspars, rather than in hydrothermal fluids. Evidence supporting this model comes from petrographic observations of tourmaline replacing the clay and/or feldspar matrix of quartz-rich sedimentary precursors of tourmalinites in the Aldridge and Prichard Formations (e.g., Ethier and Campbell, 1977; J.F. Slack, unpub. data). Recent whole-rock geochemical studies of tourmalinites in the Nagpur district, India, and the Broken Hill district, Australia, suggest protoliths of normal clastic sediments that were metasomatically altered at or below the sediment-water interface (Bandyopadhyay and others, 1993; Slack and others, 1993). A model preferred by the author involves migration of boron-rich hydrothermal fluids along permeable, sandy beds and the selective replacement of argillaceous beds (or laminae) in the upper parts of unconsolidated sedimentary sequences. This replacement may have taken place on the paleoseafloor, or tens to hundreds of meters in the subsurface. Depending on local structural and lithologic controls (i.e., cross-stratal permeability, favorable detrital mineralogy and mineral chemistry), the replacement may have formed discordant tourmalinites, stratiform tourmalinites, or both.

Previous studies of stratiform tourmalinites have favored a model involving syngenetic-exhalative processes coeval with sedimentation and/or volcanism (e.g., Slack and others, 1984; Plimer, 1988). However, I now believe that this model is applicable only to a limited number of tourmalinites, mainly those that are interlayered with, or in contact with, unequivocal chemical sediments. Candidates include stratiform sulfide deposits, iron formations, hydrothermal cherts, and fine-grained spessartine-quartz rocks or coticules (Spry, 1990). A broad association with stratiform sulfide deposits is considered insufficient evidence for an exhalative origin for tourmalinites, based on the clearly discordant nature of sulfide-related tourmalinites at the Sullivan mine (Hamilton and others, 1982) and the Black Prince mine in the Broken Hill district (Slack and others, 1993). The interlayered tourmalinites and coticules ("garnet quartzites") at Broken Hill are interpreted by Slack and others (1993) to be products of a submarine brine pool in which exhalative Fe, Mn, B, and perhaps Si, were added to clay-rich sediments on the sea floor, the clays serving as the critical source of aluminum for formation of the tourmalinites and coticules. These partly exhalative, coticule-related tourmalinites contrast with the other tourmalinites of the district that lack associated Mn-rich garnets and for which a subseafloor replacement origin seems likely. The preferential interlayering of ore-related tourmalinites with coticules at Broken Hill suggests that the presence of abundant Mn-rich garnet with tourmalinites is a potential prospecting guide in the Belt and Purcell Supergroups. The occurrence of garnet-rich tourmalinites preferentially in the immediate footwall of the Sullivan deposit supports this hypothesis.

Two other models for tourmalinite formation involve evaporitic borates and boron-rich colloids or gels. Borate minerals in marine and non-marine evaporite sequences are obviously attractive as a potential source of abundant boron. However, most borates are highly soluble and would readily dissolve during diagenesis, thus preventing their *in situ* preservation as tourmalinites (see Slack, 1982). Colloids or gels are more likely tourmalinite precursors because they have the ability to transport significant quantities of aluminum, as evidenced by the occurrence of aluminosilicate gels in ancient bedded manganese deposits in the Franciscan Complex, California (Huebner and Flohr, 1990) and in modern lacustrine sediments at Lake Magadi, Kenya (Eugster and Jones, 1968). On the basis of lithologic associations and trace-element geochemistry (e.g., low contents of rare earth elements), Slack and others (1993) suggested a colloid or gel-like protolith for a few tourmalinites (not associated with mineralization) in the Broken Hill district. I consider the colloidal origin proposed by Ethier and Campbell (1977) for the extremely fine-grained, felted tourmalinites in the footwall of the Sullivan deposit to be equally consistent with a simple replacement model involving hydrothermal dissolution of the precursor sedimentary minerals and precipitation of cryptocrystalline tourmaline by boron-rich hydrothermal fluids.

Preliminary field and petrographic studies of Belt-Purcell tourmalinites reveal features that may be useful in mineral exploration. Clearly the most obvious is the presence of sulfide minerals, especially sphalerite and galena, in both the discordant and stratiform types of tourmalinite. The occurrence of anomalous amounts (>10 vol %) of fine-grained, Mn-rich garnet within the tourmalinites (or in adjacent wall rocks) is also considered important as evidence of exhalative hydrothermal activity, because of the preferential precipitation of Mn in seawater at the sediment-water interface (rather than deep in the subsurface). Whole-rock analyses of tourmalinites may also prove valuable, based on the geochemistry of ore-related tourmalinites at Broken Hill that show elevated Fe (non-sulfide), Zn, Mn, and P, relative to barren tourmalinites in the district (Slack and others, 1993). The key is recognizing tourmalinites that formed from exhalative hydrothermal fluids at or near the sediment-water interface, as potential guides to nearby exhalative base-metal ores. In contrast are the vast majority of other tourmalinites in the Belt and Purcell Supergroups (both stratiform and discordant) that probably formed by subseafloor replacement of clastic sediments. Unlike the replacive tourmalinites, the partly exhalative tourmalinites are considered ~~true~~ time-stratigraphic markers and can be used as definitive indicators of exhalative hydrothermal activity. Such exhalative tourmalinites may be either proximal or distal to hydrothermal vents and sulfide ores (Slack, 1982; Plimer, 1988), depending on various factors such as local geochemical conditions and the extent of the submarine brine pool responsible for mineralization.

Another possible exploration guide is the presence of coexisting brown and black discordant tourmalinites. In the upper part of the footwall of the Sullivan deposit, both brown and black tourmalinites occur in complex parageneses (D.R. Shaw and J.F.

Slack, unpub. data). Although the compositions of the brown tourmalines in these tourmalinites are as yet unknown, they probably are magnesian, based on electron microprobe data on brown, Mg-rich tourmalines associated with other massive sulfide deposits (Taylor and Slack, 1984). The presence of both brown and black tourmalinites at Sullivan suggests mixing of Mg-rich seawater and Fe-rich hydrothermal fluids under high fluid/rock conditions (see Slack and Coad, 1989). Alternatively, these relations could reflect multiple periods of tourmalinization. The presence of coexisting brown and black tourmalinites elsewhere in the Belt and Purcell Supergroups therefore may be a favorable indicator for the occurrence of stratiform Pb-Zn deposits.

Many questions remain regarding the origin and exploration significance of tourmalinites. Chief among these are the nature of the fluids responsible for deposition of the tourmalinites, and whether their formation reflects low-temperature (<200°C) basin dewatering or higher temperature (200-300°C) submarine-hydrothermal processes. I believe that most of the tourmalinites that lack associated mineralization or chemical sediments (e.g., iron formations, coticules) probably formed mainly from metal-poor basinal brines, although more work needs to be done to test this hypothesis. Work is also needed on the likely role of growth faults in localizing the tourmalinites (both discordant and stratiform), the chemical reactions and hydrothermal alteration associated with tourmalinite formation, and the source of the contained boron. Geologic, geochemical, and isotopic studies of Belt-Purcell tourmalinites, currently underway, hopefully will answer these and other questions about their origin and exploration significance.

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CONTROLS OF MASSIVE SULFIDE MINERALIZATION IN THE BELT SUPERGROUP

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INTRODUCTION

Sediment hosted massive sulfide mineralization, in the Proterozoic Belt (Purcell) Supergroup of western Montana, northern Idaho, northeast Washington, and adjacent British Columbia and Alberta, was controlled by four geological factors: depositional environment, tensional tectonics, hydrothermal circulation, and metal introduction.

Only the informally named "lower Belt" (Prichard, Aldridge, and Newland Formations) hosts massive sulfide districts. These districts include Sullivan Mine, and other small mines in the Aldridge Formation of SE British Columbia and SW Alberta (Hamilton , and others, 1982; Höy, 1982; Turner, and others, this volume); the Cu-Co and Zn-Pb massive sulfide discoveries of the Sheep Creek area, central Montana, in the Newland Formation (Zieg and others, 1991,); and Zn-Pb massive sulfide occurrences of the Highland Mountains, southwestern Montana, in problematic "lower Belt" rocks of probable Prichard and Newland correlation.

DEPOSITIONAL ENVIRONMENT

The host sediments of the Belt basin massive sulfide districts are characterized by slow sedimentation, quiet basinal conditions, and reducing, anoxic conditions. The Aldridge Formation, host rocks of the Sullivan ore body are a rhythmic succession of thin- to medium-bedded, typically graded beds of very fine-grained quartz wacke interbedded with thinly laminated mudstone in beds from a few millimeters to several meters thick (Hamilton and others, 1982). Laminae and discontinuous blebs of pyrrhotite emphasize layering in the mudstone and contribute the distinctive rusty weathered surfaces of the Lower Aldridge. At the transition from Lower to Middle Aldridge Formation, the deposition of thinly laminated, sulfide-rich muds, silts, and very fine sands was interrupted by deposition of thicker, massive, lighter-colored, graded arenaceous beds interpreted to be turbidites.

The depositional environment of the Aldridge Formation is interpreted by Höy (1982) to be "deeper water" than the rest of the Purcell section. Similar rocks in the Prichard Formation of Montana contain abundant carbonaceous matter, in addition to iron sulfides, and are interpreted to have been deposited well below wave base, beneath water estimated to have been 400 m to perhaps more than 1300 m in depth (Cressman, 1989).

At Sheep Creek, Zieg (1981) describes the lower Newland Formation, which hosts massive sulfide bodies, as shales with planar very thin laminations and thin beds of microturbidite deposits with black carbonaceous laminae and very fine-grained pyrite laminae. The lower Newland deposition has been interpreted by Zieg (1986, Zieg and others, 1991) as the result of pelagic sedimentation in "subwavebase basinal environment" occasionally interrupted by slump deposits and turbidites.

In the Highland Mountains massive sulfide mineralization has been explored in sulfide rich, carbonaceous, calcareous argillite; massive, irregularly laminated, limestones; and debris flow breccia beds. This massive sulfide unit, as much as 200 feet thick, contains beds of massive (>50%) or semi-massive (30-50%) laminated pyrite or pyrrhotite, with sphalerite and minor galena (Thorson, 1984). Beds beneath the massive sulfide unit are thinly laminated, planar bedded, light- and dark-banded, pyritic or pyrrhotitic argillite which resembles the Prichard Formation. Overlying the massive sulfide unit are beds of interbedded, dark-gray to black, pyritic, carbonaceous, calcareous argillite and gray to black carbonaceous limestone which have been correlated with the Newland Formation.

TECTONIC ENVIRONMENT

Sediment-hosted massive sulfide mineralization is also favored by a tensional tectonic environment. On the largest scale the massive sulfide deposits occur during the early development of the Belt basin when the basin subsidence rate was greater than sedimentation. On a smaller scale the extensional environment is usually expressed by growth faults which control the basins in which massive sulfides were deposited.

In southeastern British Columbia and southwestern Alberta, Höy (1982) has documented several transverse faults which appear to have controlled deposition of thickened sections of Aldridge sediments, including bodies of intra-formational conglomerate. This suggests that syndepositional faulting was a local control of both of the restricted depositional conditions and the introduction of hydrothermal fluids into the basin.

Along the north edge of the Helena Embayment tongues of intraformational conglomerate and soft sediment slumps in the Newland Formation originated the upthrown, northern side of a major basin bounding fault (Zieg,1981, Zieg and others, 1991). These conglomerate tongues are interbedded with the massive sulfide lenses, indicating that the fault activity and massive sulfide deposition were contemporaneous.

Probably the best examples of multiple sets of contemporaneous faults which controlled the deposition of massive sulfides in the Belt basin are in the Highland Mountains. On a large scale, a growth fault along the south edge of the Helena embayment controlled "Lower Belt" sedimentation. On a smaller scale, the massive sulfide unit of the Highland Mountains was deposited in a series of sub-basins controlled by NW-trending normal faults contemporaneous with massive sulfide deposition.

In each of the massive sulfide districts some degree of control of massive sulfide deposition by extensional tectonics has been documented or interpreted. That control has two expressions; control of the depositional basin in which the massive sulfide accumulated, and the creation of crosscutting high permeability zones through which hydrothermal fluids were conducted to the sea floor.

HYDROTHERMAL CIRCULATION.

By far the best documented examples of the effects of hydrothermal circulation associated with a sedimentary massive sulfide have been described from the Sullivan Mine (Hamilton and others, 1982). Tremendous volumes of pre-ore breccia in a pipe beneath the ore lens have been nearly totally replaced by tourmaline. Also associated with the massive sulfide, and apparently associated with the hydrothermal system responsible for the mineralization, are zones of sericite alteration and albite-chlorite-pyrite-carbonate alteration (see Turner and others, and Leach and others, this volume). Estimations of the volumes of metals and alteration elements involved in the geochemical system include 50 million metric tons of boron in the tourmalinized zones, 60 million metric tons of sodium in the albite and tourmaline, 900 million metric tons of iron in massive iron sulfides, and 20 million metric tons of lead plus zinc (Hamilton and others, 1984). Saline fluids at temperatures greater than 250° C in the tourmalinized footwall zones, the ore, and the hanging wall sediments are indicated by ^{18}O data.

In the Sheep Creek deposits Zeig and others (1991) have described both "replacive" and exhalative processes involved in the formation of the Cu-Co mineralization of the district. Large volumes of carbonate and silica alteration are associated with the replacement processes, which are attributed to "diagenetic fluid flow and sulfide and sulfate (barite) replacive recrystallization within a pile of metal-bearing syngenetic iron sulfide." Distal to the Cu-Co mineralization in the disrupted, brecciated, replaced and altered, apparently proximal area, are extensive layers of bedded, fine grained, finely laminated iron sulfides with variable amounts of sphalerite and galena in the exhalative part of the system.

In the Butte South district a small area of Cu-Co mineralization in disrupted and brecciated carbonates may be a proximal zone similar to that in the Sheep Creek district. Distal to the Cu-Co zone there is an extensive unit of fine- to coarse- grained, finely laminated iron sulfides (both pyrite and pyrrhotite) with associated laminae of galena and sphalerite. The Zn-Pb-Fe mineralization is exposed along 10 miles strike in a series of half-graben sub-basins. The size of the system implies a large and potent hydrothermal system which carried metals into the sub-basins of deposition.

METAL INTRODUCTION

Each of the SMS districts of the Belt basin shows some aspect of zoning from a proximal area with crosscutting alteration or mineralization and brecciation or disrupted sediments to a more distal environment characterized by the intimate interbedding of sulfide layers with the host sediments on every scale from tenths of inches to hundreds of feet. The accumulated impression is that each district includes some part of a submarine hydrothermal exhalative system including both a feeder system with mineralization or alteration which replaced Belt sediments and exhalitive beds of sulfides deposited contemporaneously with sedimentation.

These SMS systems of the Belt basin are very large hydrothermal systems which affected tremendous volumes of sediments and large areas. SMS metal introduction in Sullivan area was part of a system which included the Stemwinder, Vulcan, Northstar, Kootenay King and Estelle mineralization over a distance of about 20 miles. The Sullivan massive sulfide lens was as much as 250 ft thick and is overlain by about 150 ft of hanging wall sediments with sporadic thin sulfide beds. The tourmaline altered feeder zone is known to extend for at least 1,500 ft below the massive sulfide lens (Hamilton and others, 1982)

In the Sheep Creek district Zeig and others (1991) reported three major sulfide layers, as much as 180 ft thick, in the lower Newland Formation, over a stratigraphic interval of at least 2,000 ft and have concluded that exhalative activity must have been nearly continuous throughout lower Newland deposition. Apparent centers of mineralization extend along the strike of the Volcano Valley fault for at least 15 miles.

Besides the interpretations that exhalative-appearing thinly laminated mineralization was deposited contemporaneously with sedimentation, and that replacement mineralization and alteration probably occurred under diagenetic conditions in relatively unconsolidated sediments, Pb isotope data also support a Proterozoic age for the mineralization.

CONCLUSION

The Proterozoic massive sulfide districts of the Belt basin share geological features which suggest origins through common sedimentary and exhalative processes. The widespread occurrence of deposits of this character in Middle Proterozoic sediments of the region also suggests basin-wide exhalative events during the early development of the Belt basin. This conclusion and the common occurrence of prospects displaying some of the SMS controls further suggests that additional sedimentary exhalative ore deposits remain to be discovered.

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GOLD RESOURCES IN THE GREYSON FORMATION, BIG BELT MOUNTAINS, MONTANA: PART II. MINERALIZATION AND GENESIS

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Stratiform Au occurrences in the Greyson Formation of the Proterozoic Belt Supergroup near York, Montana were recognized in the early 1980's, and have been explored by several mining companies. Our work, and this paper, is part of a research and exploration project for Independence Mining Company, current claim holders to most of the area. This paper is the second of a two-part presentation the gold resources of the Greyson Formation. For a summary of the geological setting please see the companion paper (Whipple and Morrison, this volume). The support and encouragement from IMC is greatly appreciated.

MINERALIZATION

Gold in the Greyson Formation near York occurs in resistant thin ledges, from a few feet to over 100 ft in thickness. The ledges, which have been called "reefs", persist along strike for thousands of feet. The reefs occur most frequently in the middle part of the informally designated middle member of the Greyson, unit 2 of Whipple and Morrison (this volume), and extend for at least 12 miles SE from the village of York. Multiple reefs have been explored at several locations; as many as 10 individual reefs are found in about 1600 feet of middle Greyson at York Gulch. The total gold resource in the York reefs, in rocks of greater than 0.01 oz/ton Au, has been estimated to be over 7 million ounces.

The reefs are composed of variable proportions of very fine-grained feldspar, quartz, ferroan carbonate, and illite. Bedding laminations and detailed sedimentary structures are preserved in thin to very thin beds which were originally mud, silty mud, and argillaceous silt. The rocks of the reefs appear cherty and contain minor amounts of clastic quartz and heavy mineral grains in the silty intervals, but most of the rock is fine-grained feldspar which has been interpreted to have formed by replacement of original argillic sediment (R. M. Honea, 1992, written commun.).

Wallrock outcrops around the reefs are poorly exposed, tan weathering, weakly iron-stained argillites and phyllites derived from thin-bedded, thinly laminated, clay-rich argillite, silty argillite, and occasionally, argillaceous siltstone. Color of unweathered wallrocks in drill core varies from dark green and dark greenish gray, through lighter shades of gray and greenish gray, to tan and cream colors. Beds are dominantly planar and parallel, although occasionally they are interlayered with soft sediment slumps of mud-dominated fragments in a mud matrix and with thin, graded, mud-chip debris beds interpreted to be turbidite deposits. Silty beds commonly have isolated, starved, cross-laminated silt ripples which traveled across cohesive mud beds. Clay in the reef and reef wallrocks has been determined to be illite of the 1Md polymorph. The environment of deposition is interpreted to be a quiet, restricted basin, generally starved of clastic sediment. The periodic interruption of this quiescence by mud-dominated slumps and turbidites suggests periodic, synsedimentary movement along intra-basin faults. Preliminary work with drill cores indicates that reef mineralization transgresses stratigraphy and that reefs are surrounded by a series of wallrock zones of differing character which also transgress stratigraphy and appear to be the result of alteration.

The feldspathic reef rocks have been overprinted by a complex series of veins and veinlets containing variable proportions of carbonate, k-spar, and quartz. The earliest, the "bed veins" are discontinuous veinlets, approximately parallel to bedding, commonly composed of ferroan dolomite with minor k-spar and quartz. Later en echelon zones of similar mineralogy cut the bed veins" and early bedding parallel quartz veins. Veining dominated by quartz started along bedding parallel zones before the en echelon veinlets but persisted longer to make late cross-cutting quartz veins which occasionally contain visible Au and base metal sulfides. The geometry and paragenesis of the veining indicates that the host reef beds were being inflated perpendicular to bedding in the early stages of veining and then extended parallel to bedding in the later part of the veining. Local areas of quartz veining in compressional intrafolial folds, although impressive, are much less common than the veins resulting from inflation and extension which are characteristic of the strataform mineralization over its entire strike length.

Diligent sampling of reef material, avoiding late cross-cutting quartz- carbonate veins, but including ubiquitous bed veins, gave an average Au content of the reefs at 1025 ppb Au in 32 samples (Table 1). Reef material from that same suite of 32 samples contained as much as 11.5% K_2O, and averaged 6.95% K_2O and 3.13% Na_2O. Twenty-seven argillite wallrock samples collected adjacent to reefs contained an average 3.36% K_2O and 0.93% Na_2O. X-ray diffraction patterns indicate that reef feldspar contains significant albite. In addition to the enrichment of reef material in Au, K_2O, and Na_2O, only Cr and Te have been found to be consistently anomalous and related to Au in the reefs. Chromium micas have been described from core drilled on the project and confirmed by X-ray techniques (T. H. Chadwick, 1988, written commun.). Recent microprobe studies on reef material (Bart Cannon, 1992, written commun.) have discovered inclusions of galena and Au tellurides within pyrite in both concordant and discordant veins. Elements commonly associated with most Au mineralizing systems, like As, Sb, and Hg, are not elevated in the reef mineralization in comparison to the wallrocks or unmineralized Greyson shales.

Gray- to dark greenish-gray-weathering, dark gray to black, carbonaceous and locally calcareous shale and silty shale from the type section of the Greyson Formation along Greyson Creek, 15 miles east of Townsend, Montana (Nelson, 1963) were sampled for comparison with the York reefs and wallrocks. Type-Greyson shales are similar to those at York, and the clay is also 1Md illite, indicating similar conditions of sedimentation and diagenesis. No Au mineralization, and no obvious alteration, is known from the type section, however. Table 1 shows data indicating that the York wallrocks are distinctly enriched in Au, Cr, Cu, Pb, and possibly Zn, in comparison to type-Greyson shales, although they have similar K_2O and Na_2O contents.

GENESIS

The sediment hosted stratiform Au deposits near York are unusual occurrences which have encouraged unusual proposals to explain their character and origin. Three major hypotheses have been advanced. The earliest hypothesis for the genesis of the reef mineralization depended on a syngenetic, sedimentary, perhaps exhalative, introduction of the Au and K^+ to explain the stratiform character of the reefs and the preservation of finely detailed sedimentary structures. This hypothesis was emphasized in the early exploration of the district and was revived by T. H. Chadwick (1988 written commun.).

The second hypothesis held that mineralization was controlled by Phanerozoic bedding plane thrust faults that resulted from the formation of the Cordilleran overthrust belt Zahony (1987, written commun.). Both Zahony and Chadwick consider that reef structures, such as bed veins, cross-cutting veins, and intrafolial folds, resulted from later events, either basinal uplift or formation of the overthrust belt.

We can not support these interpretations, because:

1. Reefs and surrounding wallrock alteration cross bedding gradually along strike, and more rapidly along the present dip. This distribution appears to reflect neither syngenetic control nor bedding plane thrust fault control;
2. Reef feldspathization and wallrock alteration appear to affect various lithologies, regardless of their depositional environments, casting doubt on a syngenetic origin; and
3. Reef veins and intrafolial folds are geometrically incompatible with an origin during overthrusting. The mineralogy of the veins indicates a complex paragenesis that involves K-spar, and suggests that development of these structures was integral to formation of the reefs.

The third hypothesis, a diagenetic origin for mineralization, was first postulated by Baitis, and others, (1988, 1989) based on detailed studies of textural relationships within the reefs. A supporting oxygen isotope study of the reef K-feldspar yielded a mean $\delta^{18}O_{SMOW}$ of about 10 ‰ that is remarkably uniform along a strike of about 7 miles. The $\delta^{18}O$ value of the K-feldspar is considered to be most consistent with an authigenic origin for the feldspar, and supports the interpretation of a diagenetic origin for the reefs (Susan Hall, oral commun., 1990). Based on the strong positive correlation of K_2O and Au, gold was also introduced during diagenesis. The interpretation of authigenic growth of K-spar in lower Belt sedimentary rocks at the Golden Sunlight Mine (Foster and Chadwick, this volume) may be additional support for a diagenetic origin for the reefs.

The available evidence indicates to us that the mineralization which characterizes the York reefs substantially post-dates sedimentation, but predates formation of the overthrust belt. Fine-grained sediments such as these are not normally expected to be significantly permeable, however. The reef veins, which show consistent evidence of origin by extension, could only have formed under conditions of significantly elevated pore pressure. Moreover, the abundance of

Table 1. Average content of selected chemical components in Greyson Formation at the type section and at York, Montana. (n = number of analyses)

	type-Greyson shales	n=	York wallrocks	n=	York reefs	n=
Au	7 ppb	(12)	32 ppb	(27)	1025 ppb	(32)
Te*			54 ppb	(7)	172 ppb	(8)
Cr	79 ppm	(12)	148 ppm	(32)	129 ppm	(32)
Cu	21 ppm	(12)	44 ppm	(11)	26 ppm	(23)
Pb	47 ppm	(12)	92 ppm	(11)	46 ppm	(23)
Zn	74 ppm	(12)	91 ppm	(11)	86 ppm	(23)
K_2O	3.17%	(12)	3.36%	(32)	6.95%	(32)
Na_2O	1.16%	(12)	0.93%	(7)	3.13%	(9)

*Te data from Zahony (written commun., 1987, p. 35) discontinuous, thin veins suggests that such pore pressure must have originated within the immediate host strata. But, inasmuch as the reefs now contain a fundamentally dehydrated mineralogy, these pore pressures must have originated from mineralogic changes that took place earlier in the formation of the reefs.

Another observation of particular significance is that the 1Md illite, which most characterizes the wallrocks, and which persists in small amounts within reefs, has been interpreted to originate from smectite. In contemporary sedimentary basins, such as the Gulf of Mexico, smectite converts to 1Md illite after burial to depths of 10,000 to 15,000 feet (e. g., Bruce, 1984). This mineralogic transformation is thought to depend on moderately elevated temperature, in the range of 100 to 150 °C, and on the availability of K^+. Both of these become available with burial, the K^+ being supplied by the contemporaneous dissolution of detrital feldspar. This transformation takes place over a limited vertical range of burial, liberating large quantities of water, silica, and Mg^{++}. The liberated water elevates pore pressures and increases porosity and permeability. In organic-rich sediments, the water and the permeability that results from smectite diagenesis is implicated in liberation and transport of hydrocarbons from otherwise impermeable source beds.

We propose that formation of York reefs was fundamentally dependent on burial and clay diagenesis, in an essentially two-phase process. First, transformation of smectite to illite created porosity and permeability which made it possible for fluids to penetrate and traverse these fine-grained sediments. Second, continued diagenesis, taking place laterally or deeper in the basin, provided fluids rich in K^+ which transformed illite to K-spar. Flow of these basinal fluids may have been guided by permeable sands, listric bedding-plane faults, or cross-cutting extensional faults. Progressive alteration of illitic sediment maintained permeability, through dehydration of illite to K-spar, until the illite was effectively exhausted. As the reefs were converted to brittle K-spar and rendered less permeable, increasing pore pressures formed expansion fractures, which channelized subsequent fluid flow. We draw a parallel between the York mineralization and the feldspathization, dolomitization, and regional scale fluid flow that characterizes Mississippi Valley Pb-Zn deposits. Recent papers on these deposits (e. g., Bethke and Marshak, 1990) have emphasized the coincidence of K-alteration, mineralization, and regional fluid flow. We see the primary distinction between the two deposit types as involving the degree of potassic alteration, and the kind of metals. In Belt rocks, we see a similar, although less specific, parallel to the formation of the Revett stratabound Cu-Ag mineralization. We presume the differences in all of these deposit types reflect differences in source rocks and concentrating mechanisms.

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THE SULLIVAN-NORTH STAR CORRIDOR: REGIONAL-SCALE ALTERATION, FRAGMENTAL ROCKS AND MINERALIZATION ASSOCIATED WITH THE SULLIVAN STRATIFORM LEAD-ZINC DEPOSIT, SOUTHEASTERN BRITISH COLUMBIA

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Cominco staff (e.g. Hagen, 1985) and Höy (1984) have shown that the Sullivan deposit lies along a district-scale zone of altered and fragmental rocks, the Sullivan-North Star corridor. Study of the corridor provides evidence for the structural control and associated larger-scale hydrothermal activity during the formation of the Sullivan deposit. The corridor also provides an opportunity to study the relationship between the differing alteration types and ore formation, which is critical to assessing their significance as mineral exploration guides.

The Sullivan deposit is in the Aldridge Formation of Middle Proterozoic age near the town of Kimberley in southeastern British Columbia. Aldridge strata near the Sullivan mine form a gently-folded, shallowly northeast-dipping homocline cut by small displacement north-trending faults and the large-displacement west-trending Kimberley fault (Fig.1). The Sullivan deposit lies near the transition between siltstone and argillite of the lower Aldridge Formation and overlying sandy turbidites and argillite of the middle Aldridge Formation (Fig. 2).

The Sullivan-North Star corridor is an elongate, north-south-trending zone of fragmental and disrupted sedimentary rocks that are variably altered and mineralized. Altered and fragmental rocks of the corridor extend 6 km in length, 1-3 km wide, and over a stratigraphic thickness of at least 500 m (Figs. 1 and 2). The Kimberley fault truncates the corridor to the north; at the southern end the corridor disappears under the Quaternary deposits of the St. Mary Valley. The corridor is widest adjacent to the Kimberley fault.

Three gabbro sills with an aggregate thickness of 650 m are known below the Sullivan deposit (Fig. 2). These sills are the upper part of a sill complex in the lower Aldridge Formation emplaced during deposition of Aldridge strata. The uppermost gabbro is called the Footwall sill and forms a discordant arch-like feature roughly coincident with the corridor. The top of the arch coincides with the stratigraphic level of the Sullivan orebody.

A variety of fragmental rocks are present throughout the corridor (Hagen, written commun. 1985). Bedded conglomerate beds are immediately below the stratigraphic level of the Sullivan deposit, extend across the width of the corridor, and have graded tops. The bedded conglomerate unit is typically 30-50 m thick, but exceeds 300 m thick west of the Sullivan deposit (Fig. 2). Bedded conglomerate is commonly underlain by disrupted strata and discordant zones of fragmental rock. Discordant fragmental zones, centimeters to meters in width, include angular breccia, conglomerate and pebbly sandstone. Major discordant fragmental zones are commonly associated with sulfide veins. Pebbly sandstone is composed of lithic fragments up to 10 mm in diameter scattered in a sandy matrix. Pebbly sandstone is present as massive, unbedded units that are conformable or semi-conformable, are tens to hundreds of meters thick, and transitional laterally into disrupted strata.

The Aldridge Formation is typically composed of 15-20% detrital feldspar. Altered rocks within the corridor are distinctive by a near absence of detrital feldspar and the collective abundance of muscovite, chlorite, tourmaline, sulfide minerals, garnet and epidote (Leitch and others, 1991). Major alteration types include: (1) tourmalinite; (2) muscovite; (3) albite-biotite-chlorite; (4) biotite-plagioclase hornfels; and (5) quartz-plagioclase-biotite granophyre (Fig. 3). Intensity of alteration is highly variable within the corridor and related to a number of discrete hydrothermal centers, the major ones being the Sullivan deposit and the Stemwinder vein system.

Tourmalinite alteration occurs as (1) a pipe underlying the western part of the Sullivan deposit (Fig. 3), (2) an envelope on the Stemwinder vein, (3) a cluster of small bodies extending south from the Stemwinder vein to the North Star deposit, and (4) a small body at the south end of the corridor. At Sullivan, Stemwinder, and North Star deposits, the tourmalinite is sulfide-rich; the small tourmalinite at the south end of the corridor is sulfide deficient. Tourmalinite extends to the stratigraphic level of the highest sulfide bands at Sullivan; the original extent of hangingwall tourmalinite has been obscured by later albitic alteration. Tourmalinite clasts within unaltered conglomerate are rare and suggest only limited tourmalinite formation prior to conglomerate deposition.

Muscovite alteration caused destruction of detrital feldspar and its presence is evident by the lack of biotite formation during regional metamorphism. Muscovite-altered rock is soft and grey to greenish-grey in colour (Leitch and others, 1991) and is widespread throughout much of the Sullivan-North Star corridor as: (1) large zones often associated with disseminated and veinlet pyrrhotite and roughly coincident with thick fragmental units; (2) as envelopes on pyrrhotite +/- sphalerite-galena veinlets that cut tourmalinite below the Sullivan deposit; (3) replacement of tourmalinite (e.g.Sullivan, elsewhere); and (4) as a weak fringe around tourmalinite bodies at the Sullivan and Stemwinder deposits.

Disseminated porphyroblasts of manganiferous garnet are present throughout the corridor but absent in strata to the west and east (Leitch and others, 1991). Pale pink garnet-rich beds are associated with bedded sulfide minerals at the Sullivan and North Star deposits.

Three large albitite bodies within the corridor are: (1) overlying the western Sullivan orebody (Shaw and Hodgson, 1980; Hamilton and others, 1982) (2) a kilometer west of the Stemwinder vein immediately overlying the footwall gabbro (Turner and Leitch, 1992); and (3) 3 km west of the Sullivan deposit within the conglomerate horizon (Fig. 3). Smaller zones of albitic altered sedimentary rocks occur adjacent to gabbro below the Sullivan deposit.

ORIGIN OF THE SULLIVAN-NORTH STAR CORRIDOR

Bedded conglomerate appears to have been erupted to the seafloor via feeder zones represented by conglomerate dikes and discordant breccias bodies (Hagen, written commun. 1985). Graded bedding suggests transport on the seafloor as mass flows. The well sorted and rounded nature of the conglomerate precludes simple collapse of strata on a fault scarp; instead, conglomerate dykes indicate rounding and winnowing occurred below the seafloor during upward movement. The lack of bedding in pebbly sandstones suggests subsurface disaggregation of silt and sand-rich sediments due to liquifaction rather than seafloor transport and deposition.

The energy required to erupt sediment may have been provided by gabbro intrusions at depth, fault movement, or rapid release of overpressured fluids within permeable horizons. The coincidence of some pebble dykes and large sulfide veins in the corridor suggests common structural controls for sediment eruption and hydrothermal upflow. Conglomerate eruption was closely followed by mainstage hydrothermal activity at the Sullivan deposit.

Tourmalinite appears to be the alteration type most closely linked to ore formation. Tourmalinite is spatially associated with the two major sulfide systems at Sullivan and Stemwinder, and based on stratigraphic evidence at Sullivan, is also temporally associated with ore formation (Leitch and Turner, 1992). The widespread distribution of muscovite alteration around but not directly associated with the major sulfide bodies suggests it reflects lower temperature fluid-rock interaction away from focused discharge areas. At the Sullivan deposit, muscovite alteration postdates tourmalinite formation and is associated with the waning stages of hydrothermal activity (Leitch and Turner, 1992). Garnet-rich beds associated with bedded sulfide minerals appears to reflect Mn-rich sediments exhaled on the seafloor. Disseminated garnet reflects Mn enrichment of wallrock during alteration in the corridor (Leitch and others, 1991). The close association of some albite-chlorite and albite-biotite-chlorite alteration with gabbro dike and sill contacts below the Sullivan deposit and elsewhere in the corridor indicate these types of alteration were synchronous with or postdated gabbro emplacement and are later than ore formation (Turner and Leitch, 1992).

The north-south alignment of fragmental and altered rocks, sulfide veins and discordant part of the Footwall sill indicate a northerly-trending fault zone was central to the corridor. Widening of the corridor along the Kimberley fault may suggest an ancestral fault active during formation of the Sullivan deposit; the location of the Sullivan deposit may reflect a high permeability zone along the intersection of these two fault zones. Early fault movement resulted in fragmentation and eruption of sediments, followed by widespread hydrothermal activity within the corridor. Focused discharge resulted in formation of a large seafloor sulfide body (Sullivan), as well as a smaller sulfide vein (Stemwinder deposit) and overlying seafloor deposit, now largely eroded (North Star deposit). Emplacement of gabbro intrusions was locally controlled by faults within the corridor; hydrothermal alteration associated with gabbro emplacement was focused along discordant limbs of intrusions.

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LITHOFACIES OF THE HELENA AND WALLACE FORMATIONS (BELT SUPERGROUP, MIDDLE PROTEROZIC), MONTANA AND IDAHO

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The Helena and Wallace Formations (Belt Supergroup, Middle Proterozoic), which are informally called the "middle Belt carbonate," are exposed widely in Belt terrane of the western United States and Canada. In western Montana and northern Idaho these formations are subdivided into seven lithofacies. Although the Helena Formation contains the most carbonate-rich rocks of the Belt Supergroup, the lower lithofacies is clastic-rich compared to most of the formation. In contrast to the Helena Formation, the Wallace Formation is composed mostly of clastic rocks, but the lower lithofacies is carbonate-rich compared to most of the formation. Original lithofacies relations have been modified by Cretaceous thrust faults and by strike-slip faults. Accurate restoration of lithofacies to initial sites of deposition is not yet possible, and the original size and shape of the basin are unknown. In general, the Helena Formation is mostly in the eastern and northeastern parts of Belt terrane, the Wallace Formation is in the western and southwestern parts, and the two formations intertongue in the central part (Fig. 1).

HELENA FORMATION

The lower member of the Helena Formation is a clastic and limestone lithofacies exposed mainly south of Missoula and west of Kalispell, Montana, (Fig. 1) and is composed of five principal rock types (Fig. 2): (1) laminated or mottled silty dolomite, (2) gray argillaceous and silty limestone, (3) brown siltite and fine-grained quartzite, (4) brown argillite and siltite, and (5) green argillite and siltite. Zones of each rock type range between 0.5 m and 3 m thick, but most commonly are about 1 m thick, and interbedding seems random. These rocks are partly equivalent to the cyclic lithofacies east of Kalispell and Butte, Montana, and lateral and vertical contacts appear gradational between these two lithofacies. Rocks of the clastic and limestone lithofacies of the Helena Formation were deposited on a subtidal, shallow slope on the west and southwest flank of a carbonate bank preserved along the eastern limit of the middle Belt carbonate.

The upper member of the Helena Formation east of Kalispell and Butte, Montana, is a cyclic clastic and carbonate lithofacies, but in the south part of Belt terrane west of Butte, the upper part of the Helena Formation is an argillaceous limestone lithofacies. The cyclic clastic and carbonate lithofacies is composed of cyclic sequences that contain different kinds of clastic and carbonate rocks (Fig. 2), and each cycle has an erosional base. Eby (1977) originally described cycles from the

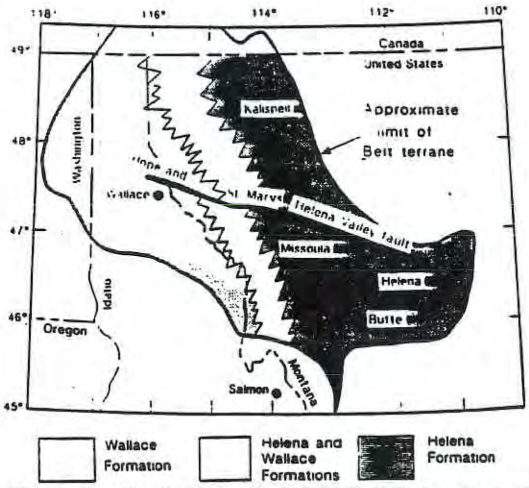

Figure 1. Distribution of middle Belt carbonate rocks, Belt terrane, USA.

region east of Kalispell and north of Helena, Montana, that were composed of six lithologies, and Eby's cycles have been generalized into: (1) gray calcareous argillite and siltite at the base, (2) silty dolomite bed above, and (3) stromatolitic and cryptalgal dolomite at the top. Cycles described by Harrison from the Kalispell region have (1) white quartzite at the base, (2) gray dolomitic siltite in the middle, and (3) gray dolomite at the top. Don Winston (oral commun., 1992) described cycles from east of the Kalispell area that have (1) gray clastic beds or intraclastic limestone at the base, (2) gray silty limestone in the middle, and (3) gray dolomite at the top. Generally a complete cycle is about 1-9 m thick, but individual cycles may be as thick as 30 m. Incomplete cycles are common, and in them uppermost or lowermost beds are missing (Eby, 1977; Schmidt and others, 1983). In the region south of Missoula and west of Butte, Montana, the Helena Formation is composed of silty and argillaceous gray limestone and calcareous siltite and argillite. North of the Helena region, this argillaceous limestone lithofacies is a tongue within the cyclic lithofacies described by Eby (1977), and this lithofacies thickens westward toward the Missoula region, where it forms most of the Helena Formation.

The cyclic lithofacies and the argillaceous limestone lithofacies of the upper member of the Helena Formation intertongue laterally toward the west, and in the southcentral part of Belt terrane near Missoula, Montana, the argillaceous limestone lithofacies overlies the cyclic lithofacies. Both lithofacies of the upper member of the Helena Formation intertongue with clastic rocks of the middle member of the Wallace Formation in the Missoula region.

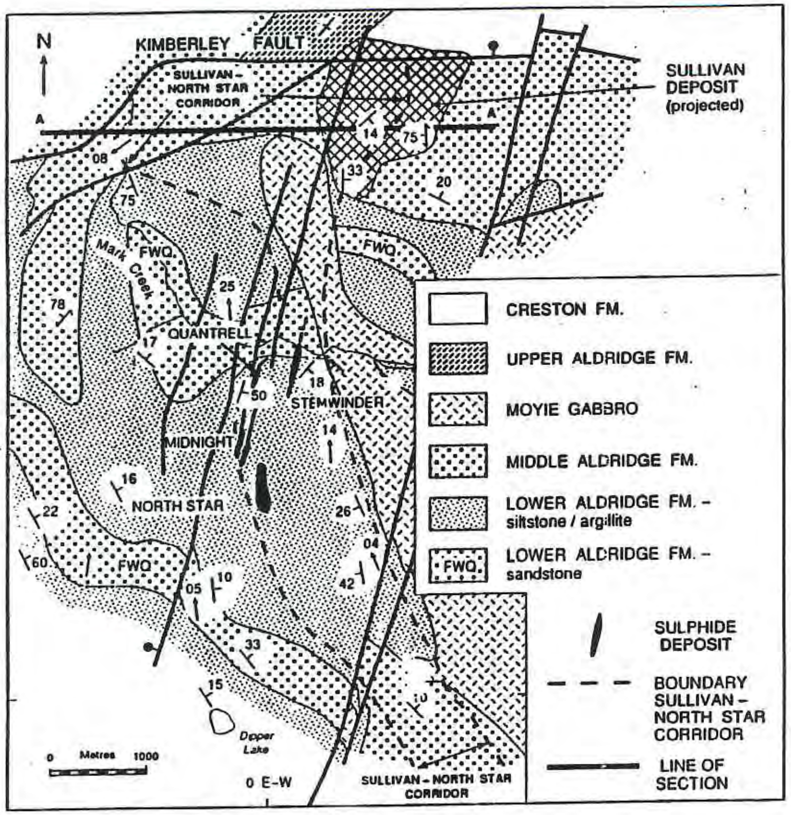

Figure 1: Geological map of the Sullivan mine area.

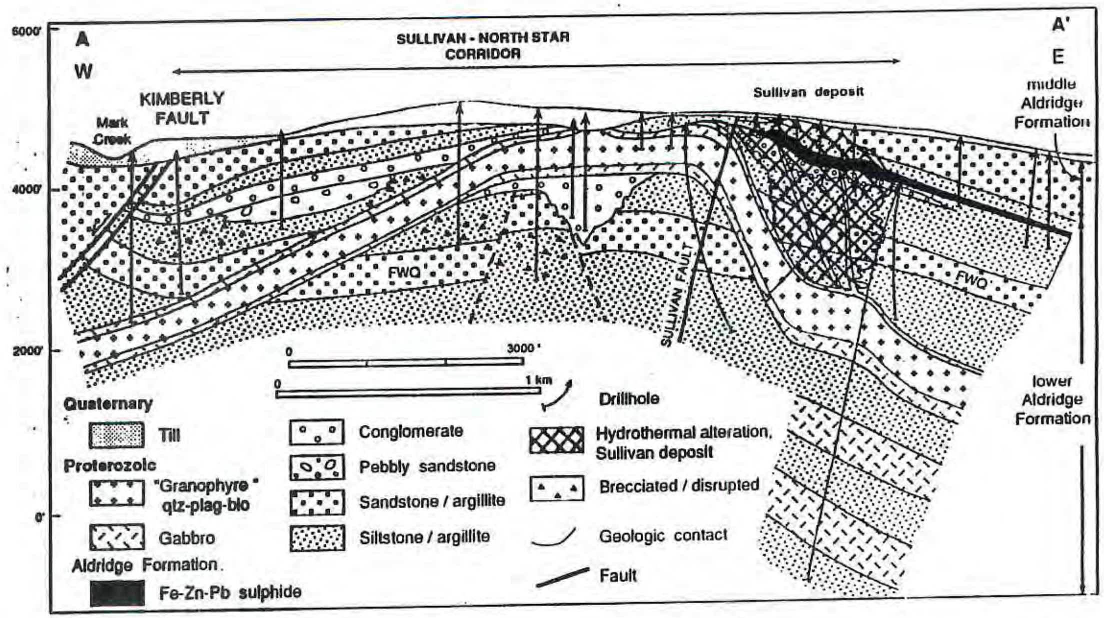

Figure 2: East-west geological cross-section of the northern Sullivan-North Star corridor. The location of the section is shown in Figure 1. The Footwall Quartzite (FWQ), a thickbedded arenite unit, is a good stratigraphic marker within the lower Aldridge Formation.

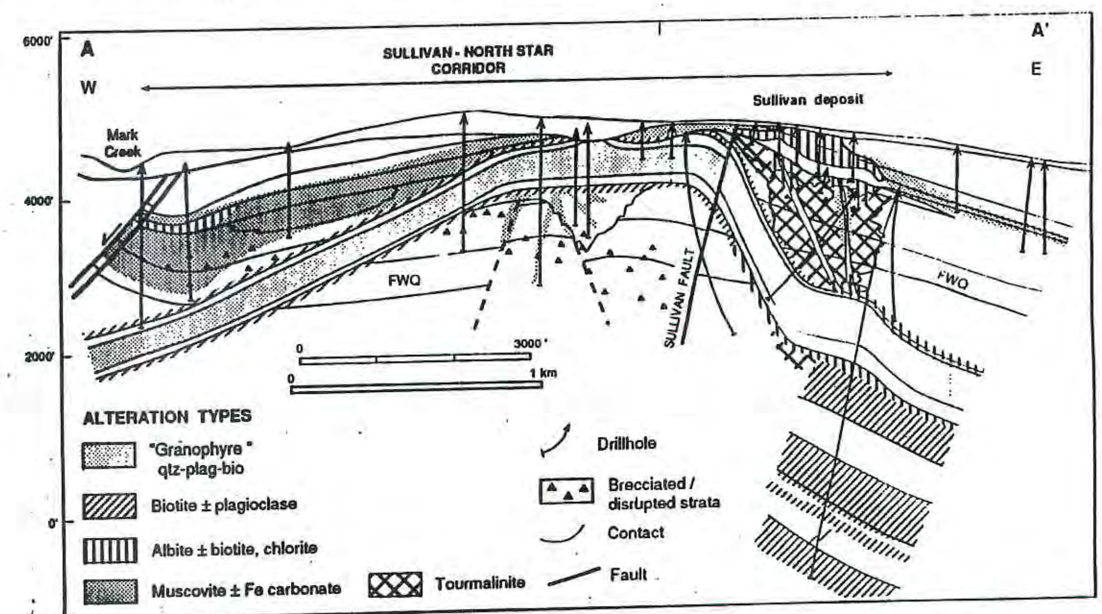

Figure 3: East-west cross-section of the northern Sullivan-North Star corridor **indicating the distribution of alteration types.** See Figure 1 for location. **Compare with Figure 2.**

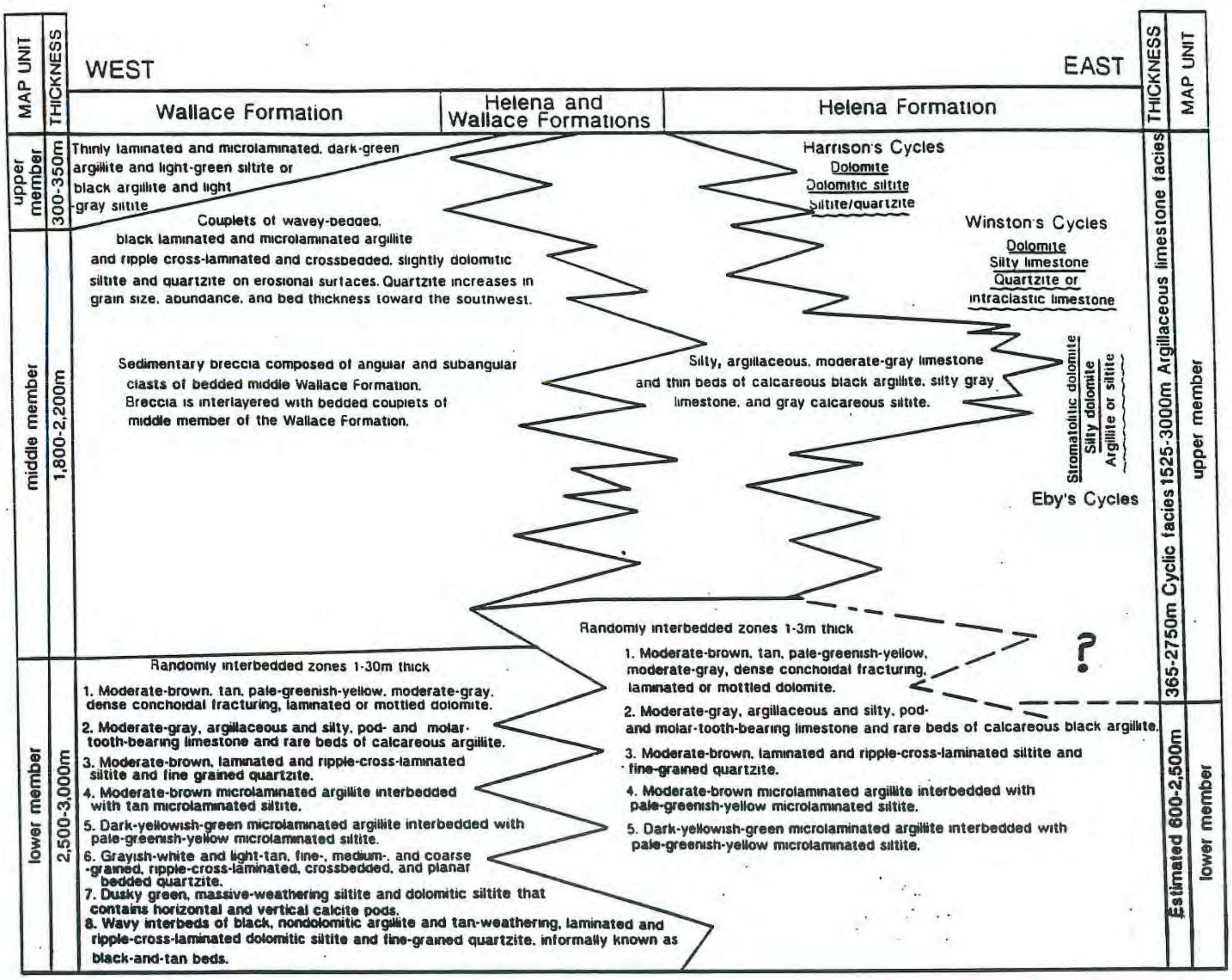

Figure 2. Diagrammatic distribution of lithofacies of the Helena and Wallace Formations (Belt Supergroup) in Belt terrane.

Rocks of the cyclic lithofacies were deposited in a peritidal carbonate bank, and stromatolites developed in the upper part of cycles along the eastern limit of Belt terrane. In the region between Missoula and Butte, Montana, the argillaceous limestone lithofacies was deposited in slightly deeper water on the west flank of the carbonate bank, and these rocks represent a more seaward facies of the laterally equivalent cyclic lithofacies.

WALLACE FORMATION

The lower member of the Wallace Formation is a dolomite and clastic lithofacies that is widely exposed in the western part of Belt terrane. This lithofacies is composed of eight principal rock types that are interbedded in a random manner (Fig. 2): (1) laminated or mottled silty dolomite, (2) argillaceous and silty gray limestone, (3) brown siltite and fine-grained quartzite, (4) brown- and tan-weathering interbedded argillite and siltite, (5) green argillite and interbedded siltite, (6) grayish-white quartzite, (7) green siltite, and (8) interbedded tan siltite and black argillite (the "black-and-tan" beds). These principal rock types form zones 1-100 m thick, although zones 2-5 m thick are most common. Quartzite beds thicken toward the southwest where they form sequences several hundred meters thick. The grayish-white quartzite, green siltite, and the black-and-tan beds distinguish the lower lithofacies of the Wallace Formation from the laterally equivalent lower lithofacies of the Helena Formation.

The depositional environment of the dolomite and clastic lithofacies of the Wallace Formation was a peritidal slope or subtidal basin in which sand wedges tapered northeastward toward the central part of the Belt basin.

The middle member of the Wallace Formation is characterized by a clastic lithofacies of interbedded black argillite and tan-weathering dolomitic siltite, the black-and-tan beds, and by a breccia lithofacies composed of subaqueous synsedimentary breccia (Harrison and others, 1986). "Pinch-and-swell" bedding structure is a characteristic feature of the black argillite and tan-weathering siltite. Black argillite beds are microlaminated and nondolomitic, and range in thickness from a few millimeters to 10 cm. Dolomitic siltite beds contain planar lamination and ripple cross lamination, are normally graded, and generally have channeled basal contacts. Synsedimentary breccia occurs in the middle member of the Wallace Formation west and southwest of Missoula, Montana, and the matrix-supported breccia contains angular clasts of the black-and-tan beds, tan-weathering dolomitic quartzite and siltite, and tan-weathering silty dolomite in a finely comminuted matrix of dolomite, siltite, and black argillite. Stick-like psuedomorphs in the breccia may have replaced shortite, which is an authigenic sodium and calcium carbonate mineral that is deposited from hypersaline water.

The black argillite and dolomitic siltite couplets represent single depositional events on a shallow shelf; traction currents transported silt and eroded the top of the underlying black argillite beds. Organic material and clay settled from calm water over the silt layer. Subaqueous mass-flow deposits slid east and northeastward off the northeastern flank of the shelf to form synsedimentary breccia deposits that are interbedded with the black-and-tan beds of the shelf.

The upper member of the Wallace Formation is a siltite and argillite lithofacies that consists of microlaminated, wavey-bedded, interbedded dark-green argillite and light-green siltite or interbedded black argillite and light-gray siltite. Dolomitic siltite beds have small-scale erosional contacts at the base of beds, and dessication cracks are common (Harrison and others, 1986).

Fine-grained clastic rocks of the upper member of the Wallace accumulated in a shallow-water basin that was periodically exposed to the atmosphere.

HELENA AND WALLACE FORMATIONS

West of Kalispell and Missoula, Montana, rocks of the cyclic beds of the upper member of the Helena Formation are interbedded with the black-and-tan beds of the middle member of the Wallace Formation, and at these places the units are combined. Interbedding of these two different rock types is on a scale of a few meters to a hundred meters.

RELATIVE SUBSIDENCE RATES

Analysis of relative subsidence rates in Belt terrane during deposition of the middle Belt carbonate is based on lithofacies thickness and intertonguing relations among lithofacies that establish time-equivalence. The thickness of lithofacies represents a relative sediment-accumulation rate if lithofacies are time equivalent. The base of the overlying Snowslip Formation (Missoula Group, Belt Supergroup) is the datum for comparison of relative subsidence rates, and lithofacies data show that north of the Hope and St. Marys-Helena Valley faults (Fig. 1) the subsidence rate during deposition of the middle Belt carbonate was greatest in the east and west parts of Belt terrane. The subsidence rate increased from north to south toward the Hope and St. Marys-Helena Valley faults. South of the Hope and St. Marys-Helena Valley faults, the subsidence rate far exceeded that of the northern part of the basin. This interpretation suggests that faults of the Lewis and Clark line, of which the St. Marys-Helena Valley fault forms the north boundary, were active during deposition of the Helena and Wallace Formations.

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STRATABOUND GOLD IN THE GREYSON FORMATION, BIG BELT MOUNTAINS, MONTANA: PART I. GEOLOGIC SETTING

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In the Big Belt Mountains of central Montana, gold is widely distributed in steeply-dipping, fine-grained sedimentary rocks of the Middle Proterozoic Greyson Formation of the Belt Supergroup. These gold concentrations are referred to as the York deposits because of their proximity to the community and mining district of York about 20 mi northeast of Helena, Montana. The sediment-hosted zone of deposits, which seems to be stratabound, has an explored strike length of about 8.5 mi. Numerous old mines and prospects occur along the mineralized zone, but although exploration companies have mapped, sampled, and drilled along its outcrop, no mineable orebody has been defined.

The gold-bearing zone consists of interbedded siltite beds of two types: orangish-weathering mineralized beds that contain abundant, intergranular potassium feldspar and locally iron carbonate and non-mineralized beds of olive, thinly laminated siltite. Gold is disseminated in potassium-feldspar-bearing siltite and is also in veinlets parallel and discordant to bedding. Preliminary sample results show gold values ranging from 0.15 ppm in disseminated occurrences to 41.0 ppm in veinlets.

The mineralized strata form resistant but laterally discontinuous, orangish-weathering outcrops that exploration companies have termed "reefs"--not to be confused with carbonate reefs. The potassium feldspar is orthoclase (Batis and others, 1989) and appears for the most part to be authigenic; K_2O in mineralized rock is as much as 12%. Locally, at the base of the mineralized zone is a distinct interval, a few tens of feet thick, of altered and bleached siltite that contains 2.4 - 4.7% K_2O but no gold. We term the interbedded succession of "reefs" and olive siltite and the underlying interval of altered siltite, the "K-spar zone".

We have subdivided the Greyson Formation into four mappable, informal units here termed successively upsection 1 through 4; stratigraphic thicknesses are estimated from map measure. Unit 1 ranges in thickness from 500 to 1,700 ft and consists mostly of very thinly laminated siltite and argillite, locally ripple cross-laminated and cemented mostly by carbonate. Thin arenite beds that are common in the lowermost part become thicker, more abundant, and coarser grained to the southeast. Unit 2 ranges in thickness from 3,500 to 4,000 ft and is mostly very-thinly laminated, olive siltite, containing discontinuous interlaminae of blackish-green siltite and argillite. This vari-colored arrangement imparts a streaked appearance to exposures. Most of the mineralized K-spar-gold-bearing "reefs" are within unit 2. Overlying unit 2 and the K-spar zone are thick, blocky beds of gray to olive siltite and interbedded, very-thinly laminated siltite of unit 3. Unit 3 is approximately 1,200 ft thick. The blocky siltite beds are as thick as 1 ft, and some are clearly, hummocky cross-stratified. At the base of this unit, siltite beds locally contain slump structures and other evidence of soft-sediment deformation. The uppermost unit, unit 4, consists of mixed siliciclastic and carbonate lithologies, all indicating deposition in a shallow-water environment. It is about 1,000 ft thick. Prominent lenticular beds of gray limestone are locally fragmental and stromatolitic. Beds of white to light-gray pyritic quartz arenite as much as 1 m thick are common in the upper part of the unit. Interbeds of siltite and argillite couplets and thinly laminated argillite are commonly grayish-green, calcareous and contain mud-chip breccia. Salt casts are rarely present in the uppermost beds below the contact with red beds of the Spokane Formation.

The red bed sequence of the Spokane Formation is in fault contact with the Greyson along much of the outcrop limit of the mineralization. We believe this structural relationship has resulted in a misunderstanding of the regional stratigraphy, the stratigraphic position of the K-spar zone, and the thickness of the Greyson Formation. Detailed mapping using stratigraphic control points and the Golden Messenger dike indicates that the K-spar zone thins along strike both northwest and southeast from the Cave Gulch drainage where it has a maximum apparent thickness of 2,300 ft. The mapped boundaries of the K-spar zone cut generally downsection to the southeast. We have no data at this time to indicate the down-dip extent of the zone.

The succession of facies in units 1 through 4 of the Greyson Formation indicates a depositional environment that shallows upward from basin plain to subtidal. Our preliminary interpretation of lamination styles and sedimentary structures in unit 2, including the K-spar zone, proposes deposition of some sequences by turbidity currents in deeper parts of a restricted basin. The gold appears to be in turbidite-hosted deposits (Keppie and others, 1986), a type that ranges from disseminations to vein deposits and from Archean to Tertiary.

Deformed beds at the base of unit 3, overlying the K-spar zone, appear to be the result of local slumping and soft-sediment deformation that we interpret to be related to active basin-growth faults. Similar structures, thought to be Proterozoic in age, are inferred at the nearby Sheep Creek copper-cobalt deposit (Zieg and others, 1991; Zieg, this volume) and the Golden Sunlight Mine (Foster and Chadwick, 1990 and this volume), which are hosted by Belt Supergroup strata. Regionally, the mineralized zone and proposed growth faults lie within the Great Falls Tectonic Zone (O'Neill and Lopez, 1985).

We have tentatively identified several volcaniclastic interbeds a few feet thick in the Avalanche Gulch area that are stratigraphically above and below the K-spar zone. In the field, these beds are dark gray, very fine-grained tuffaceous rocks that locally contain small subrounded fragments a few millimeters across. Volcaniclastic rocks interbedded with thinly laminated siltite that was deposited in a deep-water environment suggests to us episodic submarine volcanism and subsequent mixing with sediment during transport away from eruptive centers.

From our preliminary field studies, we infer that the age of the primary, disseminated gold deposits is probably Proterozoic and related to exhalative activity and hydrothermal fluids associated with basin growth faults.

Several sills and dikes intrude the Greyson Formation in the Big Belt Mountains. Most of these intrusive units appear to be Proterozoic in age and range in composition from gabbro to andesite. The Golden Messenger mine, near York, produced at least 11,000 oz. of gold from the late 1800's to about 1928 (McClernan, 1983) from quartz-ankerite veins in the dike and hornfelsed wall rocks. The mine is at the northern end of the K-spar zone, where the zone is intersected by a thick gabbroic dike of probable late Proterozoic age, called the Golden Messenger dike. The coincidence of the mine at this intersection suggests to us that gold could have been concentrated by thermal remobilization of low-grade disseminated gold in the sedimentary wall rocks. The enrichment process and geology are similar to that suggested for the Golden Sunlight Mine near Whitehall, Montana (Foster, 1991).

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THE GEOLOGY OF THE SHEEP CREEK COPPER DEPOSITS, MEAGHER COUNTY, MONTANA

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Copper deposits at Sheep Creek in central Montana occur within synsedimentary pyrite lithofacies in the northernmost exposures of the middle Proterozoic Newland Formation in the north part of the Helena embayment of the Belt basin (Zieg and others, 1991). Cominco American Inc. began exploration work in the area in 1976, presently controls 100% interest in the Sheep Creek project, and has outlined a possible resource of 4.5 million tonnes that contain 2.5% Cu and 0.1% Co in an upper sulfide zone, and 4 million tonnes that contain 4% Cu in a lower sulfide zone.

Belt stratigraphy in the northern Helena embayment north of the Volcano Valley fault (Fig. 1) consists of 1) the 800-foot-thick Neihart Quartzite, the basal Belt unit which unconformably overlies early Proterozoic crystalline basement; 2) the 2000-foot-thick Chamberlain Shale, a silty, carbonaceous black shale deposited in a current dominated environment; and 3) a 550-foot-thick section of interbedded black carbonaceous shale and "molar tooth" limestone and dolomite. This last unit, which was called Newland by Walcott (1899) and Keefer (1972), is stratigraphically below typical Newland and above typical Chamberlain (Zieg, 1981). South of the Volcano Valley fault, the 3000-foot-thick lower Newland Formation rests on Chamberlain Shale and consists mainly of thinly laminated to thinly bedded, noncalcareous to calcareous or dolomitic, very-fine-grained turbidite deposits which record subwave base depositional conditions. Mass flow deposits within the turbidite sequence include stromatolitic limestone olistoliths, dolomitic to calcareous quartz sand turbidite beds, and black silty shale-clast conglomerate; these deposits indicate an abrupt change to shallow water conditions a short distance north across an ancestral Volcano Valley fault (Godlewski and Zieg, 1984). The 1200-foot-thick upper Newland Formation consists of two carbonate-terrigenous cycles (Zieg, 1986). These record a shoaling upward from deep water turbidite-dominated deposition to subwave base carbonate deposition, to carbonate-dominated deposition within reach of storm-wave base, and finally to current-dominated silty shale deposition typical of the overlying Greyson Formation.

The Volcano Valley fault zone at Sheep Creek (Fig. 2) consists of two south dipping and intersecting fault components, with the steeper angle fault on the north and a shallower angle fault on the south (Fig. 3).

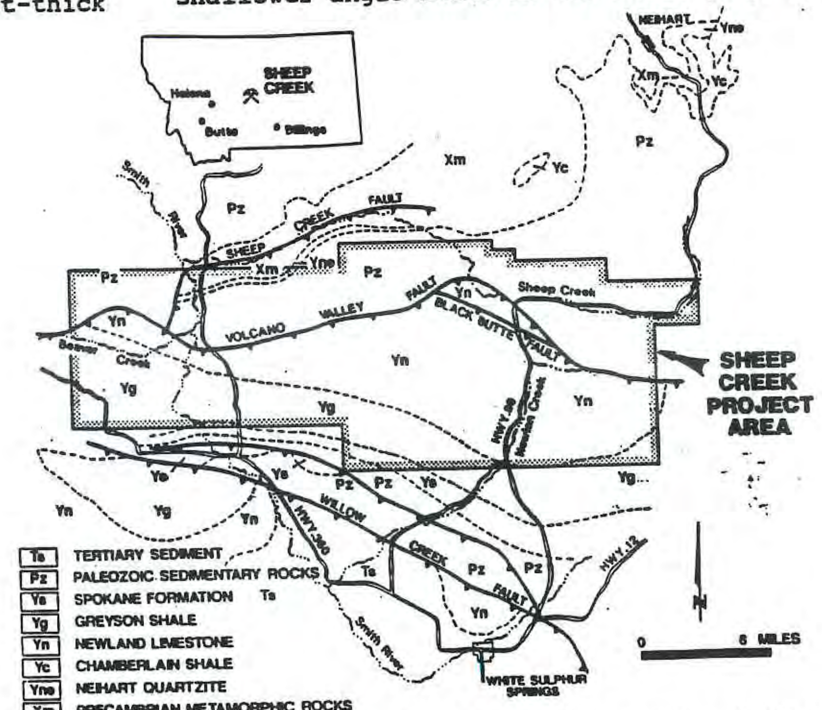

Figure 1

Geology of the Sheep Creek area showing location of project in central Montana. Mineralized area between Volcano Valley and Black Butte fault is enlarged in Fig. 2.

The shallower angle fault occurs both as a double-stranded fault with the greatest translation along its upper strand (Fig. 3), and as a single fault plane. Though Laramide deformation carried Proterozoic rocks from the south and west over Paleozoic strata along the low angle fault strand, in the area north of the Black Butte fault (Fig. 2) both the high and low angle strands of the Volcano Valley fault show normal, dip slip movement relative to Proterozoic stratigraphy. This relationship strongly suggests an original Proterozoic age normal movement along the Volcano Valley fault.

Copper deposits (Fig. 2) occur in syngenetic pyrite layers and within both stratabound and crosscutting zones of silicified host sediment. The most extensive pyrite beds occur near the upper contact of the lower Newland Formation and comprise the upper sulfide zone. Pyrite beds, and in places thin sphalerite and galena beds, are also interspersed throughout lower Newland

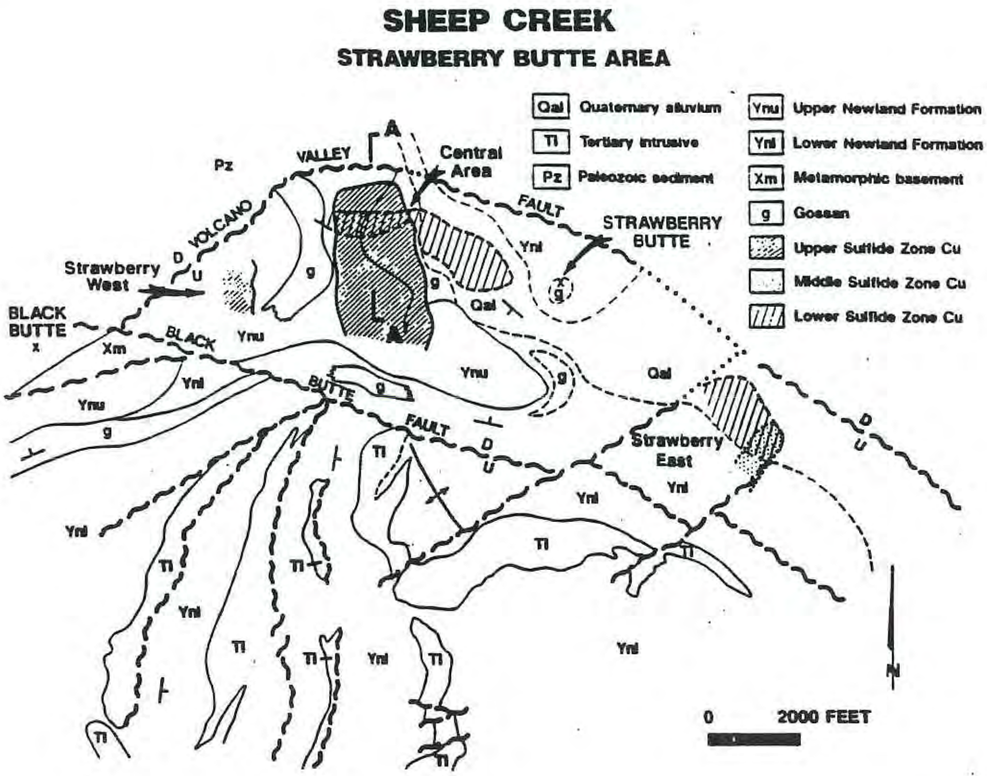

Figure 2

Geology of the Strawberry Butte area and plan locations of copper-enriched parts of the upper, middle, and lower sulfide zones.

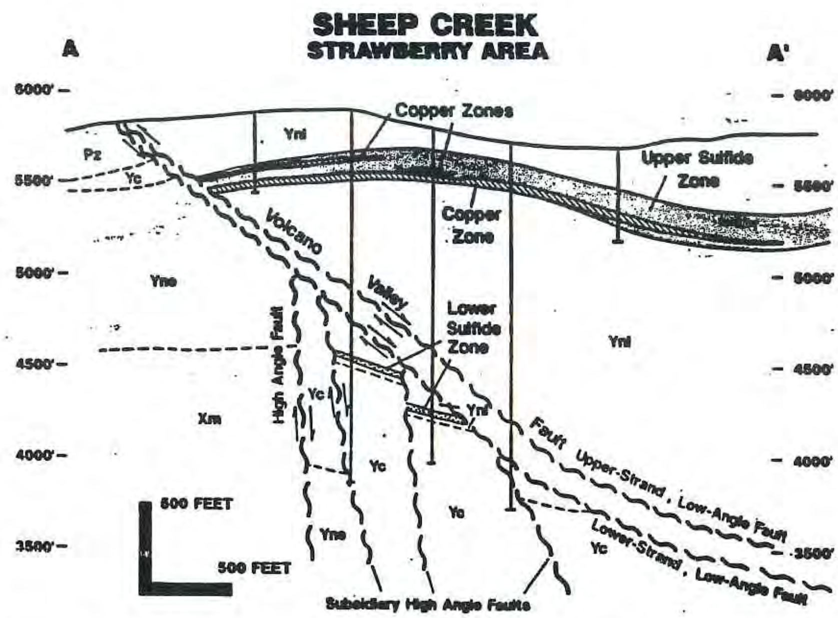

Figure 3

North-south cross-section, viewed east, through central Strawberry mineralized zone showing high-angle and low-angle components of the Volcano Valley Fault zone, and their control on the extent of the sulfide zones (see Fig. 2 for location of section and legend).

and in the lower parts of upper Newland. The nearly continuous blanket of bedded pyrite formed by the upper sulfide zone reaches over 300 feet thick and extends over a strike length of 15 miles and as much as five miles south of the Volcano Valley fault. Replacement deposits of copper sulfide occur locally in syngenetic pyrite layers of the upper sulfide zone, and within both stratabound and crosscutting zones of silicification in two other zones called the middle and lower sulfide zones. A stratabound copper horizon near the base of the upper sulfide zone reaches 40 feet thick and contains disseminated chalcopyrite, barite-chalcopyrite veins and masses, and cobaltiferous bedded pyrite. Spatially restricted concentrations of bedded pyrite overprinted by silicification, silicified fragmental rock, and copper minerals lie in the upper sulfide zone footwall, and are called the middle sulfide zone. This style of mineralization appears to have been controlled by synsedimentary faults and may have developed in areas of feeder zones to upper sulfide zone mineralization.

The lower sulfide zone lies near the base of the lower Newland and consists of stratabound silicification with disseminated clots and stringers of fine-grained pyrite, chalcopyrite, and dolomite porphyroblasts. The lower sulfide zone rests on silicified conglomerate or disturbed silicified shale. Disseminated large porphyrotopes and coarse-grained sheets and masses of diagenetic dolomite form a partial halo above and lateral to the lower sulfide zone. Sulfide and gangue textures suggest initial deposition of syngenetic pyrite beds. Subsequent development of the diagenetic dolomite halo, pyrite replacement of diagenetic dolomite, and finally silicification with chalcopyrite replacement of diagenetic dolomite and pyrite. In places, bedded pyrite sequences are stratigraphically above the lower sulfide zone. The lower sulfide zone contains higher copper grades than the upper sulfide zone, but contains little cobalt.

Silicification, copper mineralization, and in places carbonate alteration and pyritization lie below and often overprint bedded pyrite. Examples are lower sulfide zone silicification, carbonate alteration, copper mineralization, which overprint and are overlain by bedded pyrite zones; middle and upper sulfide zones; and silicified copper mineralization below bedded pyrite at the upper Newland-lower Newland contact. These pairings of syngenetic bedded pyrite and varying degrees of later alteration and mineralization suggest at least three major periods of hydrothermal activity along synsedimentary faults during Newland sedimentation. During each hydrothermal episode, exhaled fluids first precipitated bedded pyrite and later fluids altered and mineralized both sediment and earlier pyrite beds.

A synsedimentary origin for the Sheep Creek bedded pyrite and copper-bearing zones is also supported by evidence from fluid inclusion, sulfur isotope, and lead isotope studies. Fluids trapped in pseudosecondary fluid inclusions (C.H.B. Leitch, S. Hall, and G.A. Zieg, unpub. data) in hydrothermal barite, dolomite, quartz, sphalerite, and calcite average 15 weight percent NaCl equivalent, and range from 7 to 23 weight percent NaCl equivalent. Homogenization temperatures average 230°C and range from 94° to 300°C, with no evidence of boiling. Himes and Peterson (1990) showed similar results. Metastable melting at temperatures as low as -85°C, with eutectic temperatures clustered around -38° and -50°C, suggests the presence of Ca^{++} and possibly Mg^{++} in the fluid inclusions. Sheep Creek stable isotope data (C.H.B. Leitch, S. Hall, and G.A. Zieg, unpub. data) strongly suggest a two-sulfur source: a dominant seawater sulfate source for sulfur in bedded pyrite, which shows a broad range of $\delta^{34}S$ values from -12.1 to 19.7 ‰ CDT, and a deep crustal or magmatic source for Cu in chalcopyrite, which shows a narrower range from -5.1 to 7.1 ‰ CDT. Chalcopyrite in veins and veinlets in the lower sulfide zone shows $\delta^{34}S$ values which cluster even more tightly around 0 ‰. Replacement of pyrite by chalcopyrite results in a much broader range of chalcopyrite $\delta^{34}S$ values, suggesting a sulfur-deficient source fluid. One chalcopyrite-pyrite, sulphur isotopic, pair yielded a temperature of 276°C. Analyses of upper sulfide zone barite show $\delta^{34}S$ values of 13.3 to 16.3 ‰, consistent with the expected value for mid-Proterozoic marine sulfate and with sulfate minerals in the Belt basin. Strauss and Schieber (1990) obtained very similar results for pyrite ($\delta^{34}S$ -14 to 18 ‰) and barite ($\delta^{34}S$ 13.6 to 18.3 ‰). Pb isotope ratios from upper sulfide zone galena samples, collected six miles, apart are as follows:

Pb 206/204:	16.843	16.712
Pb 207/204	15.624	15.550
Pb 208/204	36.563	36.477

These values are consistent with a middle Proterozoic age for Sheep Creek base metal mineralization, and also with the results of Strauss and Schieber (1990).

The Sheep Creek copper deposits are unlike other Belt synsedimentary sulfide occurrences outside the Helena embayment. The middle Proterozoic Mount Isa copper deposits of Queensland, Australia appear the closest analogue to the Sheep Creek lower and middle sulfide zones. No analogy for the barite-capped copper deposits in the upper sulfide zone is presently known.

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