NORTHWEST GEOLOGY

Volume 23 1994

Metallogeny of the Belt-Purcell Basin (Middle Proterozoic): Southern British Columbia and Northern U.S. Rocky Mountains

Proceedings Volume From the Joint Meeting of the British Columbia Geological Survey Branch, Geological Survey of Canada, East Kootenay Chamber of Mines, Cominco Ltd., Consolidated Ramrod, and Tobacco Root Geological Society

September 30 - October 5, 1994

Edited by
David R. Lageson

Ross

- pre-Belt basement rocks
- first reconstructions based on matching up cropper provinces
  1850 to 1840 m.A. most in Boaner
  two Archean grains
- Revett
  - 4 grains 1590 to 1600 m.A.
  - non-North American signature
  - 1892 "Goller" range volcanics
  - also Nd data

- Gawler Volcanics and Hiltaba Suite Granites
- hosted by Gawler volcanics in Adelaide Geosyncline

The Geology of South Australia
Bull. 54

Priest River Complex 1576 ± 15my
on Washington - Idaho border

Rob Cameron - Flathead River Alkalic Epithermal Gold

See pg. 12
pg 26

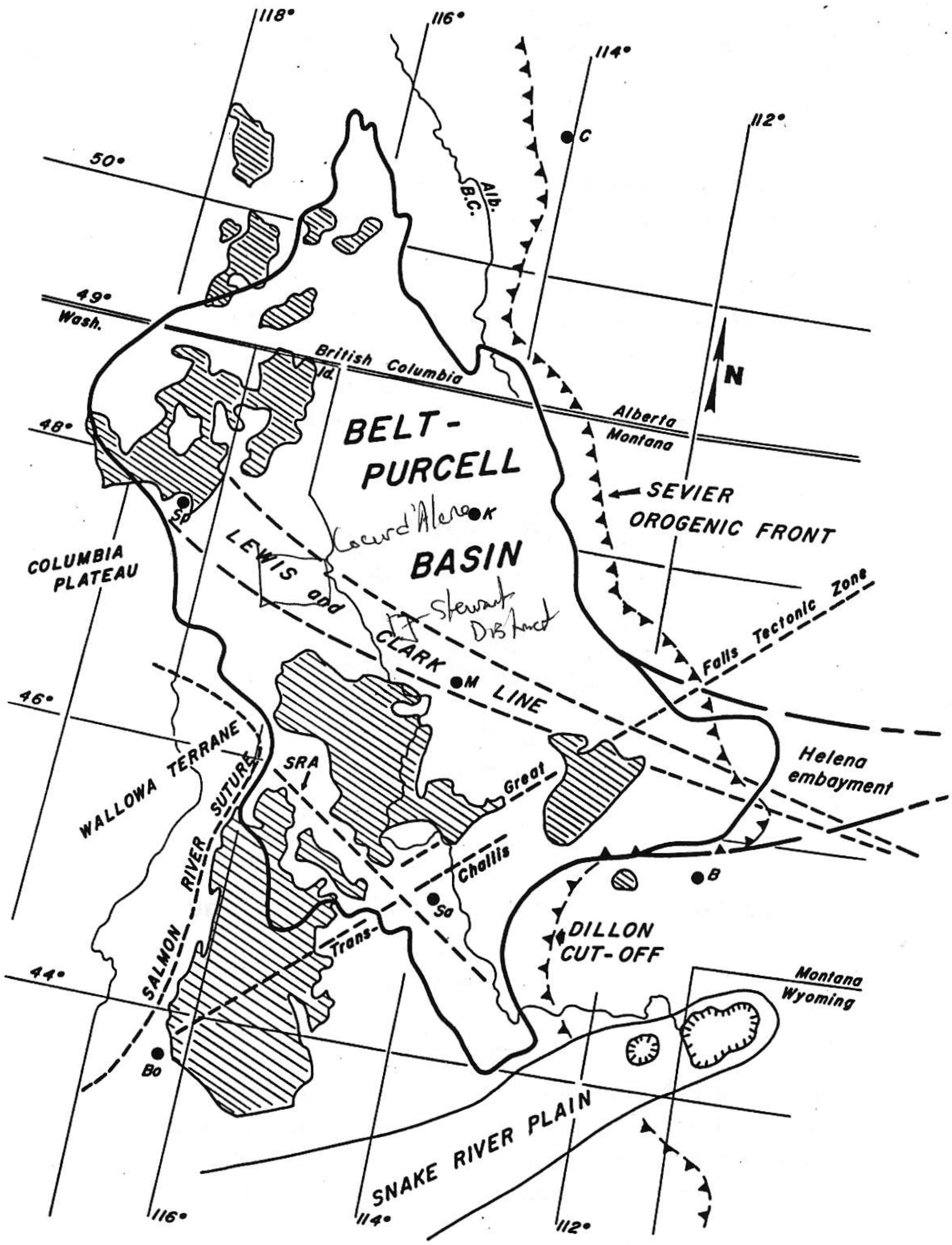
NORTHWEST GEOLOGY

A Refereed Journal Serving the Northern Rocky Mountains, Pacific Northwest and Western Canada

Published by the Department of Earth Sciences, Montana State University, Bozeman, MT 59717

Editor: David R. Lageson
Layout and typesetting: Ann Parker

The Tobacco Root Geological Society, Inc.
P.O. Box 2734
Missoula, Montana



Simplified tectonic map of the Belt-Purcell Basin and surrounding region showing some major structural features. Approximate present extent of the Belt-Purcell Basin is outlined by the heavy dark line. Areas marked by diagonal parallel lines are granitoid bodies, mostly Upper Cretaceous and Tertiary. Great Falls tectonic zone and Trans-Challis are broad zones of deformation and plutonism, the trends of which are shown by dashed lines. B = Bozeman, Bo = Boise, C = Calgary, K = Kalispell, M = Missoula, Sa = Salmon, Sp = Spokane, SRA = Salmon River Arch. Simplified from the _Tectonic Map of North America_ by W.R. Muehlberger (1992). Other sources used include Armstrong (1975), O'Neill and Lopez (1985), Bennett (1986), and Hyndman and Foster (1989). [D.R. Lageson]

Age - < 200 ma $^{87}Sr/^{86}Sr$ ratios
K/Ar — ←→ big spectrum
field geology - Proterozoic

age controversy
two stage lead?
Pb isotope - 1

Rob thinks Mesozoic
Dan Winston agrees

THE COEUR D'ALENE TYPE DEPOSITS: AN ANALOGUE TO TURBIDITE-HOSTED VEIN-TYPE GOLD DEPOSITS

IAN M. LANGE - Department of Geology, University of Montana, Missoula, MT 59812
ROBERT KELL - 10815 Oral Zumwalt Way, Lolo, MT 59847

ABSTRACT

Coeur d'Alene silver-base metal vein and Archean to Phanerozoic turbidite-hosted vein gold deposits have many similar characteristics. Formation of both ore types is compatible with compressional tectonism, commonly obliquely oriented, upon host rocks that were deposited in different tectono/stratigraphic settings.

Mineralizing fluids for both types of deposits were apparently generated by devolatilization reactions under greenschist or higher facies metamorphism during tectonism. Fluids rose into fractures and faults that developed at the ductile-brittle crustal transition depositing ore and gangue minerals, and altering wall rocks.

Compositional differences between the two ore types may be due to metal extraction from different source lithologies both at depth and along the fluid path, and different physical/chemical characteristics of the ore fluid, such as salinity, Eh and pH. Continuing or subsequent compression resulted in unmineralized, major strike-slip faults in some districts which may have offset and/or deformed the veins.

INTRODUCTION

The genesis of the Coeur d'Alene silver-base metal-bearing vein deposits (CDA) in Idaho and Montana (Fig. 1), and mesothermal gold-quartz-carbonate veins in greenschist facies metamorphosed terranes is still vigorously debated. Noteworthy are the many characteristics that these types of deposits have in common in addition to the significant differences. Even more striking are the similarities between CDA and a type of mesothermal gold-bearing veins, the turbidite-hosted gold-quartz-carbonate vein (THG) deposits (Table 1). Recent research on CDA and THG types by different workers has resulted in a better understanding of the timing and mechanisms of deformation and mineralization and component sources.

In particular, the mineral-bearing quartz-carbonate veins commonly form in steeply dipping shear zones and reverse, high angle faults in complexly deformed, siliciclastic rocks from metamorphic-like fluids. Mineralization occurred during syn to post peak regional metamorphism under greenschist facies conditions in regions undergoing compression or transpression apparently related to accretionary tectonism. The mineralizing solutions appear to have been derived at depth by metamorphic devolatization reactions associated with prograde metamorphism. The hydrothermal solutions rose and deposited quartz, carbonate and variable amounts of sulfide and precious metal minerals in the brittle to brittle-ductile crustal transition zone. Research by others suggests to us that these deposit types are analogues. Furthermore, some other deposits described in the literature appear to be very similar to the CDA deposits suggesting that CDA-type veins may not be unique but have sulfide-rich, precious metal-bearing "relatives" elsewhere.

In this report we briefly describe the CDA and gold-bearing turbidity type deposits including

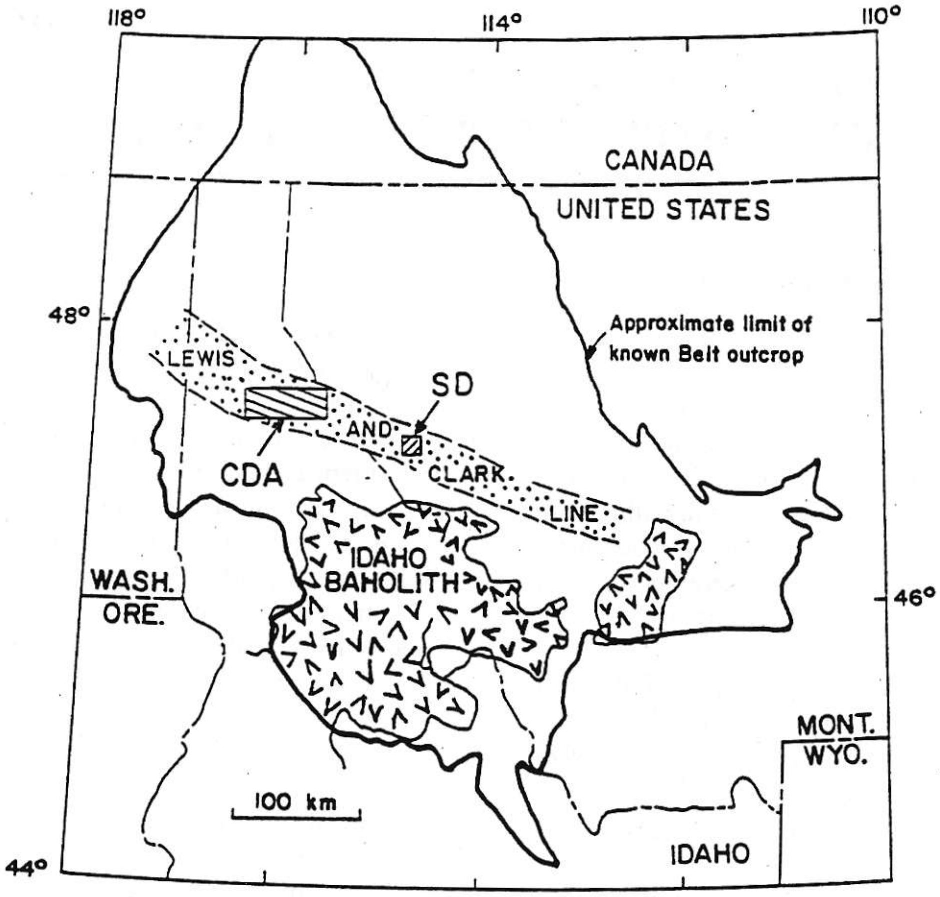


Figure 1. Location of the Coeur d'Alene (CDA) and Superior (δD) districts along the Lewis and Clark line and within the Proterozoic Belt basin of western Montana, northern Idaho, northeastern Washington and southern Alberta, Canada. Intrusive body east of the Lewis and Clark line is the Boulder batholith. (after Harrison, 1972).

their similar characteristics and tectonic settings, discuss their differences, and briefly mention some other deposits that appear similar to the CDA vein systems.

COEUR D'ALENE DEPOSITS

The Coeur d'Alene district, described by Ransome and Caulkins (1908), Fryklund (1964), Hobbs *et al.* (1965), Hobbs and Fryklund (1968), Gott and Cathrall (1980), Reid and Williams (1982), Bennett and Venkatakrishnan (1982), Venkatakrishnam and Bennett, (1988), and White (in press), is the largest silver district in the world. Production has amounted to more than one billion ounces of silver, significant amounts of Pb, Zn and Cu and 500,000 ounces of Au from sulfide mineral-rich quartz-Fe carbonate veins (Leach *et al.*, 1988). The greater CDA district (Fig. 1) is located along a 150 km and most intensely deformed portion of the west-northwest-trending, 500-km long Montana lineament or Lewis and Clark line of the northern Cordilleran fold and thrust belt (Fig. 1).

The Lewis and Clark line is a major 15- to 50-km wide intraplate tectonic boundary apparently intermittently active since the Middle Proterozoic (Harrison, 1972; Reynolds and Kleinkopf, 1977; Winston, 1986; Wallace, *et al.*, 1990). Fine-grained siliciclastic and carbonate rocks of the

- great vertical extent vs horizontal
- lack of mineral zoning
- similar fluid characteristics

- metal characteristics related to region
- magmatic waters for Coeur d'Alene

NB lack of genetically assoc. intrusive rocks with these deposits according to Lange 3

TABLE 1. Characteristics of Coeur d'Alene Silver-Base Metal (CDA) and Turbidity-Hosted Gold (THG) Deposits

Characteristics	CDA	THG
Host Rocks	feldspathic quartzite	turbidite-graywacke
	Proterozoic	Archean-Phanerozic
Tectonic Setting	intraplate	terrane margin[1]
Metamorphic grade	greenschist	greenschist
Igneous Rocks	not related	not related
Mineralization	late to post kinematic	late to post kin.
	>1 kb depth	>1kb to 3 kb depth
Wall Rock Alt.	similar-see text	similar-see text
Vein Location	brittle-ductile trans.	brittle-ductile tr.
	dilatancy zones	dilatancy zones
Vein Character.	faults, shear zones	faults, shear zones
	quartz-Fe carbonate	quartz-Fe carbonate
	sulfide-rich	sulfide-poor
	Pb-Zn-Cu-Ag-As-Sb-Fe (Cd-Ge-Ga-Hg-In-Bi)	Cu-Ag-Zn-Cd-Au-Pb-As-Sb-W-(Hg-In-Tl-Bi-Se-Te-Mo-Co-Ni)
	lack of zoning	lack of zoning
Mineralizing Fluids	5 to 12 wt % $NaCl_{eq}$	< 5 wt % $NaCl_{eq}$
	CO_2-rich	CO_2-rich
	225^0C to $>300^0C$	200^0C to 450^0C
	+/- hydrocarbons	+/- hydrocarbons
	$S^{18}O$, D/H meta.-like	$S^{18}O$, D/H meta-like

Blue Rock - diagenetic sphal.?

Sources: Fryklund, 1964; Boyle, 1986; Leach *et al.*, 1988; Goldfarb *et al.*, 1991; and others in References

1. where known

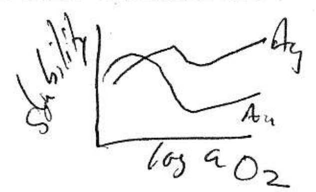
mostly in Revett Form.
- nearby veins in Pritchard Form.
Pb-Zn-Ag versus Coeur d'A. Ag-Pb

Belt Supergroup (Proterozoic Y) metamorphosed to the greenschist facies are deformed into large, upright, north-trending folds north of the Osburn fault within the Lewis and Clark zone. South of the Osburn fault to the northern margin of the Idaho batholith the fold system is complexly segmented, offset, and rotated into west-northwest trends.

The major veins in the district occur within steeply dipping fault/fracture zones that generally parallel west-northwest metamorphic foliation, and commonly flank the axial traces of anticlines and parasitic folds. White (in press) has shown that the metamorphic foliation is characterized by nearly dip-slip shearing lineation, and contains deformed grains of minerals associated with mineralization that are elongated parallel to the lineation. The coincidence of the orientation of these fabrics and mineral belts suggested to White (in press) that vein development accompanied metamorphism and tectonism.

Variable amounts of wall rock alteration and

<5 to 10 wt % NaCl 225 to 350°C
CO_2 rich + hydrocarbons



showsup (sampling problem)

sulfide mineral-bearing quartzite (blue rock) are near or adjacent to many of the veins. Altered wall rock, if present, is not always apparent because of its superposition on quartzites. It may consist of carbonatized, silicified, alkali metasomatized and sulfidized rock (Leach *et al.*, 1988). Details of the district veins are in Table 1.

TURBIDITE-HOSTED GOLD DEPOSIT CHARACTERISTICS

Total gold production from greenstone-hosted deposits including turbidite-hosted deposits is second only to that of early Proterozoic paleoplacers. Production from THG deposits has come from Archean, Proterozoic and Phanerozoic rocks. THG deposits cited by Boyle (1986) include those of the Archean Yellowknife Supergroup in the Northwest Territories and deposits in Proterozoic rocks of the Yenisey region in the former U.S.S.R., and the Pilgrims Rest and Sabie goldfields in the Transvaal System in South Africa. Phanerozoic age deposits are found in the lower Paleozoic Meguma Group in Novia Scotia, in Lower Ordovician rocks in Central Victoria, southeast Australia, in Paleozoic (or Precambrian) rocks of the Reefton goldfields on the South Island, New Zealand, the Paleozoic Cariboo series of British Columbia, and the Appalachian slate belt deposits of North and South Carolina, Georgia and Alabama.

Other THG deposits are those of the Juneau region (Goldfarb *et al.*, 1988) and within the Cretaceous Valdez Group of south-central Alaska (Goldfarb *et al.*, 1986), the Clontibret deposit in Ireland in Ordovician graywackes (Steed and Morris, 1986), the Dolaucothi gold mines of Dyfed, Wales (Annels and Roberts, 1989) and possibly the Proterozoic age Prichard-hosted gold-quartz veins near Murray, Idaho (Ransome and Caulkins, 1908). The following description of the THG type is mostly from Boyle (1986). Table 1 presents a summary of deposit characteristics.

Late kinematic veins are related to regional deformation and formed in dilatant zones at or near the brittle-ductile transition. Host rocks include primarily deep water marine turbidite sequences of graywackes and shales with or without tuffaceous units, volcanic flows, and premineralization mafic and felsic plutonic rocks all metamorphosed to the greenschist facies. The quartz-Fe carbonate veins contain minor amounts of sulfide minerals, primarily pyrite and lesser arsenopyrite, and gold. Wall rock alteration adjacent to the veins is commonly mimimal and where present, may be gold-bearing. Alteration types include silicification, carbonatization, pyritization, chloritization, sericitization and/or argillization.

Deposits may occur along major fault or shear zones (breaks or megalineaments) that parallel or transect supracrustal strata or separate distinct rock assemblages such as in the Juneau district (Goldfarb, 1988, 1991). The breaks are obliquely oriented to, and apparently formed in response to convergent plate tectonism but are not mineralized.

Most of the gold occurs in quartz veins located in bedding planes, faults (typically reverse to reverse-oblique; F. Robert, personal comm.), shears, sheeted zones, and openings in anticlines such as the saddle reefs in the Bendigo gold fields of Australia. Some replacement bodies also exist (Boyle, 1986).

DEFORMATION AND VEIN FORMATION AGE RELATIONSHIPS IN THE COEUR D'ALENE DISTRICT

The ages of deformation and vein formation in the CDA distict have long been debated. Hobbs *et al.*, (1965) formulated the "rotated-fold model". According to the model, the presently northerly striking folds now found only in the northeast part of the district, trended northwest during formation in the Precambrian. They were subsequently rotated in the Cretaceous to a north orientation north of the Osburn fault within the

Lewis and Clark line and to a west-northwest orientation south of the Osburn fault strike-slip movement within the Lewis and Clark line.

Hobbs and Frykund (1968) believed the main veins are most likely Late Cretaceous in age. This interpretation was based on the observations in eight mines of veins cutting quartz monzonite dikes related to the Late Cretaceous Gem stocks. Furthermore, the contact metamorphic aureole effects from the Gem stocks are not observed in the veins.

Subsequent confusion about vein emplacement, however, has arisen in part because of the 1500 to 1200 Ma vein lead model age dates (Zartman and Stacey, 1971) and the post-ore hydrothermal vein sericite K-Ar ages from the Bunker Hill, Lucky Friday Galena, and Sunshine mines of 829 +/- 40 Ma, 876 +/- 43 Ma, 447 +/- 25 Ma and 77 +/- 5 Ma, respectively (Leach *et al.*, 1988). They suggest that the older ages date mineralization; the two youngest ages are either reset or identify younger events. Other workers, however, believe the older age dates reflect Ar absorption during Laramide tectonism/magmatism.

Venkatakrishnan and Bennett (1988) argued that the vein components were derived from stratabound and/or stratiform deposits and migrated into fractures associated with folding related to the East Kootenay orogeny postulated by White (1959) at approximately 829 Ma. They believe the major veins formed during the Laramide orogeny (Late Cretaceous) in response to transpressional shear along west-northwest-oriented faults of the Lewis and Clark line.

Most recently, White (in press) identified five separate tectonic events in the region since Belt sedimentation. Four were influenced by an ancient west-northwest-trending basement structure that also affected Belt sedimentation. The third event produced west-northwest oriented Late Cretaceous age reverse faults, penetrative metamorphic foliation which defines metamorphic shear zones, nearly dip-slip shearing lineation, mineral belts and most of the veins that parallel the Lewis and Clark line.

Strong evidence for this tectonic event including vein formation occurring in the Mesozoic is found in the work of Fleck *et al.* (1991). They state that the high $^{87}Sr/^{86}Sr$ ratios of CDA vein siderite from the Sunshine mine, relative to undisturbed Belt carbonate rocks, could only have resulted from vein formation within the last 200 Ma or less. This Mesozoic age is also compatible with the regional $S^{18}O$ data of Criss and Fleck (1990) and Fleck *et al.* (1991). Criss and Fleck (1990) suggest that it is more than coincidental that the CDA deposits are located in the peripheral zone of steep ^{18}O gradients of the metamorphic hydrothermal system related to the intrusion of the Late Cretaceous northern lobe of the Idaho batholith. The data, they believe, support a genetic link.

The Idaho batholith was apparently emplaced into North American Precambrian crust just east of the major suture between the Seven Devils/Wallowa terrane and North America (Snee *et al.*, 1987). Terrane docking was complete by about 85 Ma to 82 Ma (Snee *et al.*, 1987), and the subduction zone-related Idaho batholith phases were intruded between 80 Ma and 65 Ma (Hyndman and Foster, 1989).

TURBIDITE-HOSTED GOLD-COEUR D'ALENE VEIN SIMILARITIES

Coeur d'Alene and turbidite-hosted gold vein types have many similar characteristics (Table 1). The CDA and most productive THG veins occur within brittle-ductile shear systems hosted by rocks that have been folded and witnessed greenschist grade metamorphism. Mineralization was commonly late or post kinematic, with veins in shears and high angle reverse faults showing complex, anastomosing structures and in some cases post vein, flexural slip margins. Wall rock alteration may be present but is usually subtle.

Juneau, Alaska - included

Within individual vein systems, mineral and/or elemental zoning is not prominent or does not occur. District scale differences, however, may be present. The vertical dimensions of veins commonly greatly exceed horizontal dimensions with some CDA veins extending more than 2,300 m vertically.

Fluid inclusion and stable isotope studies show that the CDA and THG veins formed from similar type fluids with metamorphic characteristics. Temperatures of fluid inclusion homogenization are approximately 225°C to 400°C, the fluids are CO_2-rich, reduced, and may contain hydrocarbon compounds. Fluid trapping pressures ranged from greater than one Kb to three Kb (Leach *et al.*, 1988; Goldfarb *et al.*, 1988).

Oxygen and hydrogen isotopic composition fields for fluids from CDA and THG district vein quartz overlap and fall within the metamorphic field (Fig. 2). However, primary magmatic water values also fall within the field (Fig. 2). The range of CDA δD values extends to isotopically much lighter values than those of metamorphic water. This suggested to Yates and Ripley (1985) and Constantopoulos and Larson (1991) that the mineralizing fluids contain a meteoric component. Alternatively, the light δD values may have resulted from the mixing of both primary and secondary fluid inclusion waters during analysis.

Finally, felsic plutonic rocks occur in the CDA and together with mafic rocks in some of the THG districts. The presence of these rocks has led some workers to propose magmatic metal and fluid sources. However, the available geological and radiometric data indicate the plutons are not of the same age as vein formation.

DISCUSSION

We believe that the available geologic, geochemical and isotopic data support the concept that the CDA and THG-type quartz-carbonate-sulfide-bearing veins are analogues in that they appear to have formed from metamorphically derived fluids that were liberated from sedimentary rocks during compressional tectonism. The fluids rose into faults, commonly high angle and reverse, and shear zones forming veins and altering wall rocks. Some vein systems are located near major breaks or strike-slip faults that resulted from transpressional tectonism. The Juneau district is an example located near an accretionary terrane boundary (Kerrich and Wyman, 1990). The CDA district is located within a major intraplate tectonic zone (Reynolds and Kleinkopf, 1977) affected by late Cretaceous terrane docking accompanied by Idaho batholith magmatism.

In addition to the many similar characteristics, major differences also exist between deposit types. The different characteristics include the relative elemental abundances within the veins, host rock compositions, depositional environments, and tectonic settings. Furthermore, it has not been unequivocally demonstrated that the mineralizing solutions are of metamorphic origin. Finally we discuss whether CDA-type mineralization is unique or if there are other districts which contain similar type mineralization.

Elemental Differences Between CDA and THG Veins

Overlap exists between elemental suites within CDA and THG veins, but their elemental abundances are quite different (Table 1). The CDA veins are silver-, base metal- and sulfur-rich but low in gold. The THG veins, in comparison, are gold rich but sulfide mineral poor. The element suite differences may be due to a number of factors that individually or in combinations affected the respective systems. If, as most workers believe, the hydrothermal fluids were derived at significant depths during metamorphic devoliation reactions (for example see Cox *et al.*, 1991), the composition of

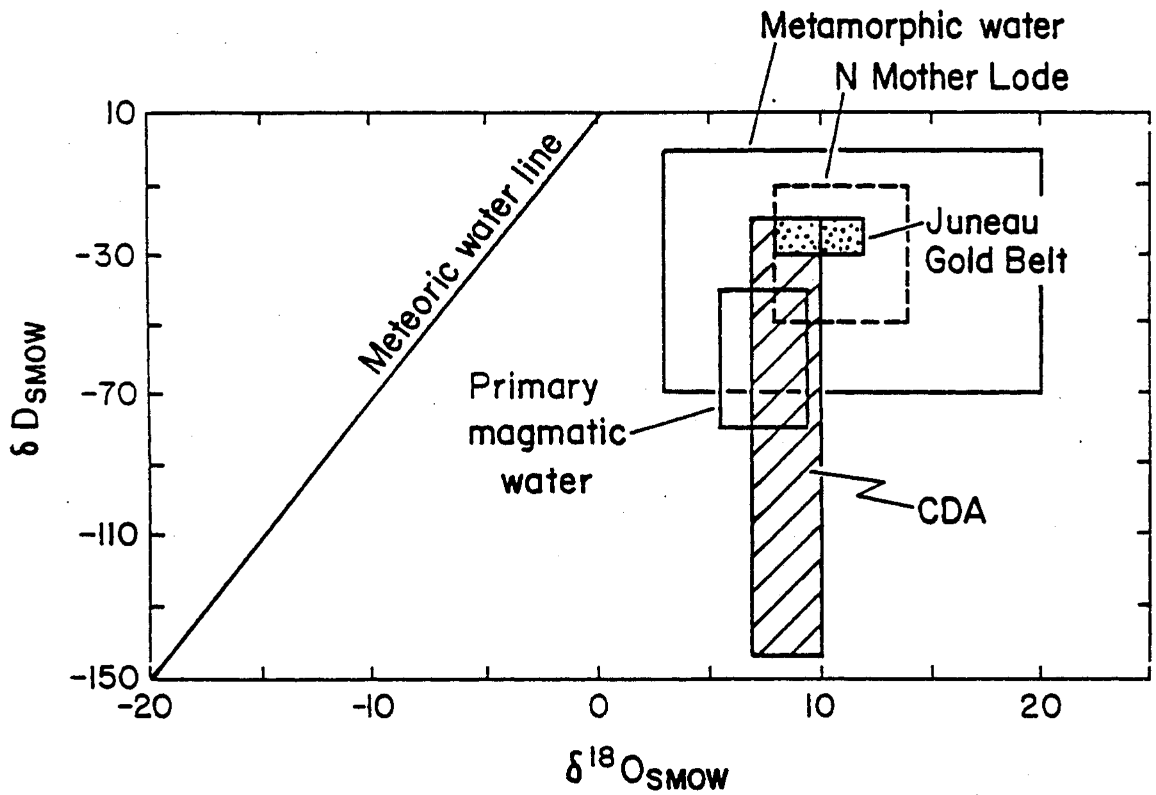


Figure 2. δD and S[18]O isotopic compositions of fluids from the CDA district (Yates and Ripley, 1985; Leach *et al.*, 1988; Constantopoulos and Larson, 1991), Juneau gold belt (Goldfarb *et al.*, 1988; 1991), Northern Mother lode (Bohlke and Kistler, 1986) relative to the present day meteoric water line, metamorphic water and primary magmatic water fields (modified after Goldfarb *et al.*, 1988).

mineralizing fluids would be influenced not only by the deep source rocks, but by interactions with lithologies and fluids encountered along the migratory paths to the deposition sites. For example, Newberry and Brew (1989) describe multiple sources for gold including the sedimentary rocks hosting the veins in the Juneau belt. Beaudoin *et al.* (1991) also describe different sources including the mantle for S, C and Pb within the Eocene age Ag-Pb-Zn veins related to the Vallalha metamorphic core complex in southeastern British Columbia.

In the CDA district lead and probably most or all of the other metals were extracted from the Belt rocks that host the deposits. The isotopic composition and age of the Pb are the same in both the CDA veins and Ravalli Group-hosted diagenetic diseminated Cu-Ag deposits such as the Troy deposit in northwestern Montana (Marvin and Zartman, 1984). Lead, zinc and silver, together with arsenic and the other metals are also found in syngenetic massive sulfide deposits in the Prichard Formation located stratigraphically below Ravalli Group rocks. An example is the Sullivan mine in southern British Columbia. (The lead isotopic ratios from the Sullivan mine are more radiogenetic than Ravalli Group-hosted sulfides (Marvin and Zartman, 1984) so the source of the Sullivan lead was not the source of CDA lead).

Therefore while the CDA mineralizing fluids may have been derived at considerable depth, lead and some or all of the other metals were probably extracted from Ravalli Group rocks. The THG veins, on the other hand, are commonly found in more gold-rich sedimentary rock packages (Boyle, 1986; Graves and Zentilli, 1982; Romberger, 1988; Titley, 1991). Interestingly, the Prichard Formation may be enriched in gold. Most of the placer gold deposits in Belt terrane are on or downstream from Prichard Formation rocks. Furthermore, the gold-bearing quartz veins of the Murray district, located north of the CDA district, are hosted by the Prichard Formation (Ransome and Caulkins, 1908).

Mineralizing solution physical/chemical characteristics may also have contributed to vein compositional differences. Available data show that mineralizing solutions for both vein types are reduced, CO_2-rich and usually contain hydrocarbon compounds such as CH_4. Their temperature formational ranges and oxygen and hydrogen isotopic compositions overlap as do fluid trapping pressures (Leach *et al.*, 1988). Two aspects which may have a bearing on vein element abundances are solution salinity and pH. The fluid inclusions in THG veins generally have five or less equivalent weight per cent NaCl (Boyle, 1986); silver-bearing CDA veins generally show values in excess of five and up to 12 weight per cent (Leach *et al.*, 1988). pH data are unavailable. Both Ag and Au are easily transported by HS^- complexes. Chloride complexes, however, are better transporters of base metals than sulfide complexes, and chloride complex ions become more effective transporters at lower pH, higher salinity and temperatures as the effectivenesss of gold-transporting HS^- decreases. (for example see, Barnes, 1979; Romberger, 1991).

Another possibility for the difference in metal abundances between CDA and THG type veins might be the oxidation state of the metal-bearing source rocks. Base and precious metals are soluble as chloride complex ions in oxidizing solutions. In contrast, the metal suite found with gold and silver in the THG deposits, but not base metals, are soluble at very low oxygen activities in the presence of sulfur (Romberger, 1986). Because both types of mineralizing solutions are anoxic, reduction of the CDA solutions, if they were originally oxic, may have taken place near or at the depositional site.

Tectonic Settings

The host rocks and tectonic settings for the CDA and THG deposits are distinctly different. The CDA vein host rocks include arenite, argillite and carbonate lithologies deposited within an intercratonic, probably shallow water (marine?) basin (Harrison, 1972; Winston, 1986). The THG-bearing turbidites were deposited within deep marine settings (Boyle, 1986). As previously noted, these lithologic differences are probably at least partially responsible for the different metal abundances in the veins. The formation of the CDA veins took place within an intraplate deformational zone. On the other hand, the Juneau district formed near a sutured and faulted terrane boundary and the Central Victoria deposits are related to a collisional event (Wang and White, 1993). (The structural setting of the other THG deposits is not clear). However, what ever the tectonic setting, the geologic relationships support the thesis that both the CDA and THG deposits resulted from metamorphism accompanying compressional tectonism. Tectonism appears to be the result of terrane docking, and a portion of the heat of metamorphism may be derived from syntectonic igneous activity in the respective region.

Source of the Mineralizing Solutions

Were the mineralizing solutions of metamorphic derivation? Oxygen and hydrogen isotope data place the hydrothermal waters mostly within the metamorphic range (Fig. 2). None the less, the metamorphic "box" also includes magmatic water, and some of the CDA δD values could be due to meteoric water contamination (Yates and Ripley, 1985; Constantopoulos and Larson, 1991). Alternatively, the isotopically light δD values may be the result of mixed primary and secondary fluid inclusion waters during processing for analysis.

However, when taken together, the geological, geochemical, and isotopic data are most compatible with metamorphically derived, vein-producing fluids. The data include: (1) mineralization occurred during or after deformation of the host rocks which accompanied regional dynamothermal metamorphism; (2) neither strong P-T gradients, nor significant mineral or element zonation developed over the commonly large vertical dimensions of the veins; (3) numerous veins of similar composition occur in the regions that are not spatially associated with igneous rocks; (4) where igneous rocks are present, geological and radiometric data show vein formation post dated igneous intrusion; (5) the mineral inclusion fluids have metamorphic characteristics; and (6) the available lead isotope data do not support an igneous source for the fluids and may, as in the CDA district (Cannon *et al.*, 1962), support derivation from the Belt Supergroup.

Coeur d'Alene-Like Deposits

The CDA district is apparently unique in regards to the quantity of contained silver within the mesothermal quartz-carbonate veins located in deformed and metamorphosed siliciclastic rocks. However, veins in at least two regions, the Cobar district of Australia and the Lifilian fold belt of easternmost Angola, northwestern Zambia and southern Zaire appear geologically similar to the CDA district veins.

The Cobar Cu-Au, Cu-Pb-Zn and Ag-Pb-Zn veins are structurally controlled within silicified, folded and cleaved Early Devonian thin bedded turbidites. Vein formation was syn to post deformation near the eastern fault-bounded edge of the Early Devonian Cobar basin (Glen, 1987). The area lacks syn deformational igneous rocks, and fluid inclusion studies revealed that mineralization took place from low salinity solutions (Glen, 1987). Glen speculated that the metal source was either the surrounding metasedimetary or basement rocks.

The Pan-African Late Proterozoic Lufilian fold belt hosts both stratiform and vein type mineralization within the Katangan Supergroup. The stratiform mineralization (Cu-Co-Ni-U-Au-Pt with lesser amounts of Se, Ce, Mo, V, and W) is within the Roan Group and was emplaced before basin folding within the outer and middle units of the fold belt (Unrug, 1988).

CDA-like quartz-carbonate veins are located on major faults which cut across fold axes in the outer and inner units of the fold belt. The veins contain Zn, Cu, Pb, U, As, and Au with minor amounts of Cd, Co, Mo, Hg, Ga, Ge, W, Ni, Ag and Rh. Gold veins also occur in stock work and shear zones within the inner tectonic unit. Because the veins post date folding and thrusting, and are higher in the stratigraphic succession than the stratiform deposits, Unrug (1988) calls upon vein formation from metamorphically driven basin dewatering outside the zone of high grade metamorphism and crustal thickening.

Silver and base metal veins are also found in other Belt basin locations within Ravalli and Missoula group rocks. The veins, while not significant metal producers, are usually spatially associated with stratiform Cu-Ag mineralization, the presumed source of the metals (Lange and Sherry, 1983). We speculate that large veins might have formed had these portions of the Belt witnessed the intensity of compressional deformation that occurred in the CDA district along the Lewis and Clark line.

ACKNOWLEDGEMENTS

The authors thank Lee Woodward, University of New Mexico, Jeffrey Mauk, University of Michigan, and Peter Mitchell and Steven R. Maning of Coeur Mines, Reno, NV, for their reviews and helpful suggestions of an earlier draft. Eric S. Cheney, University of Washington and Francois Robert, Geological Survey of Canada also provided many useful comments.

REFERENCES

Annels, A.E., and Roberts, D.E., 1989, Turbidite-hosted gold mineralization at the Dolaucothi gold mines, Dyfed, Wales, United Kingdom: Econ. Geol., v. 84, p. 1293-1314.

Barnes, H.L., 1979, Solubilities of ore minerals, in Barnes, H.L., ed., Geochemistry of Hydrothermal Ore Deposits, second edition, John Wiley, New York, p. 404-460.

Beaudoin, G., Taylor, B.E., and Sangster, D.F., 1991, Silver-lead-zinc veins, metamorphic core complexes, and hydrologic regimes during crustal extension: Geology, v. 19, p. 1217-1220.

Bennett, E.H., and Venkatakrishnan, R., 1982, A palinspastic reconstruction of the Coeur d'Alene mining district based on ore deposits and structural data: Econ. Geol., v. 77, p. 1851-1866.

Bohlke, J.K., and Kistler, R.W., 1986, Rb-Sr, K-Ar, and stable isotope evidence for the ages and sources for fluid components of gold-bearing quartz veins in the northern Sierra Nevada foothills metamorphic belt, California: Econ. Geol., v. 81, p. 196-322.

Boyle, R.W., 1986, Gold deposits in turbidite sequences: their geology, geochemistry, and history of the theories of their origin, in Keppie, J.D. *et al.*, eds., Turbidite-Hosted Gold Deposits, Geol. Assoc. Can. Special paper 32, p. 1-14.

Cannon, R.S., Jr., Pierce, A.P., Antweiler, J.C., and Buck, K.L., 1962, Lead isotope studies in the northern Rockies, U.S.A., in Engel, A.E.J., James, H.L., and Leonard, B.F., eds., Petrologic studies - a volume in honor of A.F. Buddington: Geol. Soc. Am., p. 115-131.

Constantopoulos, J., and Larson, P.B., 1991, Oxygen and hydrogen isotope geochemistry of the Star-Morning mine, Coeur d'Alene mining district, Idaho: Geology, v. 19, p. 131-134.

Cox, S.F., Wall, V.J., Etheridge, M.A., and Potter, T.F., 1991, Deformational and metamorphic processes in the formation of mesothermal vein-hosted gold deposits - examples from the Lachlan fold belt in central Victoria, Australia: Ore Geol.Reviews, v. 6, p. 391-423.

Criss, R.E., and Fleck, R.J., 1990, Oxygen isotopic map of the giant metamorphic-hydrothermal system around the northern part of the Idaho batholith, U.S.A., Applied Geochemistry, v. 5, p. 641-655.

Fleck, R.J., Criss, R.E., Eaton, G.F., Wavra, C., and Bond, W.D., 1991, Strontium and oxygen isotopic characteristics and theage of siderite-tetrahedrite veins in the Sunshine Mine, Coeur d'Alene district, Idaho: Geol. Soc. Am. Abs. withprogram v. 23, p. A173.

Fryklund, V.C., Jr., 1964, Ore deposits of the Coeur d'Alene district, Shoshone County, Idaho: U.S. Geol. Survey Paper 445, 103p.

Glen, R.A., 1987, Copper- and gold-rich deposits in deformed turbidites at Cobar, Australia: their structural control and hydrothermal origin: Econ. Geol., v. 82, p. 124-140.

Goldfarb, R.J., Leach, D.L., Miller, M.L., and Pickthorn, W.J.,1986, Geology, metamorphic setting and genetic constraints of epigenetic lode-gold mineralization within the Cretaceous Valdez Group, south-central Alaska, in Keppie, J.D. *et al.*,eds., Turbidite-Hosted Gold deposits, Geol. Assoc. Can. Special paper 32, p. 87-106.

Goldfarb, R.J., Leach, D.L., Pickthorn, W.J., and Paterson, C.J., 1988, Origin of lode-gold deposits of the Juneau gold belt, southern Alaska: Geology, v. 16, p. 440-443.

Goldfarb, R.J., Newberry, R.J., Pickthorn, W.J., and Gent, C.A., 1991, Oxygen, hydrogen, and sulfur isotope studies in the Juneau gold belt, southeastern Alaska: constraints on the origin of hydrothermal fluids: Econ. Geol., v. 86, p. 66-80.

Gott, G.B., and Cathrall, J.B., 1980, Geochemical-exploration studies in the Coeur d'Alene district, Idaho and montana: U.S. Geol. Survey Professional paper 1116, 63 p.

Graves, M.C., and Zentilli, M., 1982, A review of the geology of gold in Novia Scotia, in Hodder, R.W., and Petrak, W., eds., Geology of Canadian Gold Deposits; Can. Inst. Mining and Metal., Special Vol. 24, p. 233-242.

Harrison, J.E, 1972, Precambrian Belt basin if the northwesternUnited States: its geometry, sedimentation, and copper occurrences: Geol. Soc. Am. Bull., v. 83, p. 1215-1240.

Hobbs, S.W., and Fryklund, V.C., Jr., 1968, The Coeur d'Alene district, Idaho, in Ridge, J.D., ed., Ore Deposits of the United States, 1933-1967, v. 2, A.I.M.E., New York, p. 1417-1435.

Hobbs, S.W., Griggs, A.B., Wallace, R.E., and Campbell, A.B., 1965, Geology of the Coeur d'Alene district, Shoshone County, Idaho: U.S. Geol. Survey Prof. Paper 478, 139p.

Hyndman, D.W., and Foster, D.A., 1989, Plutonism at deep crustal levels: the Idaho batholith, Montana and Idaho: New Mex. Bur. Mines and Min. Resources Mem. 47, p. 255-264.

Kerrich, R., and Wyman, D., 1990, Geodynamic setting of mesothermal gold deposits: an association with accretionary tectonic regimes: Geology, v. 18, p. 882-885.

Lange, I.M., and Sherry, R.A., 1983, Genesis of the sandstone (Revett) type of copper-silver occurrences in the Belt Supergroup of northwestern Montana and northeastern Idaho: Geology, v. 11, p. 643-646.

Leach, D.L., Landis, G.P., and Hofstra, A.H., 1988, Metamorphicorigin of the Coeur d'Alene base- and precious-metal veins in the Belt basin, Idaho and Montana: Geology, v. 16, p. 122- 125.

Marvin, R.F., and Zartman, R.E., 1984, A tabulation of leadratios for lead minerals from mines and prospects in the Belt-Purcell basin in the United States and Canada: U.S. Geol. Survey Open File Report 84-210, 15 p.

Newberry, R.J., and Brew, D.A., 1989, Multiple sources for goldin the Juneau gold belt, Alaska: U.S. Geol. Survey Circular1035, p. 48-49.

Ransome, F.L., and Caulkins, F.C., 1908, The geology and ore deposits of the Coeur d'Alene district, Idaho: U.S. Geol. Survey Prof. Paper 62, 203p.

Reid, R.R., and Williams, G.A., eds., 1982, Society of Economic Geologists' Coeur d'Alene field conference Idaho - 1977: Idaho Bur. Mines and Geol. Bull. 24, 139p.

Reynolds, M.W., and Kleinkopf, M.D., 1977, The Lewis and Clark line, Montana-Idaho: a major intraplate tectonic boundary: Geol. Soc. Am. Abs. with Program, v. 9, p. 1141.

Romberger, S.B., 1986, Ore deposits #9. disseminated gold deposits: Geoscience Canada, v. 13, p. 23-31.

Romberger, S.B., 1988, Geochemistry of gold in hydrothermal deposits in Shawe, D.R., Ashley, R.P., and Carter, L.M.H., eds., Geology and resources of gold in the United States: U.S. Geol. Surv. Bull. 1857-A, p. A9-A25.

Romberger, S.B., 1991, Transport and deposition of precious metals in epithermal deposits, in Raines, G.L. *et al.*, eds., Geology and Ore Deposits of the great Basin Symposium Proceedings, Geol. Soc. Nevada, Reno, p. 219-232.

Snee, L.W., Lund, K., and Davidson, G., 1987, Ages of metamorphism, deformation, and cooling of juxtaposed oceanic and continental rocks near Orofino, Idaho: Geol. Soc. Am., Abstr. with Programs, v. 19, p. 335.

Steed, G.M., and Morris, J.H., 1986, Gold mineralization in Ordovician greywackes at Clontibret, Ireland: in Keppie, J.D. *et al.*, eds., Turbidite-Hosted Gold Deposits, Geol Assoc. Can. Special Paper 32, p. 67-86.

Titley, S.R., 1991, Correspondence of ores of silver and gold with basement terranes in the American southwest: Min. Deposita, v. 26, p66-71.

Unrug, R., 1988, mineralization controls and source of metals in the Lufilian fold belt, Shaba (Zaire), Zambia, and Angola: Econ. geol., v. 83, p. 1247-1258.

Venkatakrishnan, R., and Bennett, E.H., 1988, Structural controls of the Coeur d'Alene ore veins Shoshone County, Idaho: in Kisvarsanyi, G., and Grant, S.K., eds., North american conference on tectonic control of ore deposits and the vertical and horizontal extent of ore systems, proceedings volume, Univ. Missouri-Rolla, MO, p. 125-133.

Wallace, C.A., Lidke, D.J., and Schmidt, R.G., 1990, Faults of the central part of the Lewis and Clark line and fragmentation of the Late Cretaceous foreland basin in west-central Montana: Geol. Soc. Am. Bull., v. 102, p. 1021-1037.

Wang, G.M., and White, S.H., 1993, Gold mineralization in shear zones within a turbidite terrane, examples from Central Victoria, S.E. Australia: Ore Geol. Reviews, v. 8, p. 163-188.

White, B.G., (in press) Tectonically diverse structures of common trend in the Coeur d'Alene mining district: periodic reactivation of an ancient structure, in Berg, R., ed., Belt Symposium III Proceedings, Mont. Bur. Mines and Geol.

White, W.H., 1959, Cordilleran tectonics in British Columbia: Am. Assoc. of Petr. Geol. Bull., v. 43, p. 60-100.

Winston, D.W., 1986, Tectonics and sedimentation of the middle Proterozoic Belt basin, and their influence on Cretaceous compression and Cenozoic extension in western Montana and northern Idaho: in Peterson, J.A., ed., Paleotectonics and Sedimentation in the Rocky Mountain Region, United States, Am. Assoc. Petr. Geol., Memoin 41, p. 87-118.

Yates, M.G., and Ripley, E.M., 1985, Fluid inclusions and isotopic studies of the Galena Mine, Coeur d'Alene district, ID: Geol. Soc. Am. Abstr. with Program, v. 17, p. 756.

Zartman, R.E., and Stacey, J.S., 1971, Lead isotopes and mineralization ages in Belt Supergroup rocks, northwestern Montana and northern Idaho: Econ. Geol., v. 66, p. 849-860.

- superhydrostatic fluids drive mud volcanoes
- sill cuts fragmental unit

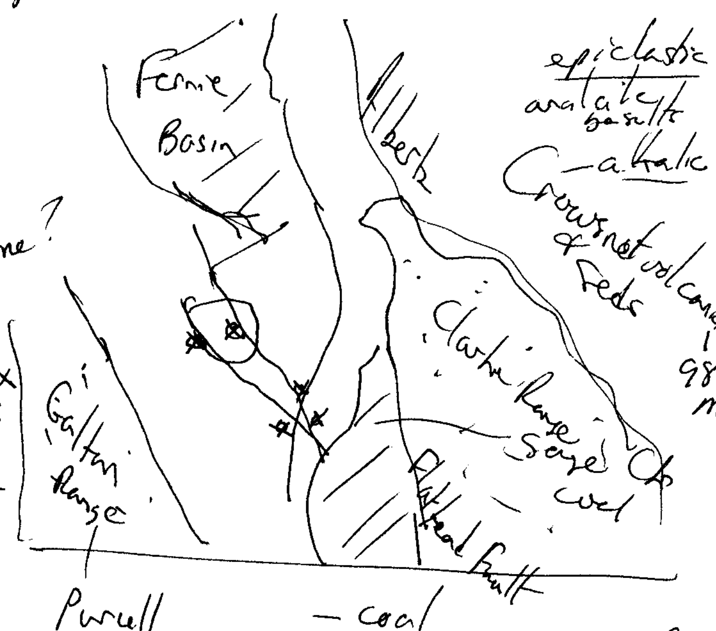
greenbed coppers
revett coppers

Rob Cameron
- Lower Cretaceous Syenite intrusions cutting Purcell ~~Sg~~ SG
- window through Lewis Thrust to Purcell and bounded by younger Tertiary faults

Zortman - Landusky
Golden Sunlight

Purcell stratigraphy from Gateway up to Fairholme? cut by alkalic intrusions and minor in overlying unit
- sills and laccoliths
- soil Au, As, base metals Sb, As

Flathead intrusions × 98.5 ma



- coal
- CO_2 reservoirs for petroleum
- phosphate
- gold

Gov'd A
- some melanogabbro sills

alkalic diatreme breccias

crosscut
- most focus on carbonate - ~~cutt~~ cut by syenite
- volcanic textures - plutonic rocks
- magnetite-bearing mafic gabbros Kspar
- get melanite garnet as accessory minerals analcite syenite

190' 1.3 g/t in silicified lst & altered syenite

- stockwork of qtz veins in centre of largest syenite - faulted porph Mo

BASE-METAL AND PGE MINERALIZATION, SEDIMENTATION AND MAFIC MAGMATISM RELATED TO RIFTING OF THE MIDDLE PROTEROZOIC BELT SUPERGROUP, WESTERN MONTANA

STEPHEN N. BUCKLEY - Flathead Tribes, P.O. Box 278, Pablo, MT 59855
JAMES W. SEARS - Department of Geology, University of Montana, Missoula, MT 59812
PETER C. RYAN - Department of Geology, Dartmouth College, Hanover, NH 03755
D. LAUER - Department of Geology, University of Montana, Missoula, MT 59812

ABSTRACT

The Belt Supergroup in western Montana hosts stratiform Cu-Ag deposits, sediment-hosted Pb-Zn-Ag, and PGE occurrences. The tectono-stratigraphic setting of these deposits suggests they formed as a result of rifting during the deposition of the Middle Proterozoic Belt Supergroup between 1,467 and 1070 Ma. The mineral deposits of the Belt Supergroup share several characteristics with world-class deposits associated with continental rifting.

SEDIMENT-HOSTED MINERALIZATION OF THE PRICHARD FORMATION

Mafic sills intrude the Prichard Formation, the lowest exposed part of the Belt Supergroup. Whole-rock chemical trends and field relations suggest that the sills intruded into wet sediments at shallow depth (Hoy, 1989; Buckley and Sears, in press). Mapping of the sill-sediment complex shows that sill intrusion and sedimentation were contemporaneous with extensional faulting, subsidence and the development of fluid overpressures within the basin. Stagnant basinal conditions and tourmalinite formation accompanied widespread deposition of base-metal sulfides. Heat flow modelling based on metamorphic isograds indicates extremely high geothermal gradients during rifting and sedimentation (Poage and Sears, 1994). Biotite and garnet isograds suggest the zone of highest heat flow occurred along the axis of the rift basin.

Palinspastic restoration of the Belt terrane with 30 degrees counter-clockwise rotation about a pole south of Helena, MT suggests an intra-continental rift setting for the Belt basin analogous to the midcontinental rift (Sears and others, 1994; Sears, this volume). The thickest parts of the sill-sediment complex outline the center of the rift basin (Fig. 1). An abrupt change in sediment facies the Lower to Middle Prichard boundary correlates in time with sill intrusion. This change from shallow to deep water deposition is shown on the palinspastic restoration to have occurred throughout most of the basin. Cressman (1989) interpreted 3 km of tectonic subsidence during this stratigraphic interval based on stratigraphic evidence. We suggest that extension, combined with sill intrusion, led to basin-wide subsidence. This event resulted in the development of fluid overpressures within the basin that drove hot sulfide-rich fluids into zones of cross-stratal permeability. These zones are now sulfitic breccias which cut bedding at a low angle (Cressman, 1985). The fluids moved through permeable sand layers in the Prichard Member E, near Perma, MT, depositing disseminated pyrite, galena, sphalerite and minor chalcopyrite.

Northwest Geology, v. 23, p. 13-18, 1994

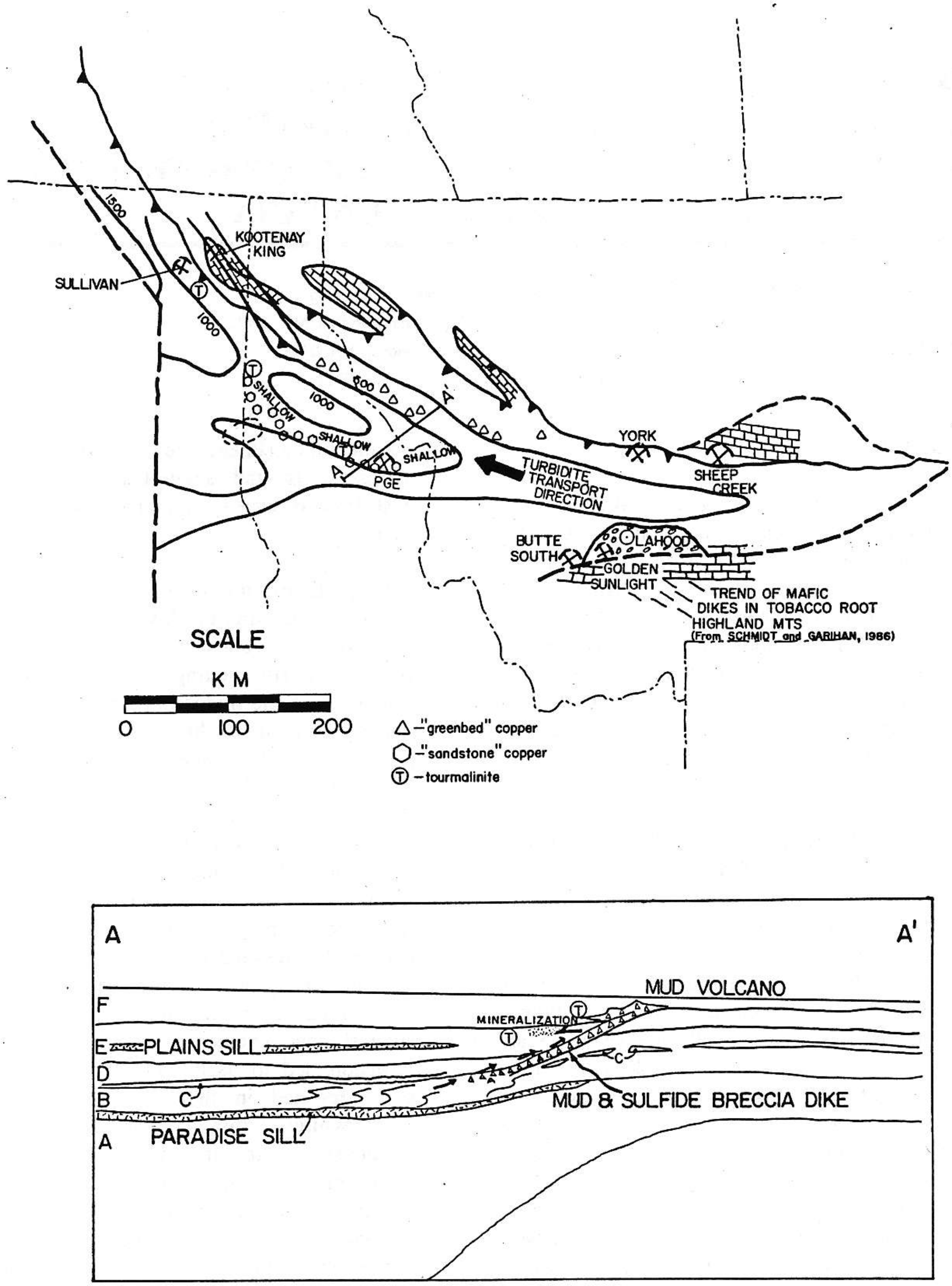


Figure 1. Model showing the restored Prichard basin with sill isopachs, turbidity transport direction and mineral deposits. Refer to Sears (this volume) for a detailed discussion of the basin restoration.

Figure 1 illustrates the relationship between sill intrusion, basin subsidence, fluid over-pressurization and base-metal mineralization in a cross-section through the restored Prichard rift-basin.

Sullivan, Red Dog, Mt. Isa, and Rammelsberg are well known deposits which share rift-related tectono-stratigraphic characteristics (Sawkins, 1990). These deposits occur within intra-continental rifts, in thick to very thick clastic sequences (5-20 km). They occur along linear zones of long-lived subsidence with contemporaneous extensional faulting. Black shale, siltstone, carbonate and turbidite host rocks indicate stagnant basinal conditions. The host sequences contain little or no volcanic rocks and exhibit indications of high geothermal gradients during sedimentation. The deposits are proximal to basin-bounding structures reflected by facies changes, thickness variations and intra-formational breccia. Ore is massive to semi-massive, conformable, and occurs in stacked lenses. Mineral zonation is characterized by an increase in the Zn/Pb ratio towards the lateral and vertical margins of the deposits.

STRATA BOUND COPPER-SILVER MINERALIZATION OF THE RAVALLI GROUP

The Ravalli Group overlies the Prichard Formation and hosts the stratiform Cu-Ag deposits of the Western Montana Copper Belt (Harrison, 1972; Lange and Sherry, 1985). The Revett Formation, the middle part of the Ravalli Group, is a sequence of interbedded quartzite, siltite and argillite (White and others, 1977). Sedimentary structures and facies changes suggest that the Revett Formation was deposited as a sequence of sheetflood deposits across broad alluvial fans onto shallow mudflats (Young and Winston, 1990; Ryan and Buckley, in press). Lange and Sherry (1985), and Ryan and Buckley (in press) showed the importance of syndepositional structure in the localization of ores in the Western Montana Copper Belt. The deposits are comprised of disseminated chalcocite and bornite and show a well-developed mineral zonation (Hayes and Einaudi, 1986). They occur in reduced beds associated with oxidized clastics often associated with dewatering features. The deposits appear to be related to zones of tensional faulting and occur in mineral belts aligned parallel to isopachs and perpendicular to paleocurrents. Figure 2 shows the relationship between the underlying mafic sills, facies patterns, proposed syndepositional structures, and Cu-Ag mineralization in a cross-section of the restored basin. The configuration of the Prichard rift-basin appears to have controlled the location of the strata-bound copper-silver deposits. Both the "greenbed" and "Revett" types of deposits follow the sill isopach pattern shown on Fig. 1.

The Cu-Ag deposits probably formed diagenetically when oxidized basinal fluids, migrating out of the basin, scavenged metals from the underlying mafic sills and oxidized sediments and deposited them at redox/Eh boundaries.

Well known examples of rift-related stratiform copper deposits include Spar Lake, White Pine, Kupferschiefer and the Zambian copper belt (Sawkins, 1990). Characteristics of these types of deposits include the disseminated nature of the ores in which bornite and chalcocite predominate. Reduced beds associated with oxidized clastics localize the ores at abrupt Redox/Eh boundaries. The deposits are underlain by mafic igneous rocks in sequences which contain evidence of high geothermal gradients during sediment deposition. Many of the deposits underly a transgressive caprock sequence such as the Empire Formation of the Belt Supergroup. Orebodies tend to be localized along zones of tensional faulting, and have high Cu/Fe ratios with accessory Ag and/or Co. Mineral zonation is well defined, consisting of Bn-Cpy-Py-Gn/Sph towards the margins of the deposit. Table 1 summarizes the characteristics of mineral deposits related to rifting (Sawkins, 1990).

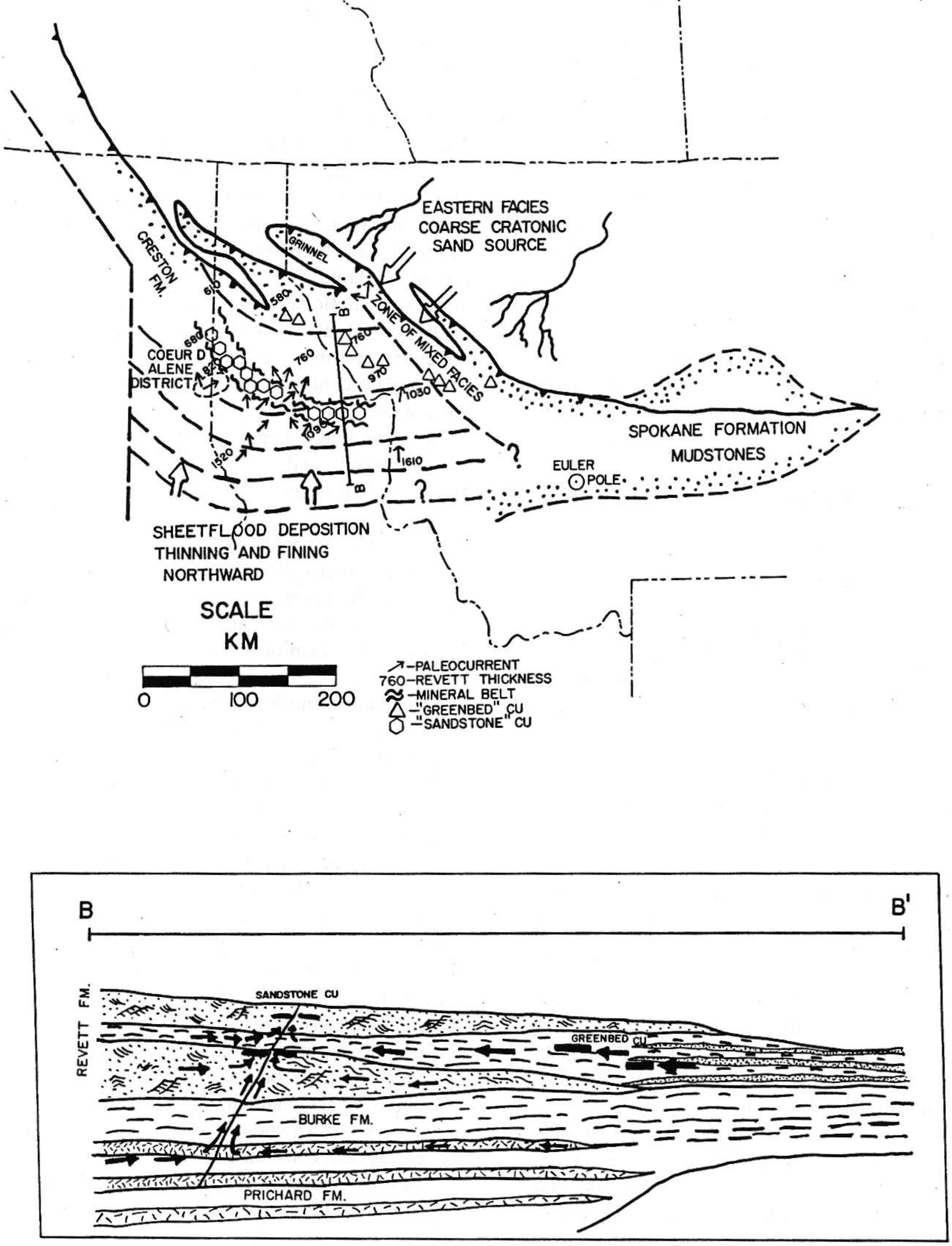


Figure 2. Model of restored Ravalli Group with isopachs, paleocurrent data and mineral occurrences.

Table 1.

COMMON CHARACTERISTICS OF SEDIMENT-HOSTED MASSIVE SULFIDE DEPOSITS	COMMON CHARACTERISTICS OF STRATIFORM COPPER DEPOSITS
Intracontinental Rifts	Intracontinental Rifts
High Heat Flow	High Heat Flow
Linear zones of long-lived subsidence with extensional faulting	Linear zones of extensional faulting
Thick to very thick clastic sequences	Oxidized clastic sequences
Black shale, siltstone, turbidity hosts indicating stagnant basinal conditions	Mafic igneous rock in the underlying section
Massive to semi-massive conformable sulfide ores in stacked lenses	Disseminated ores where chalcocite and bornite predominate
Lateral and vertical metal zonation (Zn/Pb increase towards the margin)	Well-defined mineral zonation (Bn-Cpy-Py-Pb/Zn) towards the margins, High Cu/Fe
Proximity to basin bounding structures, reflected by facies changes, thickness variations, intraformational breccias	Location at abrupt redox/Eh boundaries
Lateral metal zonation	Reduced Beds associated with oxidized clastics

Platinum Group Metals

The Revais Mining District, south of Dixon, MT hosts deposits of Platinum Group Elements in gabbro which intrudes the Ravalli Group (Crowley, 1963; Buckley, 1992). PGE minerals, including sperrylite, michnerite, and atokite occur disseminated within the gabbro along with bornite, chalcocite and chalcopyrite. Grades exceed 0.1 opt combined PGE over 3 m along 8 km of strike.

The Revais occurrences contain characteristics of both magmatic and hydrothermal gabbro-hosted PGE deposits. The gabbro lacks layering and exhibits both dike and sill geometry. The occurrences are Cu, Pt and Pd-rich and Rv, Ir, Os and Ni-poor possibly indicating a partial melt from pristine subcontinental mantle (Naldrett and Duke, 1980). The gabbro post-dates deposition of the Ravalli Group and may correlate with a younger magmatic pulse associated with rifting at the end of Middle Belt Carbonate deposition (1070 to 1,200 Ma), coeval with the Purcell Lava, and with a suite of mafic dikes in the Archean basement of southwest Montana (Schmidt and Garihan, 1986). Magmatic deposits of Cu, Ni and PGE such as Noril'sk, Duluth complex (midcontinent rift) and the Insizwa intrusion (southern Africa) are all related to mafic magnetism in intracontinental rifts (Jowett, 1989).

REFERENCES

Buckley, S.N., 1992, Geologic description of the Flathead Indian Reservation: U.S. Bureau of Indian Affairs General Publication G-92-2, p. 37-47.

Buckley, S.N., and Sears, J.W., 1995, Emplacement of sills into wet Belt Supergroup sediments at Perma, western Montana: Montana Bureau of Mines and Geology, Belt Symposium III, (in press).

Cressman, E.A., 1985, The Prichard Formation of the lower part of the Belt Supergroup, Proterozoic Y, near Plains, Sanders County, Montana: U.S. Geological Survey Bulletin 1553, 64 p.

Cressman, E.R., 1989, Reconnaissance stratigraphy of the Prichard Formation (Middle Proterozoic) and the early development of the Belt Basin, Washington, Idaho, and Montana: U.S. Geological Survey Professional Paper 1490, 80 p.

Crowley, F.A., 1963, Mines and mineral deposits (excluding fuels), Sanders County, Montana: Montana Bureau of Mines and Geology, Bulletin 34, 58 p.

Harrison, J.E., 1972, Precambrian Belt basin of northwestern United States- its geometry, sedimentation, and copper occurrences: Geological Society of America Bulletin, v. 83, p. 1215-1240.

Hayes, T.S. and Einaudi, M.T., 1986, Genesis of the Spar Lake stratabound copper-silver deposit, Montana: Part I. Controls inherited from sedimentation and pre-ore diagenesis: Economic Geology, v. 181, p. 1899-1931.

Hoy, Trygve, 1989, The age, chemistry, and tectonic setting of the Middle Proterozoic Moyie sills, Purcell Supergroup, southeastern British Columbia: Canadian Journal of Earth Sciences, v. 26, p. 2305-2317.

Jowett, E.C., 1989, Effects of continental rifting on the location and genesis of stratiform copper-silver deposits: Geol. Assoc. Canada Special Paper 36, p. 53-66.

Lange, I.M. and Sherry, R.A., 1983, Genesis of the sandstone (Revett) type of copper-silver occurrences in the Belt Supergroup of northwestern Montana and northeastern Idaho: Geology, v. 11, p. 643-646.

Naldrett, A.J. and Duke, J.M., 1980, Platinum metals in magmatic sulfide ores: Science, v. 208, p. 1417-1428.

Poage, M.A. and Sears, J.W., 1994, Thermal history model for the Middle Proterozoic Belt Basin: implications for large static heat flow at depth: Geological Society of America Abstracts with Programs, v. 26, no. 6, p. 59.

Ryan, P.C. and Buckley, S.N., 1995, Sedimentation, stratabound Cu-Ag Mineralization, and syndepostional tectonics in the Revett Formation, Flathead Indian Reservation, western Montana: Montana Bureau of Mines and Geology, Belt Symposium III (in press).

Sawkins, F., 1990, Plate tectonics and mineral deposits: Wiley and Sons, NY.

Schmidt, C.J. and Garihan, J.M., 1986, Middle Proterozoic and Laramide tectonic activity along the southern margin of the Belt basin: Montana Bureau of Mines and Geology, Spec. Pub. 94, p. 217-235.

Sears, J.W., Jacob, J.P., Poage, M.A., Sims, J.L., Skinner, L.L. and Buckley, S.N., 1994, Mid-continent rift analog for Middle Proterozoic Belt Basin: Geological Society of America Abstracts with Program v. 26, no. 6, p. 62.

Sears, J.W., 1994, Thrust rotation of the Belt basin, United States and Canada: (this volume).

White, B.G., Winston, D., and Jacob, P.J., 1977, The Revett Formation near Kellogg, Idaho: Geological Society of America, Abstracts with Program, v. 9, p. 773.

Young, A.A. and Winston, D., 1990, Sheetflood deposition in the Revett Formation (Middle Proterozoic Belt Supergroup), northwestern Montana: Geological Society of America, Abstracts with programs, v. 22, p. 51.

THE VENT COMPLEX OF THE SULLIVAN STRATIFORM SEDIMENT-HOSTED Zn-Pb DEPOSIT, B.C.

CRAIG H.B. LEITCH - Geological Survey of Canada, Mineral Resources Division, 100 West Pender Street, Vancouver, B.C. Canada
ROBERT J.W. TURNER - Geological Survey of Canada, Mineral Resources Division, 100 West Pender Street, Vancouver, B.C. Canada

In the vent zone, the western massive sulphide portion of the Sullivan deposit is underlain by an extensive, in places intensely developed, pyrrhotite ± quartz-Fe carbonate stringer network in tourmalinite (Figs. 1, 2A). This probably represents the major fluid upflow zone that formed the deposit. 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/or carbonate that appear to have formed at later stages within a more indurated feeder zone. A crescent-shaped zone around the margins of the tourmalinite pipe is characterized by the presence of sphalerite and galena in the veinlets (Fig. 2A), with associated tourmaline-destructive muscovite or chlorite alteration (Fig. 2B). This zone may represent the site of late-stage fluid flow after sealing of the main central conduit (Fig. 3). Chlorite and albite-chlorite-pyrite alteration in the footwall and hangingwall (Figs. 2B, 3) may be younger, related to hydrothermal flow generated during emplacement of Moyie sills and dykes that are unusually abundant in the footwall of the deposit (Turner and Leitch, 1992).

Quartz, and to a lesser extent sphalerite, carbonate and cassiterite, found in the footwall network of pyrrhotite veins, contain abundant pseudosecondary or secondary fluid inclusions (Leitch, 1992a). These inclusions are not seen in wallrock detrital quartz grains, which do not appear significantly recrystallized by greenschist metamorphism at 350-450°C and 2 kb (McMechan and Price, 1982; De Paoli and Pattison, 1994). Fluids trapped in the inclusions are probably samples of the mineralizing fluids, in places diluted and/or reset by metamorphism. The mineralizing fluids are characterized as moderately saline brines, 15-35 wt % ($NaCl+CaCl_2+?MgCl_2$), either Type 1 halite saturated or Type 2a undersaturated (Fig. 4). There is a possible transition to less saline 10-15 wt % (NaCl+?KCl) Type 2b fluids found in both primary inclusions in 200-500 micron tourmaline crystals formed during albite-chlorite alteration and pseudosecondary inclusions in quartz associated with the albite-chlorite alteration. Low temperature secondary inclusions in quartz, albite and sphalerite range from 3 to 20 wt % NaCl+?KCl, Th 90-150°C; the origin of these fluids is uncertain. Homogenization temperatures (Th) of the Type 1 and 2 inclusions range from 200-300°C; if, as the primary inclusions in tourmaline suggest, these reflect temperatures of venting fluids, about 30°C would need to be added to give trapping temperatures assuming water depth of 2000 m and 15 wt % salinity; if water depth was 200 m and salinity 25%, 20°C would need to be added. If Th had been reset by metamorphism (not considered likely), a pressure correction of 125°C at 1.5 kb, or 175°C at 2 kb would give Tt of 325-475°C, in agreement with studies of metamorphic equilibria (De Paoli and Pattison, 1994). Type 1 fluids are also found in inclusions in quartz at other prospects (North Star, Quantrell, St. Joe, Kidstar) in Aldridge rocks, and at the stratabound Iron Creek Cu-Co deposit in possibly equivalent age, greenschist facies rocks of Idaho (Leitch, Unpub. data). Slightly less saline brines are also

Northwest Geology, v. 23, p. 19-26, 1994

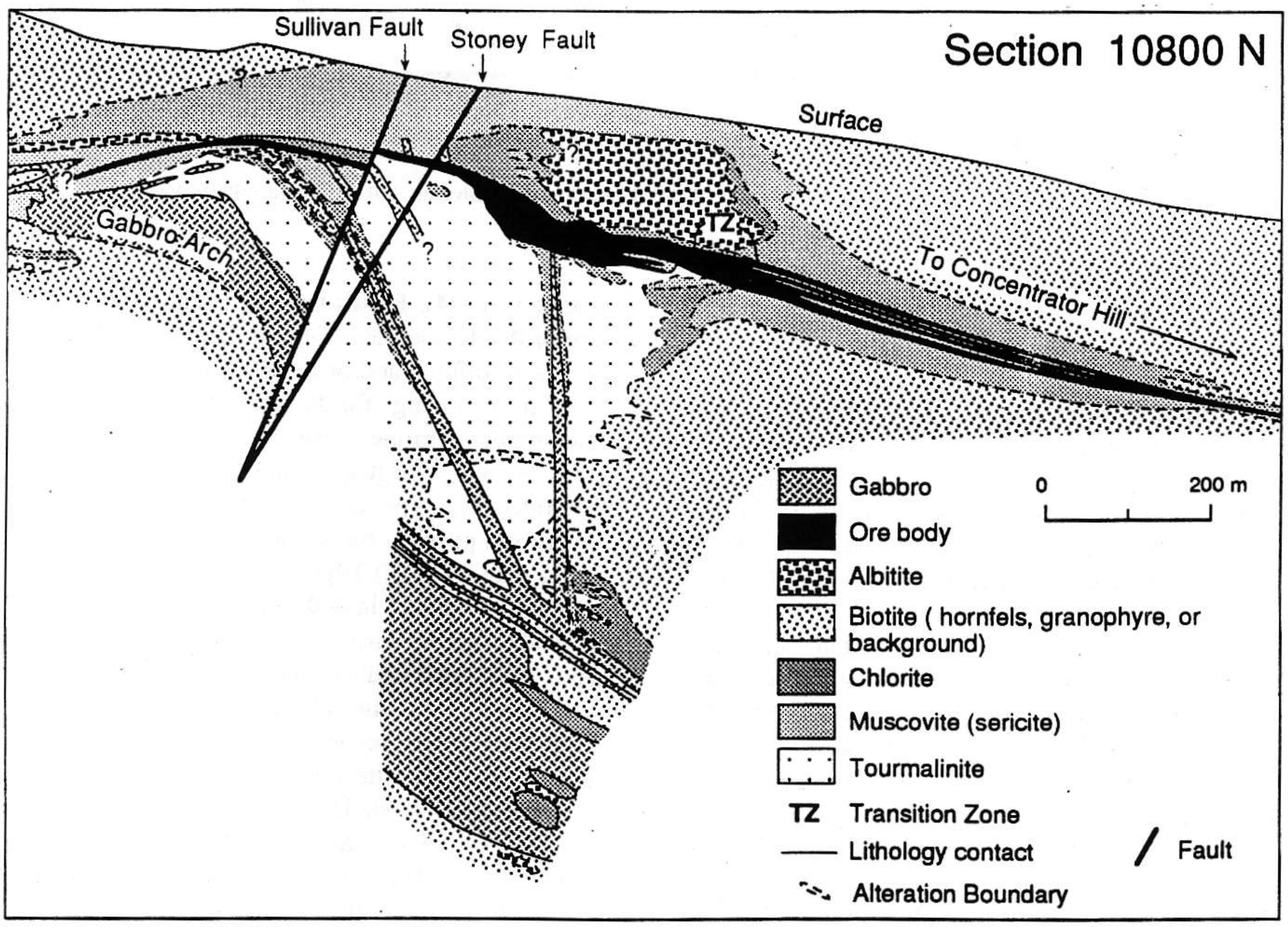


Figure 1. Geologic cross-section of the Sullivan deposit, showing extent of alteration types.

contained in fluid inclusions at the Sheep Creek stratabound Cu-Co deposit in equivalent age but unmetamorphosed rocks in Montana (Zieg and Leitch, 1993; Leitch *et al.*, in review). At Sullivan, secondary Type 2c, 3 and 4 inclusions in quartz and carbonate are dilute (< 5 wt % NaCl equivalent), with mainly low but variable CO_2+CH_4, and Th 200-350°C (Fig. 4). These low-salinity, carbonic-bearing inclusions are also found elsewhere in planar quartz veins that cut a) Mesozoic folds and b) Cretaceous stocks, suggesting they likely represent a young metamorphic overprint.

Note clay-rich (Kaolin) sediments went to muscovite in metamorphism while get biotite and feldspar in Aldridge tourmalinite tpo – down ~~#~~ 400m

At the Sullivan deposit, bedded sulphides are separated from a lens of massive sulphides by the arcuate transition zone around the eastern margin of the major footwall tourmalinite-pyrrhotite vent zone. Several minor minerals containing Sn, Ag and the semi-metals As, Sb and Bi (cassiterite, stannite, freibergite, arsenopyrite, bismuthian boulangerite and jamesonite, bournonite, gudmundite, Bi-Sb alloy, and rare molybdenite and ?bismuthinite) appear to be concentrated in the transition zone in late-stage veins associated with muscovite

- tourmalinite ± po ± qtz ± sphal ± gal
- feeder zone
- feeder zone centred on margin - Pb-Zn
- chl-po in footwall more extensive and earlier than footwall albite

- lots of tourmalinite in hangingwall - up to 200m above

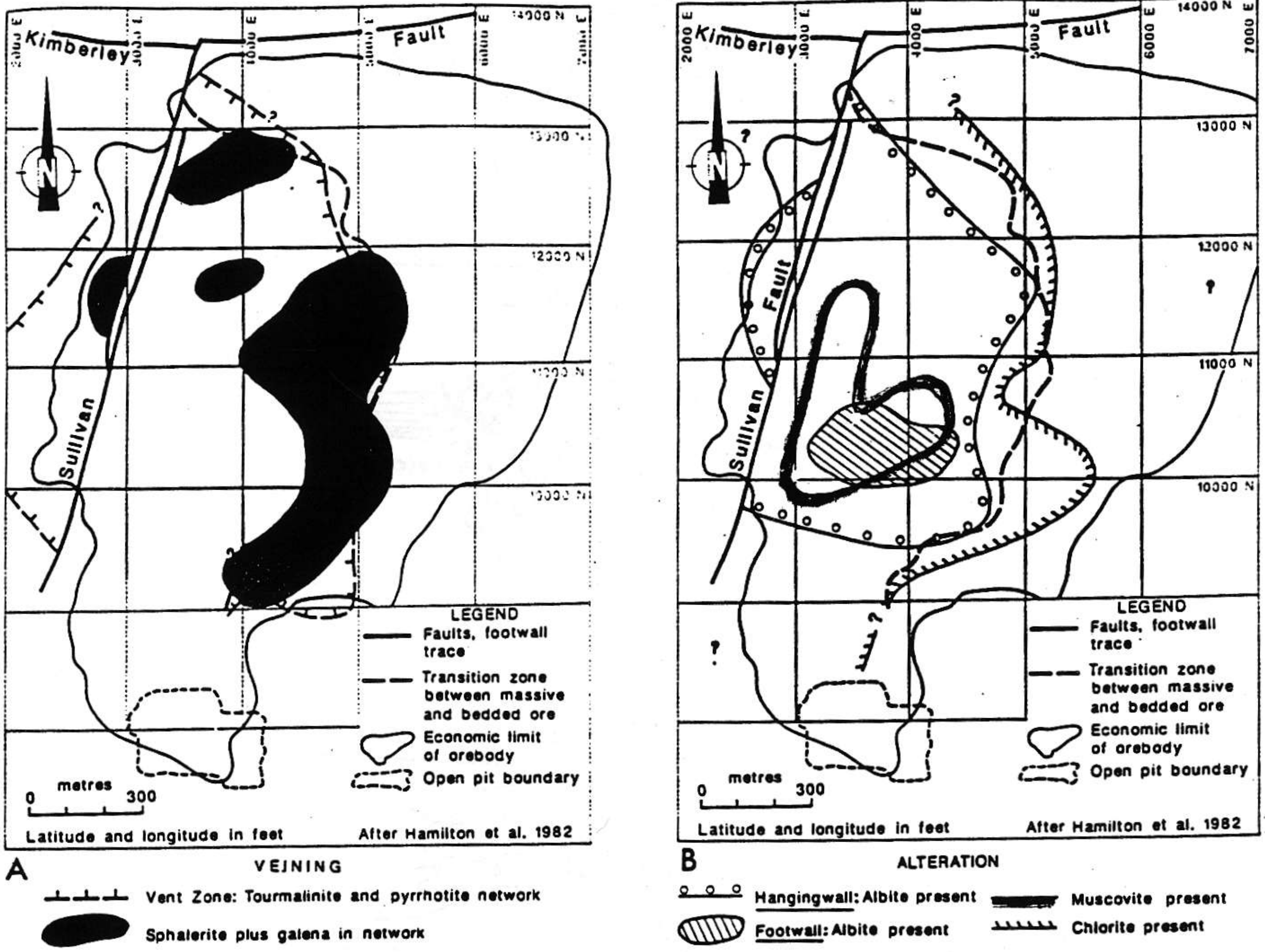


Figure 2. Plan views of the Sullivan deposit showing distribution of A) pyrrhotite-rich and sphalerite-galena-rich footwall mineralization, and B) albite-chlorite-pyrite alteration separately for hangingwall and footwall, and muscovite (shaded) and chlorite (inside hachured line) alteration from footwall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

or chlorite alteration (Leitch, 1992b). Zoning of trace metals As, Sb, Ag and Sn (Fig. 5) appears, at least in part, to reflect this distribution of minor minerals although Ag may be largely controlled by the distribution of galena.

Muscovite alteration is common in and around the Sullivan and the adjacent smaller Stemwinder and North Star deposits, forming extensive zones both crosscutting stratigraphy and enveloping each deposit (Fig. 1). Stratabound muscovite-rich rocks envelope the mineralized Sullivan horizon as far away as Concentrator Hill, 5 km southeast of the deposit. Muscovite is found regionally in the greenschist facies rocks, but appears to increase near the deposits at the expense of first feldspar (i.e. replacement of detrital grains during alteration) and then of biotite (i.e. lack of biotite growth during regional metamorphism), coincident with an increase in

py+chl
&
po+chl
(older)

Fe Core
- Py Calcite

- clasts of qtz veins in pyrrhotite massive sulphide due to Mesozoic deformation

saline solutions -15-27% NaCl+CaCl2+MgCl2

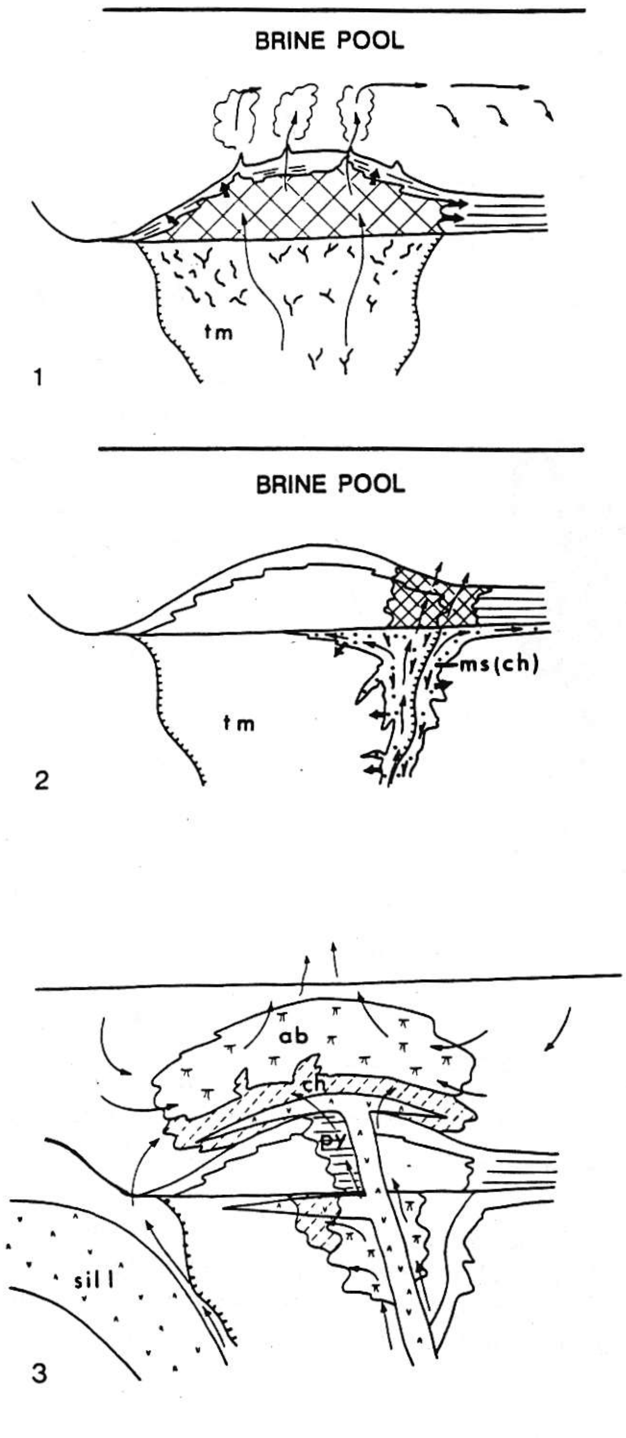


Figure 3. Postulated evolution of the hydrothermal system, Sullivan deposit (from Leitch and Turner, 1992). (1) Main stage hydrothermal flow results in sulphide sedimentation in a brine pool, with progressive replacement of sulphides by massive pyrrhotite (crosshatched), and underlying pyrrhotite network/tourmalinite (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 (ms) and ?later chlorite (ch). (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.

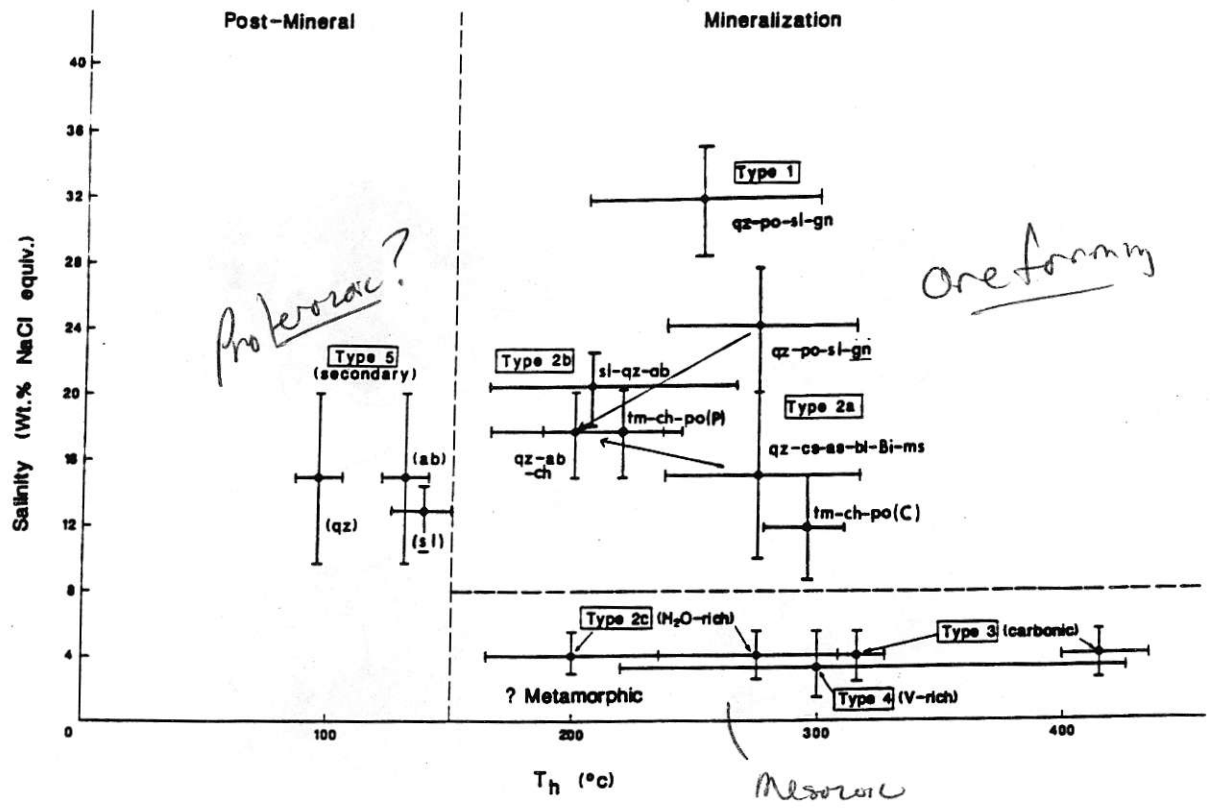


Figure 4. Salinity versus Th (homogenization temperature) plot for the fluids observed at Sullivan (after Leitch and Turner, 1991); Th values over 325°C are plotted but may be due to stretching. The division into metamorphic and mineralization related fluids is tentative, and is based on the difference in salinity between the two groups and the carbonic component of the latter. Habits of the inclusions and parageneses of associated minerals suggest a progression from Type 1 and 2a to Type 2b to Type 5; Types 2c, 3 and 4 are likely Mesozoic, based on relations seen elsewhere. Abbreviations are: ab=albite, as=arsenopyrite, Bi=native bismuth/antimony, bl=boulangerite, ch=chlorite, cs=cassiterite, gn=galena, ms=muscovite, po=pyrrhotite, qz=quartz, sl=sphalerite, tm=tourmaline, V=vapour, C=central, P=peripheral.

(1) bedded sulphides — brine pool — like Red Sea
replacement above vent — primary iron oxides, goethite, limonite
(2) alteration above fault — spessartine
(3) sills — mound in pool

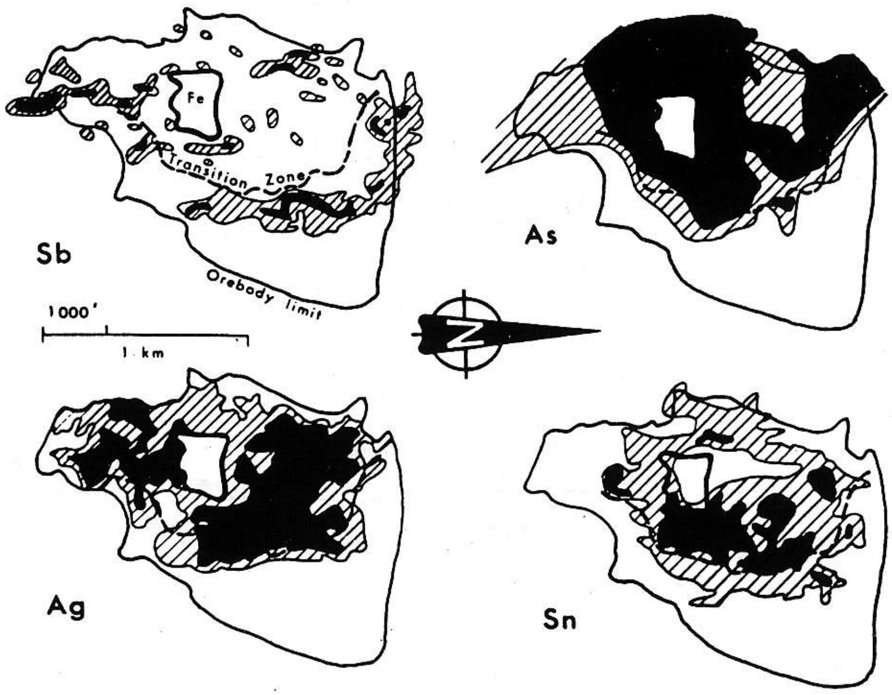


Figure 5. Distribution of minor metals in the Sullivan deposit (after Ransom *et al.*, 1985 and Freeze, 1966). White, hachured, and dark areas indicate increasing levels of each element; Fe=central pyrite-chlorite-calcite zone.

sulphide. Microprobe analyses show that there are two varieties of chlorite present in the deposit (Leitch, 1992b): (1) a widely distributed Mg-rich variety and (2) an Fe-rich variety present mainly peripherally, in the bedded ores. Preliminary analysis of the data suggests zoning of mineral compositions about the vent zone that give several vectors of potential use to exploration: Fe content of type (1) chlorites is highest in the center of the deposit, but for type (2), Fe increases toward the fringe of the deposit. There is also a zonation in carbonate, garnet and chlorite to higher Mn content toward the fringes of the Sullivan deposit. Plagioclase remaining in tourmalinized rocks is anorthite (up to An_{92}) compared to andesine (An_{38}) in regionally unaltered Aldridge strata, and albite (An_{0-2}) in albite-chlorite altered hangingwall or footwall rocks of the vent zone (Hamilton *et al.*, 1982; Leitch, 1992b). Regionally distributed clinozoisite becomes more Fe-rich (epidote) near the deposits. The Fe/Fe+Mg ratio of tourmaline, which is about 0.42 in fine felted "tourmalinite", varies from 0.9 in bluish schorl recrystallized by Moyie gabbro intrusions, to as low as 0.15 in clear or amber dravite recrystallized by albite-chlorite alteration. Traces of another borosilicate, axinite, have been discovered at the edge of the tourmalinite zone.

Calcic minerals such as plagioclase, tremolite and diopside are anomalous in the ore layers. The finely laminated nature of the major portion of the bedded ores, in which details of stratigraphy can be followed for up to 2 km, could reflect either brine pool precipitation or plume fallout in an open marine setting. However, a brine-pool model for the Sullivan deposit (Fig. 3) is supported by fluid inclusion data. The 15-27 wt % salinities in fluid inclusions are like those in anhydrite veins underlying the Red Sea brine pools, and similar to the salinity of the brine pools (13.5-25.6%: Ramboz *et al.*, 1988). Also, the presence of scapolite with composition near dipyre (Cl-rich, Na>Ca) (J. Hamilton, pers. comm., 1993), garnet coticule-bearing rocks, and primary Fe-oxide-hydroxide minerals such as goethtite and magnetite, all suggest a brine pool setting. Although evaporites are not known in the Belt Basin, they may have been present in shallow water facies underlying the Aldridge Formation, now exposed only along the eastern basin margin (Chandler and Zieg, 1994).

REFERENCES

Chandler, F.W. and Zieg, G.A., 1994, Was the depositional environment of the Sullivan Zn-Pb deposit in British Columbia marine or lacustrine and how saline was it?: a summary of the data; in Current Research, Part A, Geological Survey of Canada, Paper 94-1A, p. 123-130.

De Paoli, G.R. and Pattison, R.M.D., 1994, Metamorphic pressure, temperature, and fluid conditions of the Sullivan mine, Kimberley, British Columbia, using silicate- carbonate equilibria; Geological Survey of Canada, Minerals Colloquium, Program with Abstracts, Ottawa, Ontario, January 17-19, 1994.

Freeze, A.C., 1966, On the origin of the Sullivan orebody, Kimberley, B.C., in Tectonic History and Mineral Deposits of the Western Cordillera; Canadian Institute of Mining and Metallurgy, Special Volume 8, p. 263-294.

Hamilton, J.M., Bishop, D.T., Morris, H.C. and Owens, O.E., 1982, Geology of the Sullivan orebody, Canada, in Precambrian Sulphide Deposits (H.S. Robinson Memorial volume: R.W. Hutchinson, C.D. Spence, and J.M. Franklin, eds.); Geological Association of Canada, Special Paper 25, p.597-665.

Leitch, C.H.B. and Turner, R.J.W., 1991, The vent complex of the Sullivan stratiform sediment-hosted Zn-Pb deposit, British Columbia: preliminary petrographic and fluid inclusion studies; in Current Research, Part E, Geological Survey of Canada, Paper 91-1E, p. 33-44.

Leitch, C.H.B. and Turner, R.J.W., 1992, Preliminary field and petrographic studies of the sulphide-bearing network underlying the western orebody, Sullivan stratiform sediment-hosted Zn-Pb deposit, B.C.; in Current Research, Part E, Geological Survey of Canada, Paper 92-1E, p. 61-70.

Leitch, C.H.B., 1992a, A progress report of fluid inclusion studies of veins from the vent zone, Sullivan stratiform sediment-hosted Zn-Pb deposit, B.C.; in Current Research, Part E, Geological Survey of Canada, Paper 92-1E, p. 71-82.

Leitch, C.H.B., 1992b, Mineral chemistry of selected silicates, carbonates, and sulphides in the Sullivan and North Star stratiform Zn-Pb deposits, B.C., and in district-scale altered and unaltered equivalents; in Current Research, Part E, Geological Survey of Canada, Paper 92-1E, p. 83-93.

Leitch, C.H.B., Zieg, G.A., and Hall, S.M., (in review), Preliminary fluid inclusion and stable isotope studies of the Sheep Creek stratabound sediment- hosted Cu-Co deposit, Meager County, Montana. Submitted to Exploration and Mining Geology.

McMechan, M.E. and Price, R.A., 1982, Superimposed low-grade metamorphism in the Mount Fisher area, southeastern British Columbia--implications for the East Kootenay orogeny; Canadian Journal of Earth Science, v. 19, p.476-489.

Ramboz, C., Oudin, E., and Thisse, Y., 1988, Geyser- type discharge in Atlantis II deep, Red Sea: evidence of boiling from fluid inclusions in epigenetic anhydrite; Canadian Mineralogist, v. 26, p. 765-786.

Ransom, P.W., Delaney, G.D. and McMurdo, D., 1985, The Sullivan orebody, in Field guides to geology and mineral deposits in the southern Canadian Cordillera, D. Templeman-Kluit, ed., GSA Cordilleran Section Meeting, Vancouver, B.C., May 1985, p. 11-20 to 11-32.

Turner, R.J.W. and Leitch, C.H.B., 1992, Relationship of albitic and chloritic alteration to gabbro dykes and sills in the Sullivan mine and nearby area, B.C.; in Current Research, Geological Survey of Canada, Paper 92-1E, p.95-105.

Zieg, G.A. and Leitch, C.H.B., 1993, The geology of the Sheep Creek copper deposits, Meagher County, Montana; Program and Abstracts, Belt Symposium III, August 14-21, 1993 (Belt Association Inc., P.O. Box 1816, Spokane, WA 99210).

- silicification with fine diss. py in carbonate
- flourite ab.
- massive magnetite veins – correlation with Au?
- pyrite flooding in trachyte?
- Fluorite veining
- microsyenite's pervasively carbonatized and pyritized with several hundred ppb Au
- Bonanza veins in syenite dykes + tellurium
- locally advanced argillic alteration of syenite

High Sulphidation   Low Sulphidation   Alkalic

alkalic – two types

Continental Setting
rifting envir.
Cr. Ck 26m oz
South Dakota
Zortman-Landusky
↓
Flathead
Franklin Camp?

Oceanic
complex setting
rifting or subduction zone
↓
Porgera
Emperor
Lihir, New Guinea
↓
Quesnel trough
similar to Lihir

Porgera
Cripple Ck
Flathead

- meteoric fluids
- porphyry link
- pyritization, carbonatization
argillic
silicification
- fluorite, barite, roscoelite (V-mica)
py, gal, sph, moly, tellurides, magnetite

Pb-Zn-Ba MINERALIZATION AT WILDS CREEK: RELEVANCE TO STRATABOUND DEPOSITS ALONG THE WESTERN PURCELL ANTICLINORIUM

DEREK A. BROWN - B.C. Geological Survey Branch, 1810 Blanshard St., Victoria, B.C. V8V 1X4
PETER KLEWCHUK - Consolidated Ramrod Gold Corporation, 104-135 10th Ave. South, Cranbrook, B.C. V1C 2N1

INTRODUCTION

The Wilds Creek deposit (Leg or Legion property) is one of a series of stratabound Pb-Zn-Ba prospects and mines in upper Belt-Purcell stratigraphy along the western edge of the Purcell Anticlinorium (Fig. 1). These deposits are typically hosted in dolomitic units of the Dutch Creek Formation near the basal contact of the Mount Nelson Formation, although the stratigraphic nomenclature and correlation varies with different workers and requires a thorough review. The occurrences all lie in the hangingwall (west) of the Hall Lake fault. They have received various amounts of exploration attention and there is potential for large deposits, as shown by the Mineral King Mine located about 100 km north of Wilds Creek. It produced 1.334 million tons of 4.12% Zn, 1.76% Pb and 24.8 g/t Ag between 1956 and 1964 (Ministry of Mines, Annual Reports, 1956-1964). Mapping and drill core examination of the Wilds Creek property, about 12 km north-northwest of Creston, suggest characteristics that are of exploration significance for this part of the Belt Basin. The study is a component of a regional mapping program by the B.C. Geological Survey (Brown and Stinson, in prep.).

Galena and sphalerite were originally discovered in Wilds Creek in 1924. The showing was drilled by Newmont Mining Corporation (1954 - 6 holes), Sheep Creek Gold Mines (1961 - 2 holes), Aspen Grove Copper Mines (1964 - 5 holes), Legion Resources Ltd. (1989 - 7 holes), Kokanee Exploration Ltd. (1990 - 5 holes; 1992 - 9 holes) and Ramrod Exploration (1992 - 9 holes). There has been over 4,667 metres of drilling, excluding the 1964 program. Estimated reserves were reported at about 150,000 tons grading 6% Zn across 1 to 6 metres (Aho, 1964).

STRATIGRAPHY

The property stratigraphy comprises nine, poorly exposed units (from east to west; Fig. 2): (1) sericitic phyllite (Creston Formation); (2) thin bedded black argillite and tan dolomitic siltstone (Kitchener Formation); (3) chlorite-sericite phyllite with prominent but minor light grey quartzite beds; (4) argillite and dolomitic siltstone; (5) mineralized phyllitic dolomite and dolomitic siltstone; (6) dark grey phyllitic argillite and siltstone; (7) dolomite and limestone hosting the Pb-Zn-Ba mineralization (Dutch Creek ? Formation); (8) mafic volcanic rocks (Nicol Creek ? Formation); and (9) quartzite (Mt. Nelson ? Formation). Most units are phyllitic and internally tightly folded although unit contacts are not recognizably folded. Correlation of these units with the regional stratigraphic column remains a problem and should be considered tentative. Below are summary descriptions of the main units bounding the ore, the deposit structure and characteristics of the mineralization.

Two sequences of layered dolomite and impure quartzose dolomite (units 5 and 7) host banded

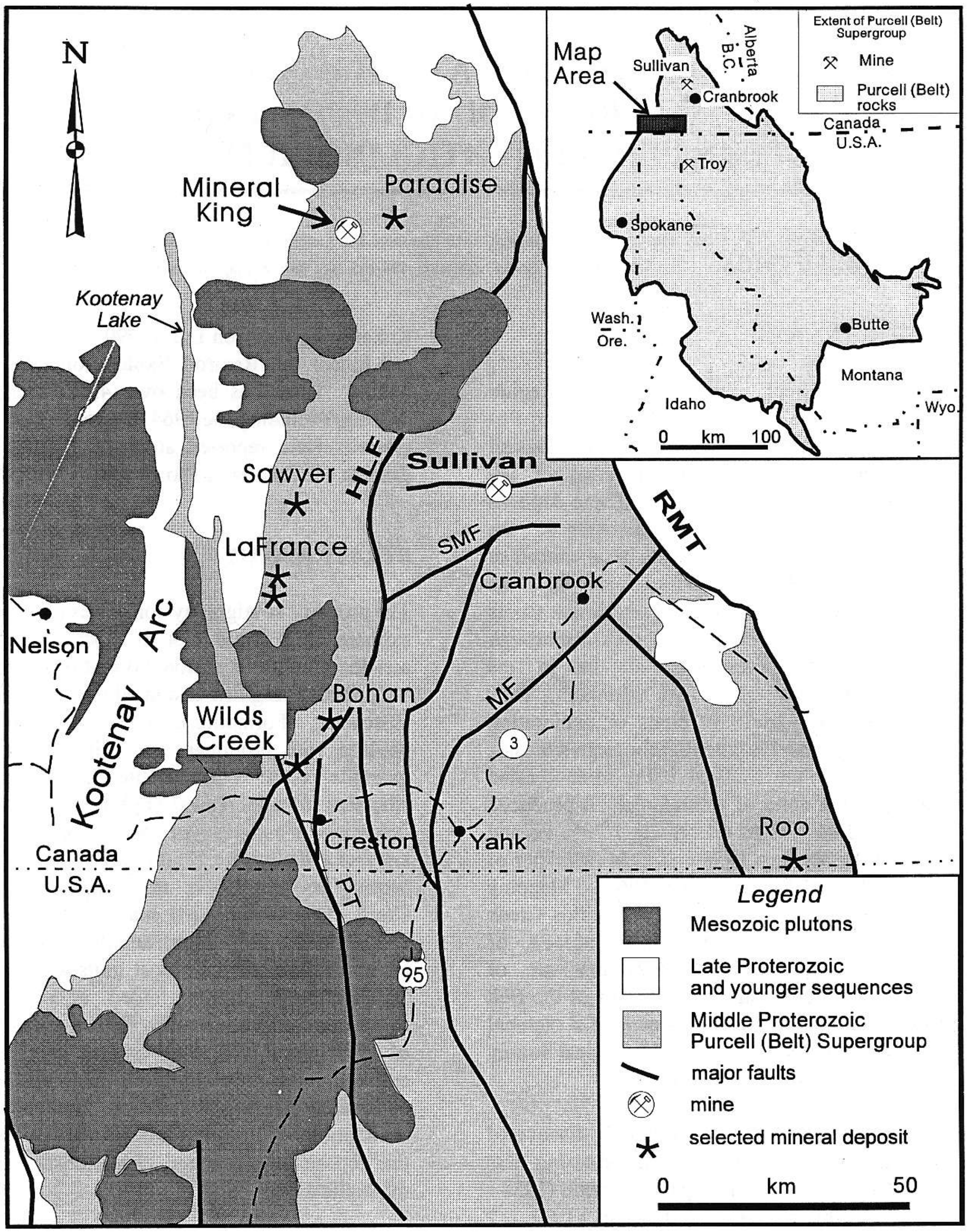


Figure 1. General location of stratabound Pb-Zn-Ba deposits and occurrences along the western flank of the Purcell Anticlinorium. Inset map illustrates the position relative to the Belt/Purcell Basin. HLF = Hall Lake fault, MF = Moyie fault, PT = Purcell Trench fault, RMT = Rocky Mountain Trench, SMF = St. Mary fault.

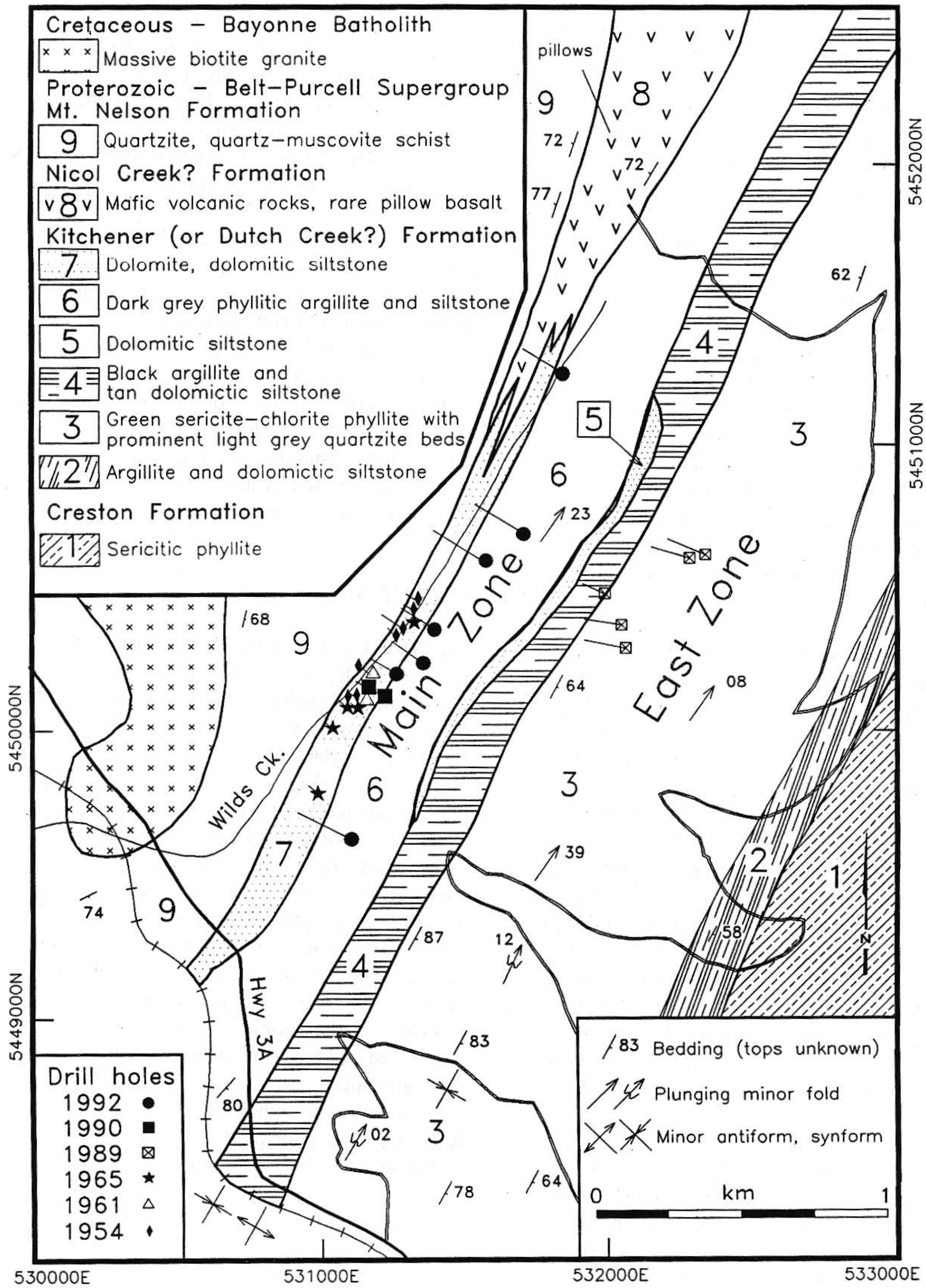


Figure 2. Generalized geology of the Wilds Creek (Leg) Property.

sulphide and are of prime economic interest. The eastern dolomite (unit 4) is more silicic and thinner than the recessive and poorly exposed white to cream coloured dolomitic siltstone of unit 7 (at least 70 metres thick). Distinctive dolomite and limestone breccia within the western dolomite horizon (unit 7) forms an important marker unit found in most of the other properties listed below. It contains angular argillite fragments and subrounded quartz and dolomite clasts up to 10 cm long. The matrix-supported argillite fragments weather out and contribute to the permeable character of this unit that now acts as an aquifer. Quartz fragments appear to be pieces of broken white quartz veins, perhaps derived from early diagenetic veins. The breccia is thought to be a solution collapse (karst) deposit and may correlate with breccias at the Mineral King Mine, described by Pope (1989), at Mt. Bohan (D. Anderson, pers. com., 1994) and at LaFrance Creek (Dave Wiklund, pers. com., 1994).

A volcanic succession (unit 8) including pillow lavas and associated tuff are exposed along a new logging road north of the mineralized dolomite. The pillows are up to 1 m long with chloritic selvages and locally with plagioclase porphyritic cores and amygdales. The brown weathering, medium to fine-grained flows are dark green on fresh surfaces, and locally have oxidized flow tops. Tuffs are deeply weathered, olive green to brown and friable. The 750 metre thick section is dominated by recessive tuff and capped by about 75 metres of flows. The volcanic unit pinches out to the south as 1 to 10 metre thick sills and rare flows within dolomitic and siliciclastic rocks . The recent recognition of these volcanic rocks is critical for regional correlation as a western equivalent of the Nicol Creek Formation.

An unmineralized, resistant, massive to thin bedded quartzite and quartz-muscovite phyllite (unit 9) forms the western edge of the property. Bedding dips fairly consistently and steeply to the southeast. The fine to medium-grained well foliated quartzite ranges from white to grey to pale green. Some layers are more chloritic and probably represent argillaceous interbeds. The quartzitic succession probably correlates with the base of the Mount Nelson Formation (Reesor, 1983) and has been interpreted to unconformably overlie the Dutch Creek Formation in the Toby-Horsethief creek area to the north (Pope, 1989).

A granitic stock, 500 m wide and 1500 m long, lies immediately west of the lower reaches of Wilds Creek. The massive biotite granite is related to the Cretaceous Bayonne Batholith that crops out farther to the northwest. Calc-silicate assemblages including coarse tremolite and biotite hornfels are prominent in the southwestern part of the property. Regional aeromagnetic data show that anomalously high values extend southward from the Bayonne Batholith under the entire property.

STRUCTURE

The Wilds Creek area lies on the northwest limb of the Goat River anticlinorium (Brown *et al.*, 1994) with a homoclinal succession from Creston Formation to Mount Nelson Formation. However, the details are more complicated and are currently being evaluated. On the property bedding and penetrative phyllitic chlorite-sericite foliations strike north-northwest and dip steeply to the east (also locally to the west). The southeast dips and apparent northwest facing direction suggest the stratigraphic succession is slightly overturned. Much of the structural style is controlled by competency contrasts of the different lithologies, tight chevron folds are abundant in the sericitic phyllite (Unit 3) and transposed bedding is common in argillite units. The mineralized carbonate (Unit 7) is phyllitic but the enclosed carbonate breccia displays no tectonic fabric, perhaps due to local flow of the carbonate unit along its contacts.

MINERALIZATION

Two separate carbonate horizons host different styles of mineralization. The "Main Zone" comprises two zones of stratiform, fine-grained, honey-coloured sphalerite and minor galena hosted in pyritic silicic dolomite (Unit 7). Baritic dolomite is important in this horizon, 1.3 metres of bedded barite is reported farther north, near LaFrance Creek (Dave Wiklund, pers. com., 1994). The "East Zone" consists of dolomite-hosted fracture controlled chalcopyrite-sphalerite-pyrite within unit 5. Recent drilling has concentrated on the Main Zone (Fig. 2). The Main Zone is at least 300 metres long and 2 to 3 metres wide, extending northward from Highway 3A to about 1250 metres elevation and lies within the western dolomitic horizon (Unit 7). Bedding parallel medium to fine grained pale sphalerite (up to 10%) and pyrite with rare galena occur within laminated dolomitic limestone and calcareous quartzite and argillite. This layering may be a primary texture. On surface in Wilds Creek the zone is intensely oxidized. Mineralization is banded in the south and becomes more silicified and massive to the north (Giroux, 1990). The East Zone is more intensely silicified than the Main Zone with abundant quartz veinlets and stockwork hosted within an eastern dolomitic horizon (Unit 5). It comprises pyrite with sporadic tetrahedrite, galena, sphalerite and chalcopyrite.

PRELIMINARY MODEL

The deposit contains stratabound Zn-Pb-Ba hosted within dolomite adjacent to mafic flows and tuffs that thicken rapidly to the north. The rapid change in the thickness of the mafic volcanic rocks is interpreted to be controlled by syn-volcanic growth faults developed during rifting. Synvolcanic block faulting synchronous with dolomite deposition would provide a heat source and conduits for the development of a hydrothermal system that could have produced the Pb-Zn-Ba mineralization at Wilds Creek. A rift-setting during extrusion of the Nicol Creek Formation farther to the east has been documented by Hoy (1993). The Roo Cu-Co-Ag prospect near Roosville (NTS 82G), a stratiform redbed copper showing above the Nicol Creek lavas may have formed in a similar setting.

CONCLUSIONS

Potentially similar prospects along the western edge of the Purcell anticlinorium include Mt. Bohan (Hall Property), LaFrance Creek (Wall and Dave claims), Mineral King and Paradise Mine (Fig. 1). The latter two are interpreted by Pope (1989) to be manto deposits. Exploration activity continues at many of these properties.

The exploration model for the Wilds Creek deposit can be summarized as follows.

- Stratigraphic interval: Dutch Creek - Mount Nelson contaet
- Local setting: association with silty dolomite, baritic dolomite and bedded barite (near LaFrance Creek) that contains a carbonate breccia unit (possible karst?); abrupt thickening of mafic volcanic rocks (Nicol Creek Formation?).
- Mineralization: stratabonnd sphalerite, galena; dolomite and barite gangue.
- Geochemical expression: strong Pb-Zn-Ba soil anomalies.
- Geophysical expression: spatial association with magnetic mafic volcanic rocks (flows and sills).

REFERENCES

Aho, A.E. (1964): Report on Liz-B Zinc Property, Creston, B.C.; consultant report, 10 p.

Aho, A.E. (1966): Report on the Creston Zinc Property; report submitted to Aspen Grove Mines Ltd., 5 p.

Anderson, D. (1989): 1989 Hall Property Year-end Report; company report; Cominco Ltd., February, 1989, 11 pages.

Brown, D.A., and Stinson, P., (in prep.), Geology and Iron Range Deposit, and diatremes, in Yahk Map Area, southeastern

British Columbia (82F/1); *in* Geological Fieldwork 1994, B.C. Ministry of Energy, Mines, and Petroleum Resources.

Brown, D. A., Bradford, J. A., Melville, D. M., Legun, A. S. and Anderson, D. (1994): Preliminary Geology of Purcell Supergroup in the Yahk map area, southeastern British Columbia (82F/1); *in* B.C. Ministry of Energy, Mines and Petroleum Resources, Paper 1994-1, p. 129-151.

Callan, N.J. (1990): 1990 Year End Report, Wall-Dave Property (NTS 82F/7, 10); company report; Cominco Ltd., December 21, 1990, 12 p.

Giroux, G.H. (1990): A geochemical and geophysical report on the Liz B - John 2 Property; B.C. Ministry of Energy, Mines and Petroleum Resources, Assessment Report 19,902, 19 p.

Hoy, T. (1993): Höy, T. (1993): Geology of the Purcell Supergroup in the Fernie west-half map area, southeastern British Columbia; B.C. Ministry of Energy, Mines and Petroleum Resources, Bulletin 84, 157 p.

Pope, A.J. (1989): The tectonics and mineralisation of the Toby-Horsethief Creek area, Purcell Mountains, southeast British Columbia, Canada; Ph.D. thesis, RHB New College, London, 350 p.

Reesor, J.E. (1983): Geology of the Nelson Map-area, East Half; Geological Survey of Canada, Open File 929.

Slingsby, A.E., Gillan, J.F. and Smith, L.J. (1981): Year End Report, 1981 Exploration Program, Kootenay Project, British Columbia (NTS 82F/7, 10); Norcen Energy Resources Limited, 23 p.

Stephenson, L. (1991): Assessment Report on diamond drill hole L90-2, Leg Property; Kokanee Exploration Ltd., B.C. Ministry of Energy, Mines and Petroleum Resources, Assessment Report 21,354.

THE FORS PROSPECT, A PROTEROZOIC SEDIMENTARY EXHALATIVE BASE METAL DEPOSIT IN MIDDLE ALDRIDGE FORMATION, SOUTHEASTERN BRITISH COLUMBIA (82G/5W).

JAMES M. BRITTON - B.C. Geological Survey Branch, 1810 Blanshard Street, Victoria, B.C. V8V 1X4
DAVID L. PIGHIN - Consolidated Ramrod Gold Corporation, 104-135 10th Avenue South, Cranbrook, B.C. V1C 2N1.

INTRODUCTION

The 1992 discovery of high-grade base and precious metal mineralization at the Fors property, 17 km SW of Cranbrook, rekindled interest in the Middle Proterozoic Aldridge Formation (Pritchard equivalent) and provides a new exploration target. Ag-Pb-Zn mineralization occurs at the top of a discordant zone of pebble wacke or "fragmental" in middle Aldridge sandstone and mudstone. Pyrrhotite, sphalerite, galena, arsenopyrite, pyrite, chalcopyrite, and bismuthinite occur in stratiform, semi-massive to massive lenses, as well as disseminations and veins. Scheelite is a local accessory. Gold values range up to 0.7 grams per metric ton; silver to 734 grams per metric ton. Best drill intersections were up to 25% combined Pb and Zn over 1 m. No tonnage estimates are available. The deposit is unusual in having extensive and varied alteration assemblages dominated by plagioclase, biotite, tourmaline, white mica, carbonate, tremolite-actinolite, talc and silica.

EXPLORATION HISTORY

Early in 1966 Helg Fors of Kimberley discovered lead-zinc mineralized float on logging roads near Moyie Lake. Subsequent prospecting by Cominco Ltd. discovered the Main showing, a small lens of bedded lead-zinc sulfides. Follow-up work from 1966 to 1983 included at least 5 shallow and 2 deeper diamond drill holes totalling 944 m (Webber, 1978a,b). No mineralization of economic interest was encountered and eventually the claims were allowed to lapse. The area was restaked and optioned to Placer Dome who explored the property in 1989.

In 1992 Kokanee Explorations (later Consolidated Ramrod Gold Corporation) commenced a diamond drill program on behalf of Chapleau and Barkhor Resources. Their first drill hole (F92-1), collared at the Main showing, intersected thin zones of disseminated to massive sulfide mineralization over a stratigraphic interval 42 m thick (Klewchuk, 1993). The highest grade intercept was 1 metre of massive sulfides with 9.35% lead, 16.4% zinc, 0.09% cadmium and 98 grams per metric ton silver (The Northern Miner, December 7, 1992). To date 12 holes (2440 m) have tested the extent of the 1992 discovery.

GEOLOGY OF THE FORS AREA

The Fors area is underlain by gently to moderately north to northeast dipping strata of the lower and middle divisions of the Aldridge Formation that have been intruded by three mafic sills (Fig. 1). The deepest intrudes near the lower-middle Aldridge contact and is at least 250 metres thick. At its closest it is 350 metres below the top of the Fors deposit (Fig. 2). The two other sills are above it.

The northeast striking Moyie fault (a reverse fault of regional significance) defines the southern limit of prospective ground. Minor

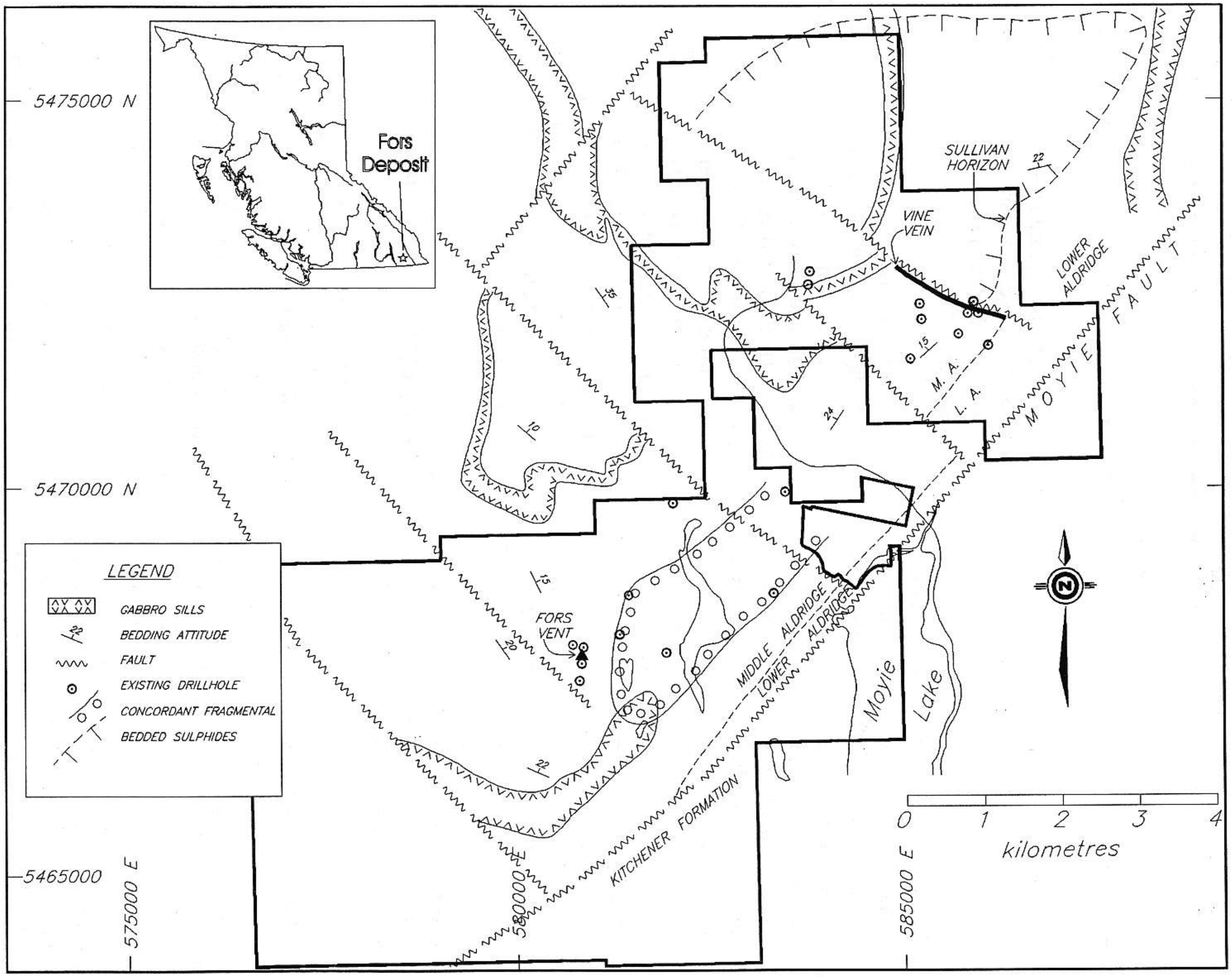


Figure 1. Geology of the Fors -Vine area.

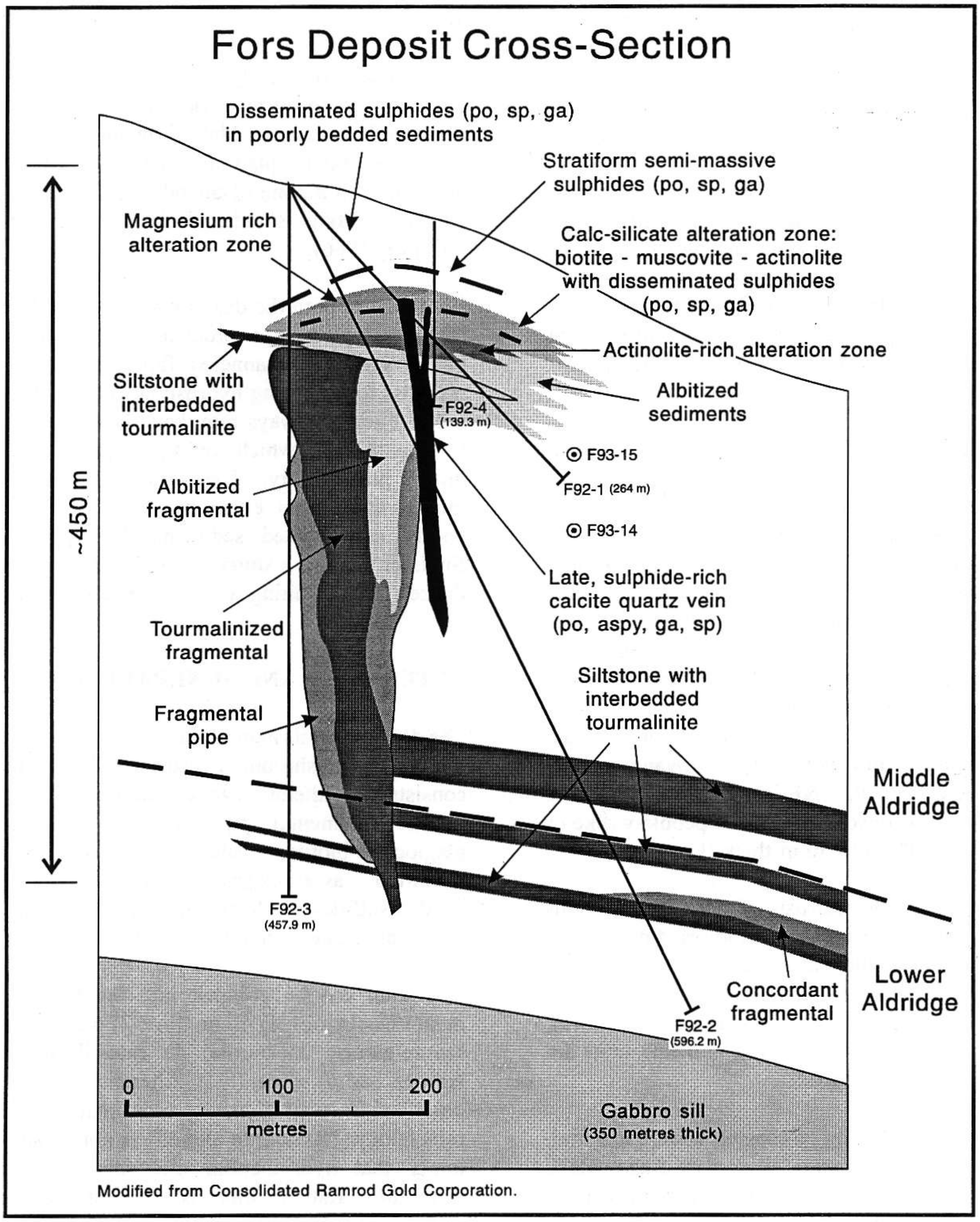


Figure 2. Schematic NE-SW cross section of the Fors deposit.

northwest and west striking high-angle faults break the stratigraphic sequence into a mosaic of structurally homogeneous blocks. Deformation is mostly limited to gentle open folds. Near faults bedding can be deflected into the plane of the fault. Shearing and tight folding occur only along the Moyie fault zone (Hoy and Diakow, 1982).

Metamorphic grade is at most upper greenschist facies (Hoy, 1993), and is attributed to simple burial. Despite metamorphic effects primary sedimentary structures are very well preserved. Only where there has been intense hydrothermal alteration or deformation are they obliterated.

STRATIGRAPHY

The oldest rocks on the property are siltstones, quartzites and silty argillites of the lower Aldridge Formation. At Fors they crop out in a thin wedge along the Moyie fault (Fig. 1) against which they have been folded. They are distinguished by rusty weathering, thin, planar bedding and coarse reddish biotite porphyroblasts that parallel bedding and grow at random angles to it. Near the top of this unit (at the stratigraphic equivalent of the Sullivan horizon), thickening to the NE away from the Fors deposit, is a concordant layer of pebble-wacke or "fragmental" up to 50 m thick (Fig. 2).

Middle Aldridge consists of thick, monotonous sequences of mostly AE turbidites: fine-grained quartzofeldspathic sandstones (mostly wackes, some arenites), siltstones and argillites with variable amounts of biotite, white mica, pyrrhotite and pyrite. Coarse and fine units are commonly interbedded.

A pipe-shaped body of coarsely clastic material, referred to herein as a "discordant fragmental", occurs at depth (Fig. 2). It consists of sand to pebble sized clasts of sandstone and siltstone in a silty to sandy matrix. Most clasts are subrounded to subangular and matrix supported. The unit is up to 100 metres in diameter and 300 meters high.

A sequence of nearly massive fine grained sediments 30 to 60 m thick was recognized by early workers near the Main showing. It consists of "intermixed quartzitic and argillaceous material with a zone of abundant pyrrhotite" in which "bedding is either lacking or obscure" (Gifford, 1966).

We interpret both the discordant fragmental and the massive unit as products of dewatering phenomena that channeled fluids upwards in response to increasing hydrostatic and lithostatic loads. Fluid pathways may have been localized by growth faults which could have provided the initial permeability. The clastic or massive fabrics result from either hydraulic milling of poorly consolidated sediments by upwelling fluids or venting a slurry of mud and sand onto the sea floor, forming a volcano-like edifice.

ALTERATION AND MINERALIZATION

The deposit and its associated alteration envelope are crudely mushroom shaped (Fig. 2). Its stem consists of an alteration zone within and adjacent to the fragmental pipe; its eap comprises plagioclase-biotite, calc-silicate and mica alteratlon assemblages with disseminated to bedded sulfides. Both stem and cap are cut by a thick, late-stage, sulphide-rich vein.

Alteration can be grouped into the following main associations: 1) tourmaline; 2) plagioclase-biotite-garnet; 3) biotite; 4) calc-silicate; 5) sericite; and 6) silica. (N.B. Alteration types are described in terms of their present metamorphic mineralogy. "Actinolite altered" is not meant to imply that hydrothermal alteration produced actinolite. Rather this is the product of metamorphism of precursor minerals that formed during alteration.)

TOURMALINE alteration consists of partial to complete replacement of original sedimentary material by microscopic grains of tourmaline. Two types of tourmaline alteration occur at the Fors. The first and probably oldest is bedded black tourmaline (Fe-rich; schorl) that prefentially affects argillaceous layers. A model for this form of tourmaline alteration is that ascending boron rich hydrothermal solutions pass through coarser sediments to become trapped by less permeable clay rich strata, reacting with them to form bedded tourmalinite (Slack, 1993). Bedded tourmalinite is locally associated with white Mn-rich garnets (<0.5 mm). The second type is light to dark brown (Mg-rich; dravite). It is mainly confined to the discordant fragmental in which both clasts and matrix are altered. Both types are locally associated with a little pyrrohtite, more rarely with arsenopyrite, sphalerite and galena.

PLAGIOCLASE-BIOTITE±GARNET alteration ("Albitization") affects large volumes of rock, including part of the fragmental pipe and bedded sediments surrounding it. It appears to be assymetrically distributed around the pipe, skewed to the NE. It is pervasive and texture destructive, and results in an aphanitic to very finely granular mottled grey, white and pink rock. It occurs as veins, patches, hairline fractures and broad diffuse areas. Its contacts are sharp to gradational. Garnets in this zone are pale pink and up to 2 mm in diameter. Pyrrhotite is common in these areas, preferentially replacing biotite.

BIOTITE AND CALC-SILICATE alteration is mainly confined to the "cap" of the deposit. It consists of complexly interlayed, coarse-grained assemblages of micas (brown biotite and muscovite), amphiboles (actinolite and tremolite) and carbonate minerals (clacite and dolomite). Because these minerals form 100% of the rock it is thought that this zone may represent a hydrothermal vent area. The cap or vent horizon is crudely stratified with a nearly massive zone of actinolite at its base, a biotite rich zone in the middle and a thin magnesium-rich zone (talc-tremolite-dolomite) at the top.

Actinolite also occurs away from the calc-silicate cap. It always appears to replace biotite.

SERICITE alteration occurs as a distal aureole around the other alteration assemblages at depth and above the bedded sulfide zone up to and including the Main showing. Locally it is texture destructive and pervasive but more commonly it is confined to bedding planes in porous, feldspathic units.

SILICA alteration is mainly confined to strata immediately overlying the bedded sulfide horizon that overlies the calc-silicate alteration cap. Core is blue grey, hard, with a diffusely granular appearance. This alteration also occurs as thin envelopes around late stage quartz veins.

MINERALIZATION

Zones of nearly massive sulfides are rare. They occur in two forms: stratiform and vein. Thickest and highest grade drill intersections were encountered in F92-1.

Conformable massive sulfides consist of fine to coarse-grained pyrrhotite, sphalerite, and galena mineralization within a plagioclase-biotite-sericite envelope. The sulfides locally contain coarse (to 8 mm) clasts of transparent quartz and have a cataclastic fabric. Upper and lower contacts approximate bedding in the enclosing sediments. A maximum thickness of 2 m was intersected. The unit lies a few meters above the top of the calc-silicate alteration cap.

A semi-massive sulphide vein almost 2 m thick, with a calcite-quartz gangue, cuts an actinolite-rich alteration zone. The vein consists mainly of granular pyrrhotite rimmed by arsenopyrite with variable amounts of sphalerite and galena and accessory scheelite, chalcopyrite and bismuthinite. One assay returned 734 grams per metric ton Ag, 16.7% Pb, and 5.40% Zn over 0.3 m.

Low grade zones of sulfides are quite widespread. They consist of disseminations, stringers, veins, small semi-massive to massive stratiform lenses and irregular patches of mainly pyrrhotite, with subordinate amounts of sphalerite, galena, pyrite, and rare arsenopyrite and chalcopyrite and bismuthinite.

DEPOSIT MODEL

A simplified genetic model for the Fors deposit is as follows:

1) pelagic, turbidite sedimentation with entrained organics and iron in a fault controlled graben or half graben results in a thick sequence of poorly consolidated sediments.

2) development of a fragmental pipe which acts as a long-lived conduit for upward migration of fluids at least until the time of formation of the bedded sulfides above the calc-silicate zone and probably until after the formation of the Main showing at surface. The upper limit of coarse fragmentals appears to lie just below the calc-silicate cap which may represent an exhalative vent deposit. The pipe may have formed in the hangingwall of a growth fault. Parallel and conjugate fractures related to this fault may have provided subsequent pathways for fluid migration.

3) tourmalinizing fluids ascend, preferentially travel along beds, but also change chemistry with time: most bedded tourmalinites are schorlitic (Fe rich); however much of the fragmental pipe is brown to pale brown, dravitic (Mg rich).

4) "albitizing" fluids ascend reacting with bedded tourmaline alteration locally but mainly spread laterally away from the pipe to the northeast.

5) potassium, iron and magnesium rich fluids deposit biotite (pale brown to bronze brown) in the main part of the vent horizon.

6) late carbonate rich fluids ("flooding") along parasitic or antithetic structures overprint biotite to produce actinolite assemblages and deposits of sulfides. Two pulses are possible: the first produced semi-massive to massive, locally stratiform Zn-Pb-Ag rich mineralization with a high base metal to iron ratio. The second pulse produced veins enriched in As, W, Ag and Bi, as well as Zn, Pb and Fe.

7) upwelling carbonate rich fluids mix with downward circulating seawater to produce Mg enriched assemblages including talc, tremolite, dolomite at the top of the alteration cap, and possibly the Mg rich tourmalinites found in the upper part of the fragmental pipe.

8) later fluids yield sericite (-silica) alteration with minor sulfides that enrich the overlying sedimentary package and formed the Main showing.

9) heat and fluids for hydrothermal alteration and mineralization may have been provided by the intrusion of a thick mafic sill into wet sediments.

10) regional metamorphism creates present silicate mineralogy and redistributes some sulfides.

CONCLUSION

The Fors prospect is a well preserved example of a small, high-grade Pb-Zn-Ag sedimentary exhalative and vein deposit hosted in Middle Proterozoic Aldridge Formation. It is associated with an unusually strong alteration assemblage variously dominated by plagioclase, biotite, tourmaline, white mica, carbonate, tremolite-actinolite, talc and silica. It is a blind discovery that resulted from drill-testing a geological model of low-grade mineralization found at surface.

It provides a new exploration target in the Sullivan camp, having some similarities with the Sullivan deposit and some important differences. Similarities include the presence of such "Sullivan indicators" as bedded sulfides, fragmental units that locally carry sulfide-bearing and tourmalinized clasts, garnet porphyroblasts, and tourmaline and albite alteration. Differences are that it is located outside the Sullivan corridor, is stratigraphically higher, has unusual alteration assemblages, and has elevated silver, gold, tungsten and arsenic.

The deposit deserves further study and would make an excellent project for a graduate student interested in water-rock interactions. Very good core storage, logging and rock cutting facilities exist at the field office of Consolidated Ramrod Gold Corporation to support such work.

REFERENCES

Gifford, R.G. (1966): Report of Geological, Geochemical and Geophysical Surveys on the Helg Groups of Claims; B.C. Ministry of Energy, Mines and Petroleum Resources, Assessment Report 834, 6 pages.

Hoy, T. (1993): Geology of the Purcell Supergroup in the Fernie West-half Map Area, Southeastern British Columbia; B.C. Ministry of Energy, Mines and Petroleum Resources, Bulletin 84, 157 pages.

Hoy, T. and Diakow L. (1982): Geology of the Moyie Lake Area; B.C. Ministry of Energy, Mines and Petroleum Resources, Preliminary Map No. 49, scale 1:50,000.

Klewchuk, P. (1993): Assessment Report on Two Diamond Drill Holes (F92-2 & 3), Fors Property; B.C. Ministry of Energy, Mines and Petroleum Resources, Assessment Report 22817, 11 pages.

Slack, J.F. (1993): Models for Tourmalinite Formation in the Middle Proterozoic Belt and Purcell Supergroups (Rocky Mountains) and their Exploration Significance; in Current Research, Part E, Geological Survey of Canada, Paper 93-1E, pages 33-40.

Webber, G.L. (1978a): Diamond Drilling Report, Vine No. 29 Claim, N.T.S. 82G/5, Fort Steele Mining Division; B.C. Ministry of Energy, Mines and Petroleum Resources, Assessment Report 6936, 5 pages.

Webber, G.L. (1978b): Geological Report, Vine Property 1978, N.T.S. 82G/5, Fort Steele Mining Division; B.C. Ministry of Energy, Mines and Petroleum Resources, Assessment Report 7087, 9 pages.

Backgd

	Pd ppb	Pt ppb	Au ppb	Cu ppm	Ni ppm
Backgd	28	22	10	760	29
High Grade	2266	1021	713	23885	238
Concentration	80	45	65	31	8

- have high pyroxene contents
- ? impact on sills of ~~dero~~ reacting with sediments in Canada (OVC)
- mineralized rock is variably altered, can appear to be fresh
- magmatic mineral phases in ore
- argillic alteration ??
- hydrothermal platinum occurrences have pt-tellerurides which are similar to same at Flathead Indian Reserve

Exploration

- system may be larger than expected
- look for dykes with pyroxene, magnetite and anomalous Cu and V in soils
- steeply cutting dykes in Prichard/Aldridge that flutter out higher in stratigraphy
- IP, VLF, magnetic surveys

COPPER AND PLATINUM GROUP MINERALIZATION IN A MAFIC DIKE-SILL BODY OF THE REVAIS CREEK MINING DISTRICT, FLATHEAD INDIAN RESERVATION, MONTANA

D.D. LAUER - 7580 Scholls Ferry Road, Beaverton, OR 97008

ABSTRACT

Palladium, platinum, copper, and gold exist within a mafic intrusive dike-sill body in the Revais Creek Mining district on the Flathead Indian Reservation, Montana. The dike portion of the intrusive body cuts sediments and diabase sills in the Prichard Formation of the Belt Supergroup and, as it enters the Burke and Revett Formations, it becomes more sill-like. Platinum and palladium can be detected throughout the body with a combined average of about 50 ppb and are concentrated in one sample as high as 12,483 ppb. Copper occurs throughout averaging about 400 ppm and is elevated in the same areas as the platinum group elements (PGE) to more than 11%. Gold also concentrates in these areas to as much as 2,250 ppb while having a background average of about 7 ppb.

High levels of PGE, copper, and gold occur in at least five areas within the body, apparently concentrated by two different processes. My data suggest that one metal concentrating process is magmatic sulfide segregation and separation from the silicate melt, while the other is hydrothermal remobilization of these metals, perhaps from a sulfide zone concentrated by the first process. The sulfide segregation PGE-deposit type typically has low Pt:Pd ratios (as low as 1:4) and up to 6% primary copper sulfides. Hydrothermally altered areas containing high metal concentrations are characterized by high Pt:Pd ratios (as much as 3:1), and secondary copper sulfides and oxides. A possible explanation for the change in the Pt:Pd ratio is mobilization and removal of palladium by hydrothermal and supergene processes. Microscopic textures, preliminary sulfur isotope analyses, and the high background levels of copper all suggest that the copper sulfides are of a purely magmatic origin.

This occurrence differs from the traditional PGE association with chromite and nickel-copper sulfides in mafic and ultramafic layered intrusions. Here, the PGE lies in close association with copper sulfides within a mafic dike-sill body. Using a JEOL-6100 scanning electron microscope, several PGE mineral grains were observed and identified in six different samples containing high PGE grades. Of these six samples, however, only one has been strongly hydrothermally altered and only one PGE mineral grain was observed. Large palladium minerals such as michnerite, atokite, mertieite, stibiopalladinite, and kotulskite dominate the PGE minerals observed. A few other noted PGE occurrences such as the Salt Chuck Intrusion in Alaska, the Coldwell Complex in Ontario and the Sudbury Footwall deposits in Ontario, show similar copper sulfide and PGE mineralogy. All of these, however, occur in large intrusive bodies, but the Revais Creek occurrence appears to be unique in that it occurs in a relatively small gabbroic dike-sill body.

tholeiite

ORIGIN AND GEOLOGIC CONTROLS ON A Ag-Pb-Zn DEPOSIT IN THE WALLACE FORMATION: SEDEX or Cd'A VEIN ORIGIN FOR THE GOLD HUNTER DEPOSIT, COEUR d'ALENE DISTRICT, IDAHO

JED THOMAS - Sage Resources, 645 Plymouth Street, Missoula, MT 59801

ABSTRACT

The Gold Hunter Ag-Pb-Zn deposit occurs as peneconformable veins in the lower member of the Wallace formation. Veins fill brittle fractures and replace nearby wallrock, forming overlapping and stacked ore shoots. Mineral trends follow the orientation of the metamorphic lineation.

Isotopic data, lithologies, structural controls, petrology and alteration patterns were used to define a geologic model of the mineralization. The results indicate a shear-zone control similar to other Coeur d'Alene deposits.

RELATION OF GOLD PLACERS IN MONTANA TO BEDROCK GEOLOGY – IMPLICATIONS FOR LODE EXPLORATION

LEE A. WOODWARD -Dept. of Earth and Planetary Sciences, University of New Mexico, Albuquerque, NM 87131-1116

ABSTRACT

At least 318 gold placers have been mined in Montana and produced about 8,875,000 ounces of gold between 1862 and 1965. Inferred primary sources of the placer gold are: (1) Archean exhalative iron-formations containing gold, (2) syngenetic and(or) diagenetic deposits in Middle Proterozoic Belt Supergroup strata, with remobilization of gold from some of the deposits during Late Proterozoic and(or) Late Cretaceous to early Tertiary magmatism, (3) Middle Cambrian carbonate strata with gold trapped in algal mats, (4) lodes hosted by Tertiary-Cretaceous intrusions and(or) their wallrocks, and (5) disseminated or stockwork gold in Tertiary and Cretaceous volcanic rocks. Secondary sources, including Tertiary placers and Pleistocene glacial deposits, have been reworked by fluvial processes.

Major tectonic features appear to have been important factors in emplacement of some gold lodes and hence have influenced the distribution of placers eroded from the lodes. For example, syndepositional faults of Middle Proterozoic age may be the principal control in localization of syngenetic lode-gold deposits in Proterozoic Belt rocks along the Lewis and Clark line and elsewhere in western Montana. In addition, some of the precious-metal deposits in the southwestern part of the state occur near basal zones of the Medicine Lodge and Grasshopper thrust plates where older, steep faults or fracture zones might have channeled intrusions and associated mineralizing solutions into favorable host rocks, mainly Paleozoic carbonate strata. The Great Falls tectonic zone appears to have controlled emplacement of several major plutons with associated gold mineralization in southwestern Montana.

Many gold placers in Montana have no known sources, suggesting the possible presence of undiscovered lode-gold deposits. Numerous potential lode targets can be defined using regional geology and the inferred bedrock sources of gold along with gold placers as physical signatures. The Prichard Formation of the Belt Supergroup is a particularly attractive target inasmuch as numerous gold placers with unknown sources occur in drainage basins underlain by the Prichard.

INTRODUCTION

At least 318 gold placers have been mined in Montana (J.S. Loen, written communication, 1994), and production is estimated at about 8,875,000 ounces from 1862 through 1965 (Koschmann and Bergendahl, 1968). This accounts for half of Montana gold production (approximately 17,750,000 ounces) from both lodes and placers during 1862-1965 (Koschmann and Bergendahl, 1968, p. 143). Most of the placer gold was mined between 1862 and 1870. Thomson (1948, in Lyden, 1987) estimated that probably ten percent of the placer mining districts accounted for about half of the placer gold. Data from Frishman *et al.* (1990) indicate that 26 placers probably produced 80 percent of the placer gold. For example, the Virginia City-Alder Gulch district produced an estimated

Northwest Geology, v. 23, p. 45-60, 1994

2,475,000 ounces from placers derived from epigenetic, auriferous pyrite-quartz veins hosted principally by Archean metamorphic rocks (Shawe and Wier, 1989).

Information about the gold placers was taken mostly from a comprehensive report by Lyden (reprinted in 1987 from a 1948 publication now out of print) and U.S. Geological Survey reports (Chesson *et al.*, 1984; Elliott *et al.*, 1992; Frishman *et al.*, 1990; Loen and Pearson, 1989; Pearson *et al.*, 1991). The placers are classified here according to the following inferred sources of the gold: sources unknown; source(s) not fully known; reworked from older placer, with lode source unknown; lode source not mined or only partly mined; and lode source mined. Many of the more productive placers were derived from known lode sources that have been mined at least in part. Emphasis in this report, however, is on placers with poorly established sources (Fig. 1) because they may in some cases indicate areas containing undiscovered lodes. Table 1 shows the numbered localities on Fig. 1 and discussed in the text.

FACTORS IN PLACER FORMATION

Loen (1992) pointed out that some of the problems concerning the problems of source areas for gold placers can be resolved by using mass balance constraints. Factors that determine the amount of placer gold concentrated include denudation rate, drainage basin area, time span, abundance of gold in source rocks, and efficiency of weathering and concentration processes. Depending on these variable factors and the amount of material eroded, economic lode deposits may or may not occur in the source areas. Gold placers form only where geomorphic processes are favorable. In Montana, these processes were favorable mainly on Pliocene pediments, Quaternary terraces, and stream channels in drainages that were not glaciated; in addition, there appears to be much reconcentration of gold from high-level Tertiary deposits into Holocene placers (J.S. Loen, written communication, 1994).

PRINCIPAL SOURCES OF PLACER GOLD

The following discussion of the inferred lode sources of the placer gold is interpretive and strongly influenced by my biases concerning the timing and styles of primary gold mineralization. Despite the problems and limitations of such an approach, it has merit inasmuch as it suggests exploration targets for lode-gold deposits. A classification of sources for the gold placers includes the following primary deposit types: (1) Archean exhalative iron-formations containing gold-quartz mineralization, (2) syngenetic and(or) diagenetic Proterozoic deposits, some of which may have been remobilized during Late Proterozoic and(or) Late Cretaceous to early Tertiary (Laramide) magmatism, (3) Cambrian carbonate strata with gold trapped in algal mats, (4) lodes hosted by Tertiary-Cretaceous intrusions and(or) their wallrocks, and (5) disseminated or stockwork gold in Tertiary and Cretaceous volcanic rocks. In addition, some placers are derived from secondary sources such as older placers (commonly Tertiary) and Pleistocene glacial deposits.

Tertiary and Cretaceous Intrusions

Large, granitoid intrusions of Late Cretaceous to early Tertiary age are abundant in southwestern Montana, and smaller, more scattered plutons extend northeasterly into the central-northern part of the state (Fig. 2). In western Montana these rocks are calc-alkaline and to the east they form the central Montana alkalic igneous province (Pirsson, 1905; Larsen, 1940; Chadwick, 1965). A few small, scattered calc-alkaline stocks are present in northwestern Montana.

Five major intrusive centers in southwestern Montana are the Boulder, Pioneer, Idaho, Philipsburg, and Tobacco Root batholiths (Klepper *et al.*, 1974; Hyndman, 1983; Arth *et*

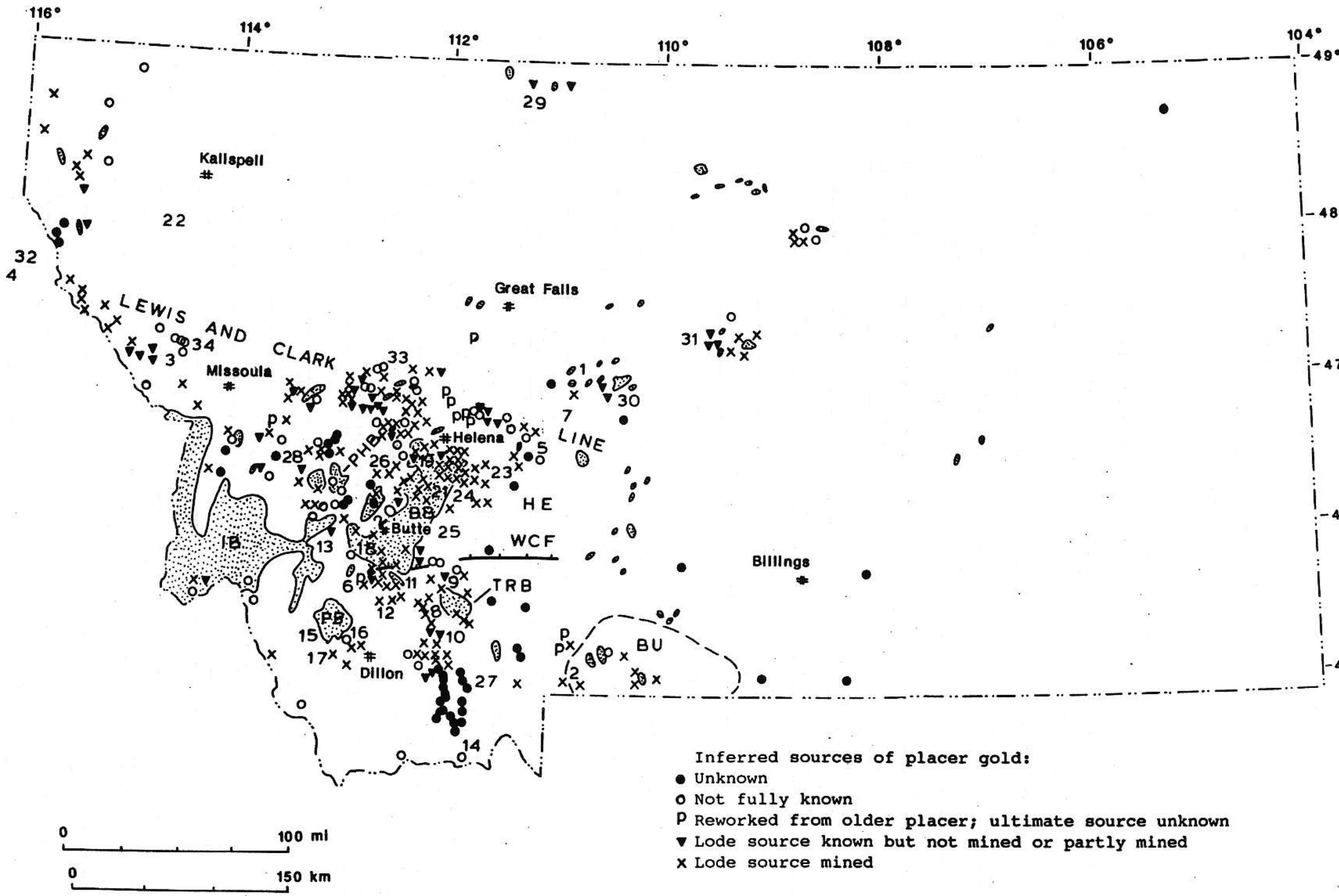


Figure 1. Locations of gold placers in Montana. Tertiary and Cretaceous plutons are stippled. HE = Helena embayment of Belt basin, WCF = Willow Creek fault, IB = Idaho batholith, BB = Boulder batholith, PhB = Philipsburg batholith, PB = Pioneer batholith, and TRB = Tobacco Root batholith, BU = Beartooth uplift, shown with dashed line. Numbered localities are referred to in text. The reader is referred to Frishman *et al.* (1990) for data concerning all the placers known in Montana.

Table 1. Numbered localities noted in text and Figs. 1 and 3, with lode and placer gold production noted where appropriate. Production data from Koschmann and Bergendahl (1968), Frishman *et al.* (1990), Shenon (1938), and Lyden (1987). N.A. = not applicable.

Map Number	Name	Approximate Gold Production in Troy Ounces Lode	Placer
1	Neihart district	67,000 total	
2	Jardine-Crevasse district	200,000	407 +
3	Alberton	N.A.	
4	Couer d'Alene region	505,000 total	
5	Big Belt Mountains	83,000	920,000
6	Fish Creek-Moose Creek area	500 to 5,000	500 to 5,000
7	Sheep Creek area	none	
8	Tobacco Root Mountains	N.A.	
9	Renova district	162,000	unknown
10	Virginia City-Alder Gulch	142,000	2,475,000
11	Silver Star district } combined	185,700	unknown
12	Rochester district		
13	French Creek area		50,000 to 250,000
14	Odell Creek area		unknown, but small
15	Dyce Creek area	unknown, but small	
16	Argenta district	65,350	unknown, but small
17	Bannack district	108,400	132,000
18	Butte district	2,362,000	363,000
19	Rimini district	194,000	4,275
20	Lowland Creek area	unknown	8,500 +
21	Montana Tunnels mine	75,000 +	
22	Hoy Heaven district	3,000	
23	Indian Creek district	37,800	42,500
24	Radersburg district	276,000	49,000
25	Elkhorn Creek area	70,000	unknown
26	Emery district	39,450	3,625
27	Ruby Creek mine	unknown	
28	John Long Mountains	37,000	75,000
29	Sweetgrass Hills	4	2,000
30	Little Belt Mountains	68,000	unknown
31	North Moccasin Mountains	650,000	unknown
32	Murray district, Idaho	98,700	197,000
33	Seven-Up Pete area	unknown	unknown
34	Ninemile district	unknown	100,000 to 125,000

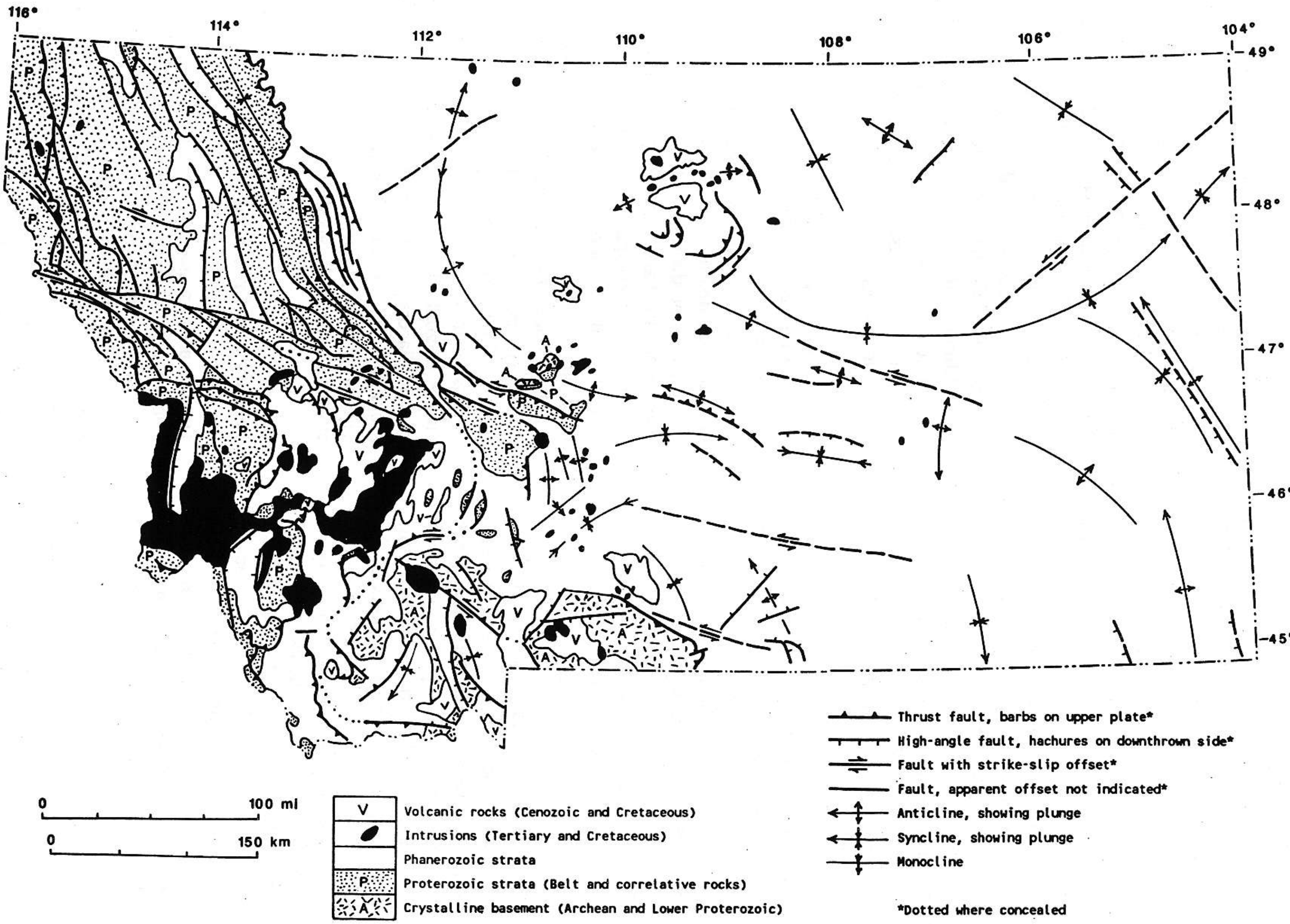


Figure 2. Generalized tectonic map of Montana. Modified from Lageson (1985).

al., 1986). The Boulder batholith is composite, consisting of calc-alkaline plutons mostly of quartz monzonite although compositions range from granite to gabbro. It covers about 6,000 km^2 and cuts Archean through Cretaceous rocks (Smedes, 1973). Elkhorn Mountains Volcanics are slightly older than the rocks of the Boulder batholith and are generally considered to be ejecta erupted from and intruded by comagmatic batholithic magmas of the same compositions (Hamilton and Myers, 1974). These volcanic rocks occur around the margins of the batholith and host numerous epigenetic gold deposits that appear to be related to plutonic rocks of the batholithic assemblage (Klepper *et al.*, 1957; Smedes, 1966).

Associated with these major batholiths are many smaller stocks scattered throughout southwestern Montana (Ruppel *et al.*, 1983; Wallace *et al.*, 1987). Intrusion of these and the larger plutons mostly post-dated Laramide (early Tertiary-Late Cretaceous) thrusting and rise of uplifts (Schmidt and O'Neill, 1982), except for the Boulder batholith which may pre-date or be contemporaneous with thrusting.

Alkalic igneous rocks of central Montana form laccolithic and stock-like intrusive complexes with associated sills and dikes (Pirsson, 1905; Larsen, 1940). This alkalic province extends from the Beartooth uplift on the south to the Sweetgrass Hills near the Canadian border (Fig. 3, location 29). Gold mineralization in the alkalic province was accompanied by varying amounts of silver and base metals and ranges from igneous-hosted stockworks and breccia pipes to replacements in adjacent strata, mainly Paleozoic carbonates (Giles, 1982). Gold production from this province is at least 1,850,000 ounces and reserves are estimated at 1,040,000 ounces, for a total of 2,890,000 ounces.

The principal mining districts in Montana are in the southwestern and west-central parts of the state and are found in the Tertiary-Cretaceous plutons and in the adjacent wallrocks of Archean through Cretaceous age (Billingsley and Grimes, 1917; Pardee and Schrader, 1933; Koschmann and Bergendahl, 1968; Woodward, 1986). Over 30 mining districts are within or adjacent to the Boulder batholith alone and are estimated to have produced at least 7,200,00 ounces of gold (Koschmann and Bergendahl, 1968; Krohn and Weist, 1977). The Butte district (Fig. 3, location 18) at the south end of the batholith has dominated production, with about 2,700,000 ounces of gold (Miller, 1973; Krohn and Weist, 1977; Elliott *et al.*, 1992). The Golden Sunlight deposit near the Willow Creek fault (Fig. 3) has production and reserves of approximately 3,000,000 ounces of gold.

Numerous gold placers throughout Montana are inferred to have been derived from lodes hosted by Tertiary-Cretaceous intrusions and their contact aureoles. One of the better examples is the Bannack district of southwestern Montana where placer production of 132,000 ounces was interpreted to have come from replacement bodies in Mississippian limestone near the contact with granodiorite (Koschmann and Bergendahl, 1968, p. 145). Other good examples are the Pioneer district (Loen, 1986), the Ophir district (Loen, 1990), and the German Gulch (Siberia) district (Loen and Pearson, 1989); each of this districts produced more than 200,000 ounces of gold. Figure 1 shows the abuntlance of gold placers in this part of the state and their proximity to Tertiary-Cretaceous intrusions. Similarly, gold placers in central Montana are clustered around the alkalic intrusions.

Tertiary and Cretaceous Volcanic Rocks

Tertiary and Cretaceous volcanic rocks ranging from rhyolite to basalt are widespread with large exposures in the central-northern, central, and southwestern parts of Montana. Small and widely scattered outcrops are found in the central-western part of the state. Upper Cretaceous Elkhorn Mountains Volcanics are found near the Boulder batholith; these rocks were discussed in connection with intrusions of

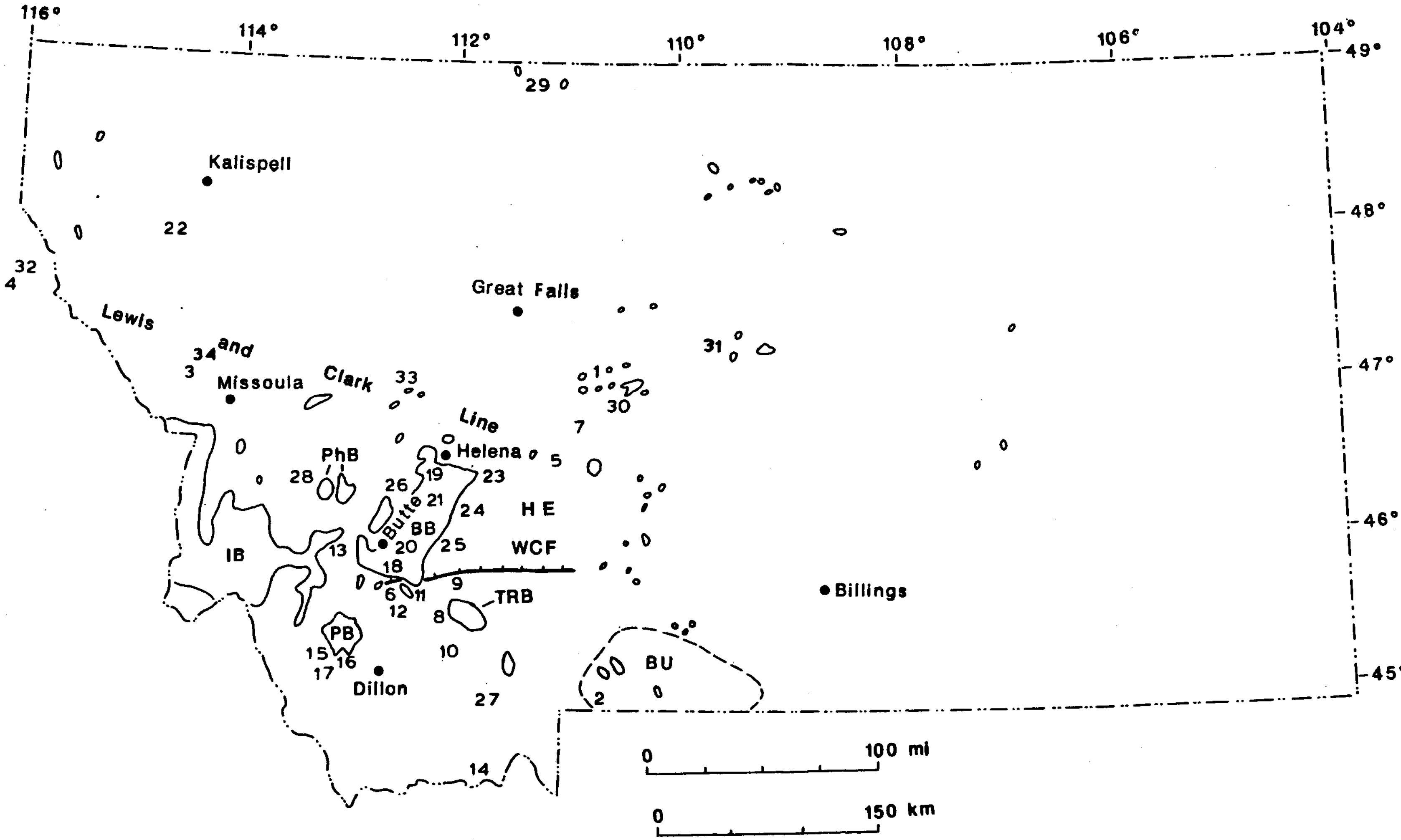


Figure 3. Map of Montana showing major features and numbered localities noted in text. Tertiary and Cretaceous plutons outlined with solid lines include Idaho batholith (IB), Boulder batholith (BB), Philipsburg batholith (PhB), Pioneer batholith (PB), and Tobacco Root batholith (TRB). Beartooth uplift (BU) shown with dashed line. WCF is Willow Creek fault. HE is Helena embayment of Proterozoic Belt basin. Numbered localities are referred to in text.

Tertiary-Cretaceous age.

Large, low-grade, disseminated gold deposits occur in Oligocene rhyolite flows, tuffs, and breccias at the Porphyry Dike and Paupers Dream properties (Knopf, 1913; Pardee and Schrader, 1933; Ruppel, 1963) in the Rimini district near the north end of the Boulder batholith (Fig. 3, location 19). In the Lowland district (Fig. 3, location 20) Tertiary felsic volcanic rocks host gold and silver deposits (Knopf, 1913; Roby *et al.*, 1960). A diatreme with quartz latite matrix cuts Tertiary ignimbrites and contains a large, disseminated and vein gold-silver deposit at Montana Tunnels (Sillitoe *et al.*, 1985) in the Colorado district (Fig. 3, location 21). Silver-lead mineralization with minor gold occurs in Tertiary volcanic rocks of the Hog Heaven (Flathead) district (Johns, 1970; Lange and Zehner, 1992) in northwestern Montana (Fig. 3, location 22). At the Seven-Up Pete area (Fig. 3, location 33) andesite hosts gold-bearing quartz-sulfide veins (Elliott *et al.*, 1992) with an estimated 5.2 million ounces of economically recoverable gold at the McDonald Meadows deposit. Other occurrences of mineralized Tertiary volcanic rocks are widely scattered in the southwestern part of the state.

Gold placers inferred to have been derived from Tertiary and(or) Cretaceous volcanic rocks are present in the Boulder Batholith region (Figs. 1, 3) at the Rimini (location 19), Lowland (location 20), Elkhorn Creek (location 25), and Emery (location 26) districts (Lyden, 1987). Some of the gold in placers at Indian Creek (location 23) and Radersburg (location 24) was probably derived from veins in Elkhorn Mountains Volcanics as well as from lodes hosted by intrusions (Klepper *et al.*, 1971).

Archean and Proterozoic Crystalline Basement Rocks

The principal examples of Archean age gold mineralization are the Jardine-Crevasse Mountain deposits (Seager, 1944) in the southwest part of the Beartooth uplift (Fig. 3, location 2). Gold ore occurs concordantly in folded, stratiform sulfide-silicate facies iron-formation (Cuthill *et al.*, 1989) interpreted to be exhalative. It is controversial whether the gold-quartz ore is primary and stratabound with minor remobilization or consists of structurally controlled veins folded into conformity with the iron-formation (Cuthill *et al.*, 1989).

Although there are several areas with oxide facies iron-formation in southwestern Montana (Bayley and James, 1973), major gold mineralization is reported only for the large sulfide-silicate facies iron-formation at Jardine-Crevasse Mountain. Minor gold is reported in quartz veins in oxide facies iron-formation at the Ruby Creek mine (Fig. 3, location 27) in the Gravelly Range (Heinrich and Rabbitt, 1960).

Numerous gold lodes are hosted by crystalline basement rocks in southwestern Montana, but most of these deposits probably are epigenetic and are related to emplacement of adjacent Tertiary-Cretaceous plutons. O'Neill and Schmidt (1989) proposed that gold-bearing ores hosted by Archean basement rocks in the Rochester and Silver Star mining districts (Fig. 3, locations 12 and 11, respectively) were concentrated during Cretaceous magmatism but were scavenged from the host rocks.

Middle Proterozoic Strata

The Middle Proterozoic Belt Supergroup (Fig. 4) is composed mainly of fine-grained siliciclastic rocks (shale, argillite, and siltite) with subordinate quartzite and carbonates, and minor conglomerate and breccia. This sequence thickens westward from a zero edge due to erosional truncation in the Little Belt Mountains of central Montana (Fig. 3, location 1) to a maximum of about 20,400 m near Alberton in western Montana (Fig. 3, location 3) where neither the top nor base of the section is exposed (Harrison, 1972, p. 1219). Facies changes across the Belt basin are, with a few exceptions,

			Western Montana	Central Montana
Cambrian			Flathead	Sandstone
Middle Proterozoic	Belt Supergroup	Missoula Group	Pilcher Quartzite	
			Garnet Range Fm.	
			McNamara Formation	
			Bonner Quartzite	
			Mt. Shields Formation	
			Shepard Formation	
			Snowslip Formation	
		Middle Belt Carbonate	Wallace Formation	Helena Limestone
		Ravalli Group	Empire Formation	Empire Formation
			St. Regis Formation	Spokane Formation
			Revett Formation	
		Lower Belt	Burke Formation	Greyson Shale
			Prichard Formation	Newland Limestone
				Chamberlain Shale
				Neihart Quartzite
Early Proterozoic and Late Archean			Basement not exposed	Crystalline rocks

Figure 4. Generalized correlation chart for units in the Belt Supergroup.

relatively minor. The lowest unit of the Belt Supergroup on the west is the Prichard Formation, composed principally of quartzite, siltite, and laminated argillite, generally considered to be correlative with a transgressive sandstone, shale, and limestone sequence in central Montana.

Gold and silver occurrences hosted by the Belt Supergroup and correlative strata are clustered in five major areas: (1) along the west-northwest-trending Lewis and Clark line, (2) in a northerly trending zone in northwestern Montana, (3) in a north-northeast-trending zone in the John Long Mountains of central-western Montana (Fig. 3, location 28), (4) along the westerly trending Willow Creek fault zone, and (5) adjacent to Tertiary-Cretaceous intrusions (Woodward, 1992). Other occurrences are numerous but appear to be scattered and without well-defined major trends.

Widespread evidence exists for syngenetic and(or) diagenetic copper and silver mineralization in strata of the Belt Supergroup and correlative rocks. Many of the epigenetic deposits hosted by Proterozoic strata appear to be remobilized from syngenetic or diagenetic deposits with concurrent enrichment. In Montana the Prichard Formation hosts at least 28 known gold deposits that mostly appear to be epigenetic. They are stratabound within the Prichard and commonly are concordant with bedding. This supports gold remobilization from syngenetic deposits in the Prichard Formation. Large amounts of placer gold appear to have been derived from lodes hosted by the Prichard Formation, particularly in the Murray district of northern Idaho (Shenon, 1938).

Baitis (1988) reported that stratiform orthoclase-rich sandstones, siltstones and ferroan carbonates containing minor amounts of pyrite and 0.01 to 0.05 ounce per ton gold occur in the Greyson Shale in the York district of the Big Belt Mountains. These deposits have been interpreted to be syngenetic, diagenetic, or epigenetic (Thorson *et al.*, in press) with local remobilization during Laramide (Late Cretaceous-early Tertiary) deformation (Woodward, 1992).

About 65 km east-northeast of York there is a massive sulfide deposit hosted by the Newland Formation at Sheep Creek. This deposit is next to the Volcano Valley fault (Fig. 3, location 7), a growth fault with the south side down during deposition of the Newland Limestone (Godlewski and Zieg, 1984). Himes *et al.* (1988) reported that subsurface mineralization at Sheep Creek consists of thinly laminated to massive pyrite, chalcopyrite, cobaltiferous pyrite, sphalerite, gold, and galena interbedded with clastic sediments and barite; the sulfide bodies are associated with debris-flow conglomerates and are interpreted to have formed from exhalative activity near a growth fault (Himes *et al.*, 1988).

Widespread but volumetrically minor magmatism that occurred in the Belt basin during Middle and Late Proterozoic time may have been important in mineralization processes. Mafic sills, dated at 1,075-1,200 Ma, occur in the northern and east-central part of the Belt basin (Reynolds, 1984). Diabasic dikes and sills that are about 750 to 830 Ma (Reynolds, 1984) and clearly post-date deposition of the Belt rocks are present at numerous localities. These intrusions, together with Tertiary-Cretaceous plutons, may have caused multiple remobilization of metals hosted by Belt rocks (Woodward, 1992).

McClernan (1984) speculated that lode deposits in the Big Belt uplift (Fig. 3, location 5) contain remobilized paleoplacer gold initially trapped in algal mats of the Newland Limestone. The ultimate derivation of gold was interpreted to be from Archean terrane south of the Willow Creek growth fault that bounds the south side of the Helena embayment (Fig. 3).

Middle Cambrian Strata

Ruppel (1985) proposed that gold might be trapped in algal reefs in the upper part of the

Meagher Limestone or in the overlying Park Shale, both Middle Cambrian, in the western Tobacco Root Mountains of southwestern Montana (Fig. 3, location 8), possibly with redistribution during diagenesis of the shale. The lode-gold deposits occur as quartz veins and replacements in the Meagher Limestone and in the underlying Wolsey Shale and Flathead Sandstone and extentl downward into crystalline basement rocks of Archean age. No gold deposits in the area are known in the sedimentary rocks above the Meagher Limestone. These lodes and associated gold placers are restricted to areas where the Meagher is present and contains algal reefs in its upper part. The reefs are interpreted to fringe an island of crystalline rock that was surrounded by the Cambrian sea (Ruppel, 1985).

The ultimate source of the gold inferred to have been trapped in the algal reefs would presumably have been the deeply weathered Archean crystalline rocks exposed to erosion during the Middle Cambrian. If the lode-gold deposits were derived from the Cambrian rocks during compaction, dewatering, and diagenesis of the Park Shale, and were redeposited in and below the Meagher Limestone, it follows that the lodes must be of Cambrian age also.

This speculative model was used by Ruppel (1985) to explain many of the gold deposits of southwestern Montana (Fig. 3) including some of the lodes and associated placers in the Renova (location 9), Virginia City (location 10), Silver Star (location 11), Rochester (location 12), and French Creek (location 13) mining districts, and the Odell Creek (location 14) and Dyce Creek (location 15) areas. However, the gold in these areas could have been derived from sources other than from algal mats. Ruppel further suggested that some of the hydrothermal deposits associated with Late Cretaceous and Tertiary intrusions contain gold remobilized from Middle Cambrian rocks, particularly in the Fish Creek-Moose Creek (location 6), Argenta (location 16), and Bannack (location 17) districts.

Miscellaneous Secondary Sources

Tertiary placers reworked by fluvial processes and Pleistocene glacial deposits are sources for gold found in some placers in Montana (Loen, 1989). Placers along the lower parts of major rivers, such as the Missouri and Yellowstone, are generally thought to be the result of reworking of placers upstream (Lyden, 1987). The precise sources of the gold have not been determined for most of these deposits and consequently the ultimate lode sources are unknown.

Gold-bearing glacial deposits are present in northwestern Montana at Libby Creek, Fisher River, Stillwater River, and Vermillion Creek, in the Gold Creek drainage in Powell County of central-western Montana and in the Gravelly Range in the southwest part of the state (Lyden, 1987). These deposits generally contain small amounts of gold and most have been unprofitable to mine. Numerous alluvial placers containing minor amounts of gold inferred to have been derived from glacial deposits are found in the Gravelly Range. Some alluvial placers in the Gold Creek area appear to have concentrated gofd derived from subeconomic glacial deposits (Loen, 1994).

USE OF PLACER GOLD MORPHOLOGY

In a study of 468 gold grains from quartz veins and four placer localities along a distance of 13 km from the lode in the Pioneer district near the north end of the Philipsburg batholith (Fig. 1), Loen (1993) observed changes in morphology of medium and coarse gold nuggets. Transport from lode sources resulted in decreased grain size, increased rounding, polishing, folding of grains, decreased percentages of grains with primary mineral inclusions, and progressive increase in mean flatness index. Loen (1993) also concluded that chemical accretion did not affect the morphology of the placer gold grains; an increase in fineness (i.e., a higher gold to silver ratio) limited to thin (tens of microns) rims of the gold grains was attributed to supergene weathering processes (Loen, 1986). The changes in placer gold morphology demonstrated by Loen (1993), used in conjunction with the regional geology, styles of lode-gold mineralization, sedimentologic studies of gold-bearing sediment dispersal, and geochemical studies, provide an effective means of tracing placer gold to its source.

IMPLICATIONS FOR EXPLORATION

Gold placers (Fig. 1) can be used to identify potential target areas for lode exploration and in Montana, most lode-gold discoveries were preceded by discovery of related gold placers (McCulloch, 1989). Of particular interest in exploration are those placers with no known source, as there is a possibility that an undiscovered economic lode may be present and was the source for the placer gold. It is also possible that a sub-economic lode was the source or that the source was entirely eroded. With these caveats in mind, the following discussion outlines exploration approaches for lode-gold deposits using placers with no known sources. In addition, the regional geology and style of primary gold mineralization must be known in order to use this approach. McCulloch (1989) discussed methods of defining local and regional lode targets: local targets involve a single gold-bearing drainage extending from a central area; parallel gold-bearing drainages confined to one side of a major ridge define a regional target; and dense patterns having large numbers of placers with no recognizable relations to one another also mark regional targets.

Tertiary-Cretaceous intrusions and their adjacent wallrocks are major hosts for gold mineralization in the southwestern and central parts of the state. Undiscovered mineral deposits may be present in southwestern Montana near basal zones of the Medicine Lodge and Grasshopper thrust plates where older, steep faults or fracture zones might have channeled intrusions and associated mineralizing solutions into favorable host rocks, mainly Paleozoic carbonates (Ruppel and Lopez, 1984). Mesozoic clastic rocks are also important host rocks, and at the Beal mine form the source area for 250,000 ounces of placer gold in German Gulch (Loen and Pearson, 1989).

Placers that have no known sources but appear to be spatially associated with Tertiary-Cretaceous intrusions or their wallrocks are found near the Little Belt Mountains (Fig. 3, location 30), the Moccasin Mountains (Fig. 3, location 31), the Boulder batholith, the Pioneer batholith, the Idaho batholith, and several plutons in the Beartooth uplift of southern Montana. Lode targets in these areas are primarily porphyry-hosted stockwork or disseminated ore bodies, and replacements and disseminations in flanking Paleozoic carbonates.

Exploration for gold hosted by volcanic rocks might have greater success in the region of the Boulder batholith as this is where most of the major, volcanic-hosted deposits have been found. Recent exploration activity, however, northwest of Helena in the Seven-Up Pete area (Fig. 3, location 33) indicates that large lode-gold deposits are found in Tertiary volcanic rocks in other areas also. Sheared and brecciated andesite lavas host low-grade gold and silver deposits there (Pardee and Schrader, 1933).

The northeast-trending Great Falls tectonic zone (O'Neill and Lopez, 1985) is a broad zone that encompasses the Boulder and Pioneer batholiths and many gold placers in southwestern Montana (Fig. 1) and adjacent parts of Idaho. This zone may have controlled emplacement of many Tertiary-Cretaceous intrusions.

Exploration for lodes in drainages underlain by crystalline basement rocks should focus on exhalative iron-formations or on deposits such as quartz veins related to Tertiary-Cretaceous magmatic activity. It is also possible that paleoplacers might be found in some of the quartzites and metaconglomerates of Archean age in southwestern Montana although no such deposits have been described; deposits of this kind could have been overlooked by prospectors who focused on quartz veins.

If basement growth faults were active during Belt sedimentation and are the major control in localization of lode gold deposits concentrated along the Lewis and Clark line, then the trends of these postulated faults in Belt strata are favorable for mineral exploration. These growth faults presumably mark the margins of differentially subsided basement blocks and are subtle features that can be inferred by detailed stratigraphic and sedimentologic studies as discussed by Winston (1983, 1986a, 1986b).

Numerous gold placers of unknown sources along the Lewis and Clark line occur in streams draining areas underlain by Belt strata, suggesting opportunities for lode exploration. A particularly attractive exploration target is the Prichard Formation which hosts lode deposits in the Murray district, Idaho (Fig. 3, location 32) at the western end of the Lewis and Clark line. Total gold production between 1884 and 1934 was 295,700 ounces, with 98,700 ounces from lodes (Shenon, 1938). In addition, five placers in the Ninemile district (Fig. 3, location 34), about 30 km northwest of Missoula, probably were derived from lodes in the Prichard and suggest that exploration would be warranted there.

Middle Cambrian strata may be the source of some of the gold deposits in southwestern Montana (Ruppel, 1985). If this hypothesis is correct, there could be undiscovered gold deposits, including veins, fine-grained disseminations in the Cambrian rocks, and placer gold weathered from these rocks and deposited in present stream gravels, or in ancient stream gravels having different drainages than the present ones (Ruppel, 1967). Gold placers with unknown sources are present in this part of the state and present opportunities for exploration of potential lodes hosted by Middle Cambrian rocks.

CONCLUSIONS

Major primary lode sources of gold placers in Montana are inferred to be: (1) Archean iron-formations containing gold-quartz mineralization, (2) syngenetic and(or) diagenetic Proterozoic Belt-hosted deposits, some of which may have been remobilized during Late Proterozoic and(or) Tertiary-Cretaceous magmatism, (3) Cambrian carbonate strata with gold trapped in algal mats, (4) veins, replacements, and skarns associated with Tertiary-Cretaceous intrusions, and (5) disseminated and stockwork gold deposits in Tertiary and Cretaceous volcanic rocks. Secondary sources include Tertiary placers and Pleistocene glacial deposits.

The majority of the gold placers are in the southwestern part of the state (Fig. 1). This area is geologically complex, with Archean to Lower Proterozoic crystalline basement rocks, Middle Proterozoic strata of the Belt Supergroup and correlatives, Paleozoic and Mesozoic strata, thrust faults of the Cordilleran fold and thrust belt, major intrusions of the Boulder, Tobacco Root, Philipsburg, Pioneer, and Idaho batholiths along with satellite plutons, and uplifts of the Rocky Mountain foreland, and basin-range fault-blocks. Regional exploration programs there should consider all of the inferred lode sources noted previously.

Numerous gold placers also occur along the Lewis and Clark line (Fig. 1), a major structural zone and locus of mineralization (Billingsley and Locke, 1941) extending east-southeasterly from the Couer d'Alene mining district of northern Idaho (Fig. 1, location 4) to west-central Montana. This lineament is interpreted to have been marked in part by growth faults during Proterozoic time with localization of sedimentary exhalative mineral deposits during deposition of Belt (Middle Proterozoic) strata. The Great Falls tectonic zone (O'Neill and Lopez, 1985) appears to be a fundamental assemblage of diverse structural features that influenced the tectonic development of southwestern Montana, including Tertiary-Cretaceous plutons and volcanic centers and their associated gold deposits.

Many gold placers in Montana have no known sources, suggesting the possible presence of undiscovered lode-gold deposits. Some of the lodes may be of sufficient size and grade to be economic. Numerous lode exploration targets can be defined using gold placers, regional geology, and styles and timing of mineralization as guides.

ACKNOWLEDGEMENTS

I thank Ian Lange, Edward Ruppel, and Jeffrey Loen for reviewing the manuscript and offering many helpful suggestions.

REFERENCES

Arth, J.G., Zen, E., Sellers, G., and Hammerstrom, J., 1986, High initial Sr isotopic ratios and evidence for magma mixing in the Pioneer batholith of southwest Montana: Journal of Geology, v. 94, p. 419-430.

Baitis, H.W., 1988, York-A Proterozoic "shale" - hosted gold system in Montana: Northwest Mining Association, 94th Annual Convention, November 30 - December 3, 1988, Spokane, Washington, Abstract Booklet, p. 7.

Bayley, R.W., and James, H.L., 1973, Precambrian iron-formations of the United States: Economic Geology, v. 68, p. 934-959.

Billingsley, P., and Grimes, J.A., 1917, Ore deposits of the Boulder batholith of Montana: American Institute of Mining Engineers Transactions, v. 58, p. 284-361.

Billingsley, P., and Locke, A., 1941, Structure of ore districts in the continental framework: American Institute of Mining Engineers Transactions, v. 144, p. 9-64.

Chadwick, R.A., 1965, Volcanism in Montana: Northwest Geology, v. 1, p. 1-20.

Chesson, S., Griffith, T., and Wallace, R.R., 1984, Maps showing mineral occurrence data for the Wallace 1°x2° quadrangle, Montana and Idaho: U.S. Geological Survey Miscellaneous Field Studies Map MF-1354-F.

Cuthill, J., Oliver, D., and Hofer, W., 1989, The structure of the Jardine gold deposit, Jardine, Montana: Montana Geological Society 1989 Field Conference Guidebook, v. I, p. 371-373.

Elliott, J.E., Loen, J.S., Wise, K.W., and Blaskowski, M.J., 1992, Maps showing locations of mines and prospects of the Butte 1°x2° quadrangle, Montana: U.S. Geological Survey Miscellaneous Investigation Series Map I-2050-C, with 147 page pamphlet.

Frishman, D., Elliott, J.E., Foord, E.E., Pearson, R.C., and Raymond, W.H., 1990, Map showing the location of significant lode and placer gold mines in Montana: U.S. Geological Survey Open-File Report 90-0241, 62 p., 2 plates.

Giles, D.L., 1982, Gold mineralization in the laccolithic complexes of central Montana: Denver Region Exploration Geologists Symposium, The Genesis of Rocky Mountain Ore Deposits: Changes with Time and Tectonics, p. 157-162.

Godlewski, D.W., and Zieg, G.A., 1984, Stratigraphy and depositional setting of the Precambrian Newland Limestone: Montana Bureau of Mines and Geology Special Publication 90, p. 2-4.

Hamilton, W., and Myers, W.B., 1974, Nature of the Boulder batholith of Montana: Geological Society of America Bulletin, v. 85, p. 365-378.

Harrison, J.E., 1972, Precambrian Belt basin of northwestern United States: Its geometry, sedimentation, and copper occurrences: Geological Society of America Bulletin, v. 83, p. 1215-1240.

Heinrich, E.W., and Rabbitt, J.C., 1960, Pre-Beltian geology of the Cherry Creek and Ruby Mountains areas, southwestern Montana: Montana Bureau of Mines and Geology Memoir 38, 40 p.

Himes, M.D., Bourns, F.T., Golden, H.C., Perello, J.A., Taylor, S.A., Winder, R.J., Rankin, P.W., Jennings, S., and Zieg, G.A., 1988, The Sheep Creek project, a Proterozoic sediment-hosted massive sulfide deposit, Meagher County, Montana: Northwest Mining Association, 94th Annual Convention, November 30 -December 3, 1988, Spokane, Washington, Abstract Booklet, p. 17.

Hyndman, D.W., 1983, The Idaho batholith and associated plutons, Idaho and western Montana, in J.A. Roddick, ed., Circum-Pacific Batholiths: Geological Society of America Memoir 159, p. 213-240.

Johns, W.M., 1970, Geology and mineral deposits of Lincoln and Flathead Counties, Montana: Montana Bureau of Mines and Geology Bulletin 79, 182 p.

Klepper, M.R., Weeks, R.A., and Ruppel, E.T., 1957, Geology of the southern Elkhorn Mountains, Jefferson and Broadwater Counties, Montana: U.S. Geological Survey Professional Paper 292, 82 p.

Klepper, M.R., Ruppel, E.T., Freeman, V.L., and Weeks, R.A., 1971, Geology and mineral deposits, east flank of the Elkhorn Mountains, Broadwater County, Montana: U.S. Geological Survey Professional Paper 665, 66 p.

Klepper, M.R., Robinson, G.D., and Smedes, H.W., 1974, On the nature of the Boulder batholith of Montana: Geological Society of America Bulletin, v. 82, p. 1563-1580.

Knopf, A., 1913, Ore deposits of the Helena mining region, Montana: U.S. Geological Survey Bulletin 527, 143 p.

Koschmann, A.H., and Bergendahl, M.H., 1968, Principal gold-producing districts of the United States: U.S. Geological Survey Professional Paper 610, 283 p.

Krohn, D.H., and Weist, M.M., 1977, Principal information on Montana mines: Montana Bureau of Mines and Geology Special Publication 75, 150 p.

Lageson, D.R., 1985, Tectonic map of Montana: Bozeman, Montana, Department of Earth Sciences, Montana State University, scale 1:3,500,000.

Lange, I.M., and Zehner, R.E., 1992, Geologic map of the Hog Heaven volcanic field, northwestern Montana: Montana Bureau of Mines and Geology, Geologic Map 53, scale 1:50,000, with text, 6 p.

Larsen, E.S., 1940, Petrographic province of central Montana: Geological Society of America Bulletin, v. 51, p. 887-948.

Loen, J.S., 1986, Origin of gold placers in the Pioneer distrldt, Powell County, Montana: unpublished M.S. thesis, Fort Collins, Colorado State University, 164 p.

Loen, J.S., 1989, Climatic and tectonic controls on the formation of Tertiary gold placers, Pioneer district, Powell County, Montana: Montana Geological Society 1989 Field Conference Guidebook, v. I, p. 375-381.

Loen, J.S., 1990, Lode and placer gold deposits in the Ophir district, Powell and Lewis and Clark Counties, Montana: unpublished Ph.D. dissertation, Fort Collins, Colorado State University, 264 p.

Loen, J.S., 1992, Mass-balance constraints on formation of gold placers: Possible solutions to "source-area problems": Economic Geology, v. 87, p. 1624-1634.

Loen, J.S., 1993, Implications of changes in placer gold morphology, Pioneer district, Montana: Northwest Geology, v. 22, p. 29-41.

Loen, J.S., 1994, Origin of placer gold nuggets and history of formation of glacial gold placers, Gold Creek, Granite County, Montana: Economic Geology, v. 89, p. 91-104.

Loen, J.S., and Pearson, R.C., 1989, Maps showing locations of mines and prospects in the Dillon 1°x2° quadrangle, Idaho and Montana: U.S. Geological Survey Miscellaneous Investigations Series Map I-1803-C, scale 1:250,000, contains 85 p. pamphlet.

Lyden, C.J., 1987, Gold placers of Montana: Montana Bureau of Mines and Geology Reprint 6, 120 p.

McClernan, G.H., 1984, Gold deposits of the Big Belt Mountains: Montana Bureau of Mines and Geology Bulletin 121, p. 33-40.

McCulloch, R., 1989, Gold placers: A signature for lode-gold deposits in Montana: Montana Bureau of Mines and Geology Bulletin 128, p. 25-34.

Miller, R.N., 1973, Production history of the Butte district and geological function, past and present, in Miller, R.N., ed., Guidebook for the Butte Field Meeting, Society of Economic Geologists, Butte, Montana, August 18-21, 1973: Butte, Montana, Anaconda Company, p. F-1-F-10.

O'Neill, J.M., and Lopez, D.A., 1985, The Great Falls Tectonic Zone: American Association of Petroleum Geologists Bulletin, v. 69, p. 437-447.

O'Neill, J.M., and Schmidt, C.J., 1989, Tectonic setting and structural control of gold deposits in cratonic rocks of the Rochester and Silver Star mining districts, Highland Mountains, southwestern Montana: Montana Geological Society 1989 Field Conference Guidebook, v. I, p. 393-402.

Pardee, J.T., and Schrader, F.C., 1933, Metalliferous deposits of the greater Helena mining region, Montana: U.S. Geological Survey Bulletin 842, 318 p.

Pearson, R.C., Loen, J.S., Trautwein, C.M., Moll, S.H., and Ruppel, E.T., 1991, Map showing mineral resource assessment for placer gold and silver, Dillon 1°x2° quadrangle, Idaho and Montana: U.S. Geological Survey Miscellaneous Investigations Series Map I-1803-F, scale 1:250,000.

Pirsson, L.V., 1905, Petrographic province of central Montana: American Journal of Science, v. 20, p. 35-40.

Reynolds, M.W., 1984, Tectonic setting and development of the Belt basin, northwestern United States: Montana Bureau of Mines and Geology Special Publication 90, p. 44-46.

Roby, R.N., Ackerman, W.C., Fulkerson, F.B., and Crowley, F.A., 1960, Mines and mineral deposits (except fuels), Jefferson County, Montana: Montana Bureau of Mines and Geology Bulletin 16, 122 p.

Ruppel, E.T., 1963, Geology of the Basin quadrangle, Jefferson, Lewis and Clark, and Powell Counties, Montana: U.S. Geological Survey Bulletin 1151, 121 p.

Ruppel, E.T., 1967, Late Cenozoic drainage reversal, east-central Idaho, and its relation to possible undiscovered placer deposits: Economic Geology, v. 62, p. 648-663.

Ruppel, E.T., 1985, The association of Middle Cambrian rocks and gold deposits in southwest Montana: U.S. Geological Survey Open-File Report 85-207, 26 p.

Ruppel, E.T., artd Lopez, D.A., 1984, The thrust belt in southwest Montana and east-central Idaho: U.S. Geological Survey Professional Paper 1278, 41 p.

Ruppel, E.T., O'Neill, J.M., and Lopez, D.A., 1983, Preliminary geologic map of the Dillon 1° x 2° quadrangle, Montana: U.S. Geological Survey Open-File Report 83-168, scale 1:250,000.

Schmidt, C.J., and O'Neill, J.M., 1982, Structural evolution of the southwest Montana transverse zone, in R.B. Powers, ed., Geologic Studies of the Cordilleran Thrust Belt: Rocky Mountain Association of Geologists, Denver, v. I, p. 167-180.

Seager, F.G., 1944, Gold, arsenic, and tungsten deposits of the Jardine-Crevasse Mountain district, Park County, Montana: Montana Bureau of Mines and Geology Memoir 23, 111 p.

Shawe, D.R., and Wier, K.L., 1989, Gold deposits in the Virginia City-Alder Gulch district, Montana: U.S. Geological Survey Bulletin 1857-G, p. G14-G19.

Shenon, P.J., 1938, Geology and ore deposits near Murray, Idaho: Idaho Bureau of Mines and Geology Pamphlet No. 47, 44 p.

Sillitoe, R.H., Grauberger, G.L., and Elliott, J.E., 1985, A diatreme-hosted gold deposit at Montana Tunnels, Montana: Economic Geology, v. 80, p. 1707-1721.

Smedes, H.W., 1966, Geology and igneous petrology of the northern Elkhorn Mountains, Jefferson and Broadwater Counties, Montana: U.S. Geological Survey Professional Paper 510, 116 p.

Smedes, H.W., 1973, Regional setting and general geology of the Boulder batholith, Montana, in R.N. Miller, ed., Guidebook for the Butte field meeting, Society of Economic Geologists, Butte, Montana, August 18-23, 1973: Butte, Montana, Anaconda Company, p. A-1-A-6.

Thorson, J.P., White, B.G., and Baitis, H.W., in press, Gold resources in the Greyson Formation, Big Belt Mountains, Montana: Part II. Mineralization and genesis: Montana Bureau of Mines and Geology Special Publication.

Wallace, C.A., Schmidt, R.G., Lidke, D.J., Waters, M.R., Elliott, J.E., French, A.B., Whipple, J.W. Zarske, S.E., Blaskowski, M.J., Heise, B.A., Yeoman, R.A., O'Neill, J.M. Lopez, D.A., Robinson, G.D., and Klepper, M.R., 1987, Geologic map of the Butte 1°x2° degree quadrangle, western Montana: U.S. Geological Survey Open-File Report 86-292.

Winston, D., 1983, Middle Proterozoic Belt Basin syndepositional faults and their influence on Phanerozoic thrusting and extension: American Association of Petroleum Geologists Bulletin, v. 67, p. 1361.

Winston, D., 1986a, Generalized geologic map of the Belt-Purcell basin, Montana, Idaho, Washington and Wyoming, U.S.A.; Alberta and British Columbia, Canada: Montana Bureau of Mines and Geology Special Publication 94, 311 p., Sheet 1.

Winston, D., 1986b, Middle Proterozoic tectonics of the Belt basin, western Montana and northern Idaho: Montana Bureau of Mines and Geology Special Publication 94, Belt Supergroup, p. 245-257.

Woodward, L.A., 1986, Tectonic origin of fractures for fissure vein emplacement in the Boulder batholith and adjacent rocks, Montana: Economic Geology, v. 81, p. 1387-1395.

Woodward, L.A., 1992, Tectonic setting of Au-Ag deposits hosted by Proterozoic strata along the Lewis and Clark line, west-central Montena, in Basement Tectonics 8: Characterization and Comparison of Ancient and Mesozoic Continental Margins - Proceedings of the 8th International Conference on Basement Tectonics (Butte, Montana, 1988), Kluwer Academic Publishers, Dordrecht, The Netherlands, p. 653-663.

MONTANA SAPPHIRES - THE VALUE OF COLOR

David W. Baker - Paine Gulch, Monarch, MT 59463

ABSTRACT

A model for magma genesis provides insight into the origin of high quality "cornflower blue" sapphire in the Yogo dike near Utica, Montana. Ultramafic lamprophyre magma rose through the lithospheric mantle in the Eocene and "pooled" at or near the base of the abnormally thick (50-55 km) and abnormally cool (550°-600°C) crust, where it assimilated high alumina rocks in the eclogite metamorphic facies. Corundum (sapphire) formed by replacing kyanite in a desilication reaction. Rare inclusions in the sapphire include primary igneous analcime that has a stability range of 8-14 kb water pressure and 600-640°C and nearly pure carbon dioxide. Under reducing conditions ferrous iron and titanium were incorporated in the corundum, creating the violet-blue hue. The reducing conditions also limited the solubility of iron in corundum as shown by the low iron content and the weak peaks for Fe^{+3} in absorption spectra for the sapphire.

Although corundum is stable over a wide range of conditions, gem-quality corundum apparently forms in the temperature range of 600°-750°C (and pressure range of 6-14 kb). The lithosphere underlying the three principal placer deposits of sapphire in western Montana—Missouri River near Helena, Dry Cottonwood Creek, and Rock Creek—is much thinner than under central and eastern Montana. The geothermal gradient is much higher; and the 600°-750° temperature "window" is at mid-crustal depths of 25-35 km, where conditions are predominately oxidizing and ferric iron is abundant. The iron content in these sapphires is typieally 5 times higher than at Yogo. Heat-treating can enhance the color, but can not eliminate the high iron content. The resulting "steel" blue is commercially viable, but not as attractive as cornflower blue.

Mining the placer deposits involves relatively low cost, high volume surface operations. The normally high costs of underground mining at Yogo are increased due to the complexity caused by pre-intrusion and post-intrusion karst development. Thus, heat-treated steel blue sapphire from the placer deposits can supply a high volume market at competitive prices, whereas the much more expensive cornflower blue sapphire from Yogo Gulch will continue to supply a smaller up-scale market.

INTRODUCTION

Montana sapphire deposits have been mined for the last 100 years. Before the introduction of cheap synthetic corundum in the 1920's, Montana sapphire was used for abrasives, watch bearings and other industrial purposes. Now the value of Montana sapphire lies in its color and use as gemstones. The main sapphire deposits in the state are (from east-northeast to west-southwest): Yogo Gulch (Judith Basin County), Missouri River near Helena (Lewis and Clark County), Dry Cottonwood Creek (Powell County), and Rock Creek (Granite County) (Clabaugh, 1952) (Fig. 1). There are systematic changes in the sapphires which correlate with the geologic setting. The best known deposit is at Yogo Gulch (Clabaugh, 1952; Voynick, 1987; Dahy, 1991). This article describes why color in Yogo sapphire creates value and how this deposit of colored gemstones formed, and then discusses how the regional geology and geophysics affects the other deposits in terms of color and value.

Northwest Geology, v. 23, p. 61-75, 1994

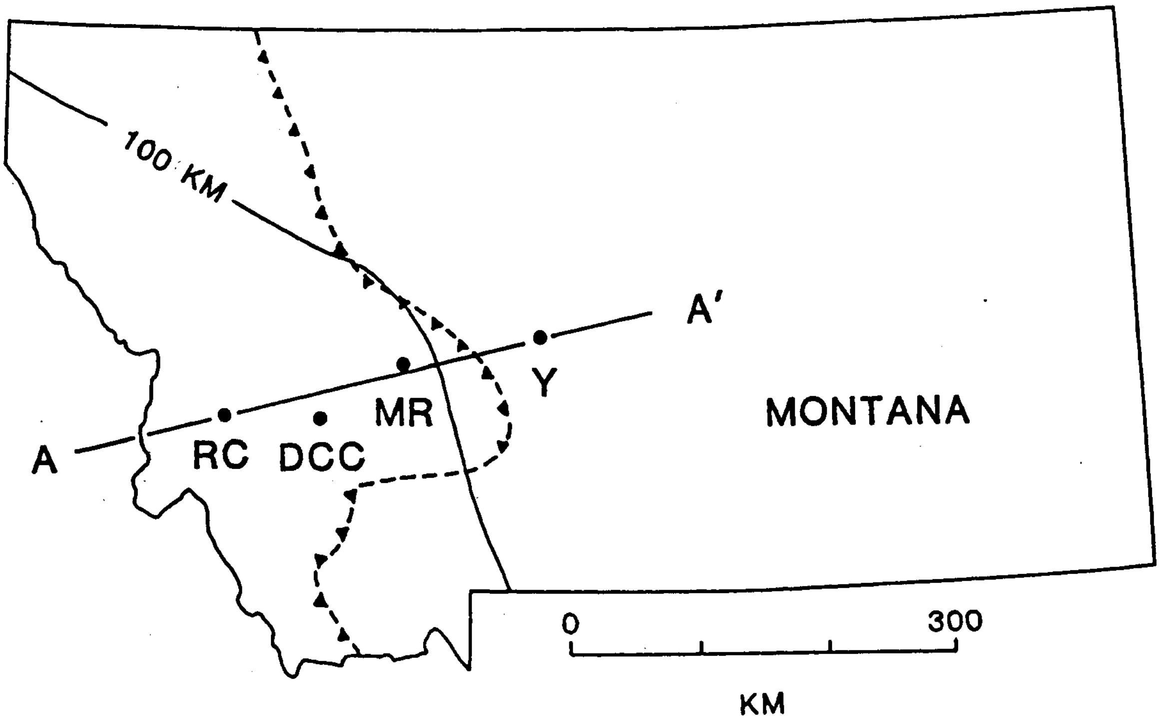


Figure 1. Location of major sapphire deposits in Montana. RC - Rock Creek, DCC - Dry Cottonwood Creek, MR - Missouri River, Y - Yogo. Dashed line with sawteeth - eastern limit of thin-skinned tectonics (= fold-and-thrust belt); thick-skinned tectonics (=Laramide foreland uplifts) lie to the east. Solid line - 100 km contour on the base of the lithosphere. A-A' shows location of cross section in Fig. 4.

COLORED CORUNDUM

The Value of Color and the Structure of the Retina. The value of colored, precious gemstones is a function of hue, saturation or chroma of the color, clarity, cut, weight in carats, rarity, and location of the deposit (Hughes, 1990; Newman, 1994). In suites of gemstones showing a range of hues but with the other parameters the same, certain hues are most highly valued. Value versus wavelength curves correlate with the spectral response of the retina, suggesting a scientific basis for the adage, "Beauty is in the eye of the beholder." The electrical outputs of the red-, green-, and blue-sensitive cones of the retina are "processed" so that "red minus green" and "blue minus yellow" signals are sent to the brain via the optic nerve (Boynton, 1979, p. 207-250). The (absolute value of the) spectral output of these two "color channels" (Boynton, 1979, Fig. 7.4; Wyszecki and Stiles, 1982, p. 648-653) is a good approximation to a plot of dollars versus wavelength on color grading charts for sapphires and emeralds, and to a lesser extent, for rubies (Fig. 2). Both curves peak on the violet side of blue (sapphire), the yellow side of green (emerald) and the orange side of red (ruby). The cornflower blue of Yogo sapphires is thus a "world-class" hue. [Kelly and Judd (1976) give quantitative definitions of the color

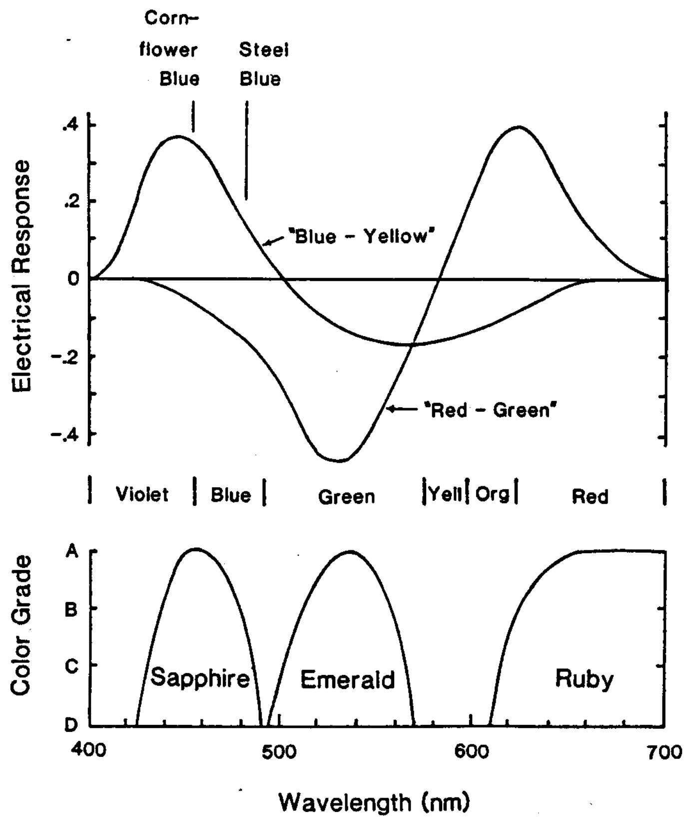


Figure 2. Comparison of electrical response of opponent type signals sent to the brain via two "color channels" in the optic nerve (Wyzsecki and stiles, 1982, p. 646-648) with generalized color grading curves for sapphire, emerald and ruby (*cf.* Roberts, 1986; Rubin, 1992; Suwa, 1993).

hue names used in this article. Cornflowers (*Centaurea cyanus*)—better known as bachelor's buttons—come in hues ranging from blue through magenta to red, paralleling the hues of Yogo sapphires. The blue pigment in this flower is an organic complex with iron and/or aluminum (Bayer, 1958; Bayer *et al.*, 1960; Asen and Jurd, 1967; Asen and Horowitz, 1974).]

Color in Blue Sapphire. Cornflower blue sapphire from Yogo Gulch owes its color to an intervalence charge transition (IVCT) in which light absorbed by the crystal causes electrons to jump from Fe^{+2} atoms to Ti^{+4} atoms (Townsend, 1968; Nassau, 1983, p. 140-145). The Fe and Ti substitute for Al in the corundum lattice in trace concentrations—0.14 and 0.013 atomic percent, respectively (Meyer and Mitchell, 1988). Absorption spectroscopy shows that this vivid, violetish blue results from transmission of some red light through the gemstones in addition to blue light (Fig. 3). The addition of Fe^{+3} absorbs the reddish component and shifts the hue of the sapphire towards blue-green (Schmetzer and Bank, 1980). Addition of chromium produces violet, purple, magenta and red hues (Schmetzer and Bank, 1981). Some chromium-bearing Yogo sapphires exhibit the alexandrite effect, appearing blue in daylight and red under incandescent light.

Cornflower Blue Sapphire Formed under Reducing Conditions. Studies of aluminum oxide ceramics show that the high temperature solubility of iron in corundum increases with oxygen fugacity (Meyers *et al.*, 1980). The low concentration of iron in Yogo sapphire and the lack of strong peaks for ferric iron in the absorption spectra indicate that the corundum formed under

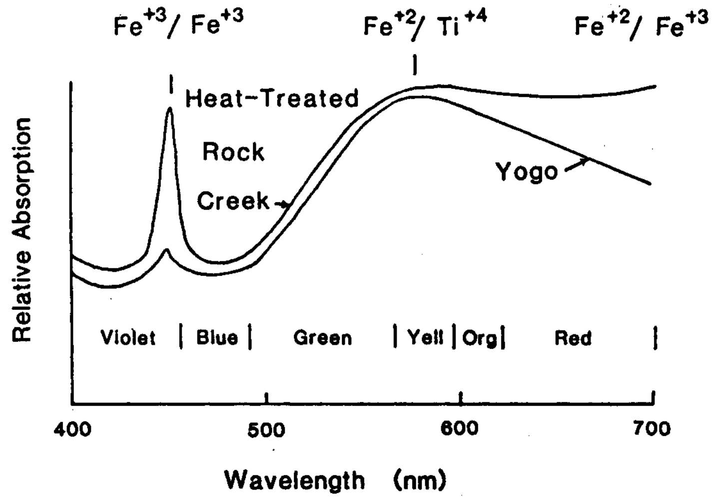


Figure 3. Absorption spectra for blue sapphire from Montana. Polarized light, beam parallel corundum *c* axis, electric vector **E** normal to *c*. Yogo curve (courtesy G. Rossman, Caltech), Rock Creek curve (after Themelis, 1992). The strong absorption by Fe^{+2}/Fe^{+3} at red end of spectrum is the tail of a broad band centered at 875 nm. As discussed by Billmeyer and Saltzman (1981) and Wyzsecki and Stiles (1982), these curves can be integrated to give CIE tristimulus values, Munsell color values, color opponent coordinates known as CIELAB, and the dominant wavelength.

reducing conditions. Sapphire crystals coated with small hercynite grains are abundant—another indication of reduction rather than oxidation.

ORIGIN OF MONTANA SAPPHIRE

The Yogo Sapphire Deposit

Sapphires at Yogo Gulch occur in a 5 km long dike, located on the northeast flank of the Little Belt Mountains, one of the foreland uplifts in Central Montana (Baker, 1991). In surface exposures the dike cuts Mississippian-age limestone of the Madison Group and siltstone and shale in the Big Snowy group (Zimmerman, 1966; Dahy, 1988, 1991). In the subsurface, Paleozoic sedimentary rocks overlie Proterozoic-age rocks—the base of the Belt Supergroup—deposited near the edge of the Helena embayment. The underlying Archean basement complex consists of high grade metamorphic rocks. The igneous host rock consists primarily of clinopyroxene, phlogopite, and analcime (Clabaugh, 1952; Dahy, 1988, 1991). It also contains about 4% titaniferous magnetite (Meyer and Mitchell, 1988), providing a magnetic signature which could be used in geophysical prospecting. It is an ultramafic lamprophyre called an ouachitite (Clabaugh, 1952, p. 14; Meyer and Mitchell, 1988; Brownlow and Komorowski, 1988; Dahy, 1988, 1991). The dike contains abundant globules or ocelli of carbonate that Dahy (1988, 1991) interpreted as evidence of immiscibility of a separate carbonatite liquid. The Yogo dike is part of the Central Montana Alkalic Province (CMAP), an igneous province characterized by mafic and ultramafic igneous rocks with unusually high alkali contents (dominantly potassium) that formed in the mantle far below the 50 to 55 km thick crust (Hearn, 1989; Baker and Berg, 1991, Baker, 1992).

Plate Tectonic Model for the CMAP. The onset of igneous activity during the Laramide orogeny, which formed the Northern Rocky Mountains, advanced from southwest to the northeast across Idaho and Montana during Late Cretaceous and Paleocene time, about 80-58 million years (m.y.) ago, as a result of shallowing of the subducted Farallon plate (Dickinson, 1979; Bird, 1984, 1988; O'Brien *et al.*, 1991; Baker, 1992). The great Idaho batholith and the smaller Boulder batholith were emplaced to the west of the CMAP during this period. Sizeable gold deposits were emplaced in the Moccasin, Judith and Little Rocky Mountains (Baker and Berg, 1991). The short lapse of volcanic activity (or magma gap) during the late Paleocene (about 58-54 m.y.), when the subducted Farallon plate scraped the underside of Montana lithosphere, was when the greatest crustal shortening and major uplift of the Little Belt, Big Snowy, Beartooth and other foreland ranges along deep-seated thrust faults in the basement occurred. This was followed by an igneous flare-up across Montana, Wyoming and Idaho during the Eocene (about 54-48 m.y.), corresponding to the decoupling and sinking of the Farallon plate (Baker, 1992). Major Eocene mineralization included the lead-zinc-silver and sapphire deposits in the Little Belt Mountains (Baker and Berg, 1991) (and sapphire-bearing volcanics at Rock Creek).

Magma Genesis. As the Farallon plate sank under central Montana in the Early Eocene (about 50 million years ago), it released water stored as hydroxyl in phlogopite mica. This, in turn, triggered partial melting of asthenosphere (O'Brien *et al.*, 1991) at a depth of approximately 150 to 200 km. As the melt rose through the mantle lithosphere, which is about 100 km. thick under the Little Belt Mountains (Eggler and Furlong, 1991), it assimilated low-melting material, particularly phlogopite-rich veins (O'Brien *et al.*, 1991, Fig. 12; Irving *et al.*, 1991). Neodymium and lead isotopes acquired from this assimilated material give (inherited) Precambrian radiometric ages, even though the rocks were emplaced in the Eocene (Dudás *et al.*, 1987; O'Brien *et al.*, 1991). [Of course, potassium-argon dates of Laramide igneous rocks in the Little Belt Mountains give Eocene ages (Marvin *et al.*, 1973).]

Magma Density and Buoyancy. Mafic and ultramafic magmas rise to the base of the crust where they tend to "pool" because the density contrast between magma and country rock is less in the crust than in the mantle (Philpotts, 1990, p. 466). The magma density, estimated from the chemical analysis given by Clabaugh (1952) using the method of Bottinga and Weill (1970), was 2.6 g/cm^3 compared with 2.9 - 3.0 for lower crustal rocks.

"Pooling" at or near the base of the crust allowed the magma to cool by heating and locally melting crustal rocks. Dahy (1988) mapped a rhyolite sill and a rhyolite-cored laccolithic dome only 1.3 and 2.3 km from the Yogo dike, respectively. These granitic magmas were generated from the heat supplied by mantle-derived mafic and ultramafic magmas "pooling" in the lower crust. Embry (1987) described how a mantle-derived shonkinite magma formed the Yogo Peak stock at the head of Yogo Gulch (*cf.* Baker *et al.*, 1991), but before it could solidify, a granitic magma derived from the lower crust rose up through the center of the shonkinite intrusion and caused physical mixing of the two magmas.

Sapphire Genesis. Pirsson (1897; 1900, p. 554) and Clabaugh (1952, p. 57) surmised that Yogo sapphires formed when the ultramafic lamprophyre magma assimilated some metamorphosed shale. However, they did not envision the great depth at which this occurred.

Kyanite-Bearing Xenoliths from the Lower Crust. The kyanite-quartz xenolith in the Yogo dike described by Clabaugh (1952, p. 16) and the kyanite-garnet-quartz xenolith with accessory rutile described by Dahy (1991) are the kind of mineral assemblages expected for a metamorphosed shale near the base of the crust where the conditions were 550-600°C (Eggler *et al.*, 1988, Fig. 5), 50-55 km depth (Prodehl and Lipman, 1989, Figs. 2 and 24; Braile *et al.*, 1989, Fig. 3), and approximately 12 kilobars or 1.2 gigapascals of pressure. At these temperatures, pressures above 10 kilobars correspond to the eclogite metamorphic facies, characterized by absence of hydrous phases such as muscovite (Philpotts, 1990, p. 328). A long history of burial and metamorphism allowed the water to escape. A plausible original mineralogy is a relatively common rock type—shale composed of kaolinite and quartz ± iron oxides. Hall (1987, p. 271-275) described the process of assimilation of metapelites by basic magmas. Silica is removed from aluminosilicates (in this case kyanite) to form corundum (*cf.* Helmley *et al.*, 1980) and from garnet to form spinel. I suspect that a comprehensive search for xenoliths in the Yogo dike will find some with corundum replacing kyanite, similar to the description of Altherr *et al.* (1982) of gem-quality corundum replacing kyanite in gneiss undergoing anatexis in Tanzania.

Sapphire-Bearing Cognate Xenoliths. Heating country rock to generate granitic magmas and assimilating some pelitic rock lowered the magma temperature, resulting in crystallization of clinopyroxene. Dahy (1988, 1991) described the bright green clots in the Yogo dike consisting mostly of clinopyroxene, but also with phlogopite and sapphires, which were (presumably formed as cumulates and) subsequently brought up with the magma as cognate xenoliths according to Meyer and Mitchell (1988) aluminum coordination in the clinopyroxene phenocrysts in the Yogo dike indicates low pressure (i.e. lower crust) rather than high pressure (i.e. mantle) crystallization. They suggest a crystallization temperature for phlogopite in the magma of 900°C.

Analcime Inclusions in Sapphire. The rare small inclusions in Yogo sapphires noted by Gübelin and Koivula (1986) contain analcime, pyrite, calcite, rutile, zircon, and a dark mica that is most likely phlogopite. The occurrence of white "snowballs" of analcime is unique among Montana sapphires. Roux and Hamilton (1976) determined experimentally that primary igneous analcime has a very restricted stability range of 8-14 kilobars and about 600°-640°C. These pressures correspond to the lower part of the abnormally thick crust under central Montana (Bird, 1984), but the temperatures require some additional heating by magma from the mantle. Since analcime is not a metamorphic mineral, its presence as inclusions in the sapphires is additional evidence that the sapphires formed in the Eocene by assimilation reactions rather than as metamorphic corundum in the Precambrian that had been liberated from matrix in the Eocene. The barrel-shaped crystal habit, typical of metamorphic corundum, has not been observed in the Yogo deposit.

Carbon-Dioxide Inclusions in Sapphire. Roedder (1972, Plate 6; 1984, p. 491-492) described two-phase fluid inclusions in a cut 10.2 carat (ct) Yogo sapphire. The inclusions contain liquid carbon-dioxide and gas bubbles that homogenize at 31°C, indicating a composition of nearly pure CO_2. Fluid inclusions of nearly pure CO_2 in a variety of minerals are well-known from deeply buried, high-grade metamorphosed crustal rocks in the eclogite and granulite metamorphic facies (*cf.* Roedder, 1984, p. 361-380), from mantle-derived basaltic magmas (Roedder, 1984, p. 488-492), and in xenoliths of mantle rocks in alkali basalts (Roedder, 1984, p. 514-532). Both Koivula (1986) and Schmetzer and Medenbach (1988) found nearly pure CO_2 fluid inclusions in blue sapphire from Sri Lanka in the hypersthene-granulite metamorphic facies.

Separate globules or ocelli of carbonate in the Yogo lamprophyre (Dahy, 1988, 1991) and fluid inclusions of CO_2 in Yogo sapphire are direct petrographic evidence for liquid immiscibility in the mantle-derived Yogo magma. According to Roedder (1984, p. 383-384, 504-505) when such unmixing occurs, water remains in the silicate melt. Thus, the incompatible requirements of high water activity in order to form primary igneous analcime inclusions in some sapphires and high CO_2 activity to form the fluid inclusions in other sapphires are explained by separate liquids—a water-rich silicate magma with small blebs of carbonate magma and/or CO_2.

Ascent Through the Crust. The buoyancy imparted by the volatiles to the Yogo magma caused it to rise through the crust. As the magma ascended through the crust, it decompressed and cooled, generating many changes in the chemistry of crystallizing phases. Clinopyroxenes became progressively enriched in iron, titanium, aluminum and some acquired a rim of acmite (Meyer and Mitchell, 1988). Titaniferous magnetite grains and sapphire crystals were partially resorbed by the magma, leaving in the latter case reaction rims of dark-green spinel (=hercynite) (Clabaugh, 1952, p. 18; Dahy, 1988; Meyer and Mitchell, 1988; Brownlow and Komoroski, 1988). Many of the sapphire crystals had grown as thin plates (basal pinacoids) (*cf.* Berezhkova, 1980; Hughes, 1990, p. 149), but developed rhombohedral faces (and etch pits) by dissolution during ascent of the magma (*cf.* Berezhkova, 1980; Pratt, 1897; Clabaugh, 1952, p. 18-21). Nepheline formed hexagonal prisms about 1½ cm long and 1 cm in diameter (Weed, 1900, p. 457; Clabaugh, 1952, p. 16; Dahy, 1991), but these are now pseudomorphs consisting of fine-grained calcite and quartz (L.G. Zeihen, personal communication, 1994). The size of the nepheline crystals appears to rule out extremely rapid ascent of the magma, such as occurred in Montana diatremes (Hearn and McGee, 1984).

The fluid inclusion of CO_2, 0.75 mnn in diameter, in the Yogo sapphire specimen at the Smithsonian Institution was too large to survive the trip to the surface (*cf.* Roedder, 1984, p. 70-75). High pressure CO_2 exceeded the tensile strength of the sapphire, causing rupture and escape of CO_2 along the fracture. However, the fracture healed, and the CO_2 coalesced into many small inclusions with a "fingerprint" appearance (Roedder, 1972, Plate 6; 1984, p. 492). Pressure inside the inclusions at 31°C is now only 70 bars.

Other Montana Sapphire Deposits

The other large sapphire deposits in Montana—Missouri River near Helena, Dry Cottonwood Creek, and Rock Creek—are all placer deposits with volcanics or shallow intrusives as inferred source rocks (Clabaugh, 1952). Sapphire in these deposits differs significantly from Yogo sapphire. The crystals show oscillatory color zoning and contain abundant inclusions, particularly rutile. The second order prism $\{11\bar{2}0\}$, absent at Yogo, is common. The iron content is typically 5 to 10 times greater than at Yogo (Emmett and Douthit, 1994), indicating far more oxidizing conditions. The colors are pastel with a much greater range of hues than at Yogo. Greens and yellows, characteristic of ferric iron, are well represented. These features are consistent with growth at shallower depths in the crust under more oxidizing conditions and subsequent transport to the surface as xenocrysts.

Missouri River near Helena. Although no kyanite-bearing metamorphic rocks crop out in the Helena area, kyanite is found together with sapphire at El Dorado Bar of the Missouri River near Helena, suggesting that the kyanite and corundum may have been brought up in the same magma. Mertie *et al.* (1951), Clabaugh (1952), and Zeihen (1987) mention a sapphire-bearing andesite dike at Canyon Ferry Dam near the upriver limit for alluvial sapphires along the Missouri River, but conclude that a far larger source is needed to account for the volume of sapphire in the placer deposits down river. The dike and a nearby pluton appear to be associated with the Late Cretaceous to Early Paleocene Boulder batholith and Elkhorn Mountains volcanics (Mertie *et al.*, 1951). The sapphire- and gold-bearing placer deposits in the strath terraces along the Missouri River contain pebbles of volcanic rocks other than andesite, indicating other candidates for source rocks. Quaternary faulting altered drainage patterns in the Helena Valley, including that of Prickly Pear Creek (Stickney, 1987). Thus, Eocene-age Lowland Creek volcanics in the Prickly Pear Valley, 30 km to the southwest near Clancy (*cf.* Wallace *et al.*, 1986) should be included in the list of possible source rocks for the placer deposits.

Dry Cottonwood Creek. The placer deposit at Dry Cottonwood Creek is underlain by Tertiary-age rhyolitic tuff and the Boulder batholith. The source rocks suggested by Clabaugh (1952) are the Eocene-age Lowland Creek volcanics at the head of the drainage (Wallace *et al.*, 1986). Gravels in the placer deposit are composed of rhyolitic tuff, andesite and basalt (American Gem, 1994).

Rock Creek. Clabaugh (1952) discussed the occurrence of sapphire-bearing pebbles of andesite in the Rock Creek deposit. There are also pebbles of rhyolite tuff and basalt (American Gem, 1994). Wallace *et al.* (1986) mapped volcanics with andesites, latites and rhyolites in the area around the placer deposit—the presumed source rock for the sapphires. These Eocene (?) volcanics postdate the nearby Idaho batholith (LaTour, 1974; Wallace *et al.*, 1986).

Source Rock. All three deposits are adjacent to a batholith and to large volumes of volcanic deposits. Thus they occur where large volumes of crustal rocks were undergoing anatexis. A cross section of crust, lithospheric mantle and asthenosphere shows systematic changes from Yogo to Rock Creek (Fig. 4). The geothermal gradient is much higher in the west than in the east. The lithosphere thins abruptly at the Rocky Mountain front, coinciding with a transition from thick-skinned to thin-skinned tectonics. (Plate tectonic models for the "keel" in the mantle lithosphere under central and eastern Montana are discussed by Baker, 1992.)

Sapphire Genesis. Although corundum is stable under a wide range of conditions (*cf.* Berman, 1988), one can make a good case for gem-quality corundum forming in the temperature range of 600° to 750°C. Suggested pressures are in the range of 6 to 14 kb. These conditions are close to or somewhat above the melting curve for granite. Altheer *et al.* (1981) found that gem-quality corundum formed in Tanzania at 695°C and 7.7 kb from kyanite during anatexis of gneiss. Okrush *et al.* (1976) determined that gem-quality corundum in Kashmir formed at 620°C and 6 kb. Blue sapphire from cordiorite- and/or sillimanite-bearing granulites in Sri Lanka formed within this range of conditions (*cf.* Keller, 1990, p. 9-10; Philpotts, 1990, p. 328, 358). Blue sapphire at the Thurein Taung mine in Myanmar (formerly Burma) occurs at the contact between nepheline-rich urtites and regionally metamorphosed marble and gneiss (Kane and Kammerling, 1992); and the eutectic for nepheline-rich rocks is within this range (Philpotts, 1990, p. 201). The analcime inclusions in Yogo sapphire have a restricted stability of 600° to 640°C. The stability range of corundum with respect to pressure and temperature can be explored for the mineral assemblages at known sapphire localities using several commercially available programs for the personal computer and collections of thermodynamic data for rock-forming minerals, such as that of Berman (1988). However, this is beyond the scope of the present paper.

When the temperature "window" of 600° - 700°C for sapphire genesis is plotted on the cross section of the lithosphere (diagonal hatching in Fig. 4), the difference in source regions between Yogo and the other deposits is obvious. Under the three western deposits this temperature interval occurs in the middle third of the crust at depths of approximately 25 to 35 km where rocks with ferric rather than ferrous iron are abundant and the rocks are more oxidized. Higher oxygen fugacity meant far greater amounts of iron were dissolved in the sapphire. Under Yogo the temperature "window" extends from the base of the crust down into the mantle. However, the sapphire was generated in the crust, not the mantle, because the kyanite-bearing xenoliths—the presumed source of aluminium for the sapphire—are rich in quartz, a mineral absent in the mantle. For quartz-rich rocks to exist at such depths in the lower crust (the stippled region in Fig. 4) without melting required an abnormally low geothermal gradient. The low geothermal gradient was a consequence of the abnormally thick lithospheric mantle. Mantle-derived ultramafic magma "pooling" in the lower crust supplied the necessary heat to create the sapphire.

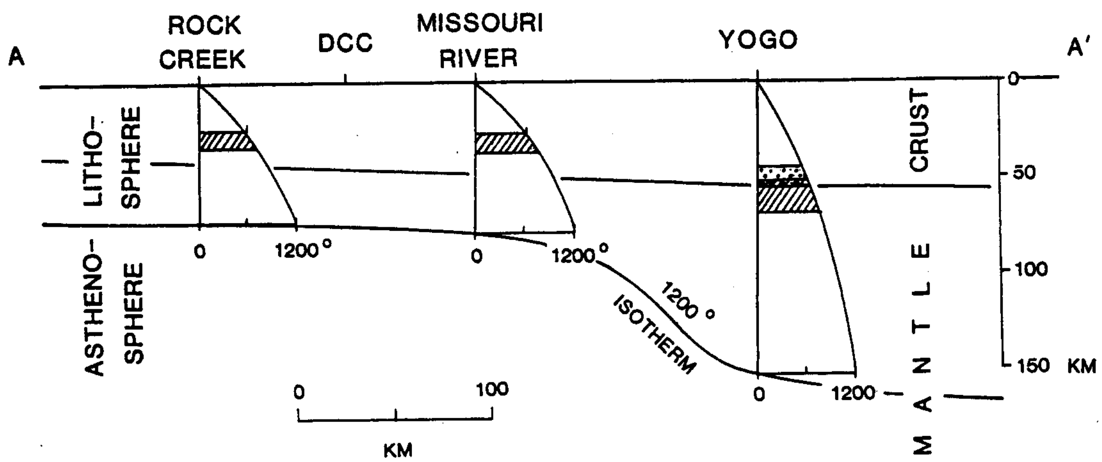


Figure 4. Cross section A-A' in Fig. 1, showing crust and lithospheric mantle, and temperature profiles at Rock Creek, Missouri River by Helena and Yogo. DCC - Dry Cottonwood Creek crustal thickness from Prodehl and Lipman (1989). Lithosphere thickness from Eggler and Furlong (1991) and Iyer and Hitchcock (1989). Base of lithosphere is assumed to be 1200°C after Eggler and Furlong (1991). Diagonal hatching - suggested thermal window of 600°-750° for sapphire generation. Stippling - suggested source for Yogo sapphire.

Heat Treating. Emmett and Douthit (1994) demonstrated that a significant percentage of Rock Creek sapphire can be heat-treated to produce blue or yellow sapphire. Heat-treating changes the valence of iron. Heating sapphire in an atmosphere with low oxygen fugacity, reduces as much iron as can be matched pairwise with titanium, compensating the electrical charges. However, this leaves substantial ferric iron in the lattice which cannot be removed without over-reduction. Overreduction causes particles of hercynite or metallic iron to precipitate, adding an undesirable gray component to the color. As shown in Fig. 3, the ferrous-ferric iron transitions centered at 875 nm strongly absorb the red end of the spectrum of transmitted light, eliminating much of the violet tint and producing "steely" blues rather than cornflower blues (*cf.* Schmetzer, 1987; Schmetzer and Kiefert, 1990).

Thus, it appears that the massive production of heat-treated sapphire by American Gem Corporation in Helena will produce very substantial quantities of "pure" blue or steel blue sapphire geared to mass marketing at lower prices, whereas the cornflower blue of the natural, untreated Yogo sapphire will preserve its up-scale market for those preferring the violetish-blue that gives a greater single-channel response in the human visual system.

THE IMPACT OF MINING COSTS

Mining the three placer deposits involves surface operations with relatively low mining costs and high volume. In contrast underground mining at Yogo is much more expensive. A number of mining ventures at Yogo have failed because of the inability to contain mining costs (Smith, 1964; Barron 1982; Voynick, 1987). Perhaps the most significant geological factor has been the myth that the Yogo dike is a simple tabular body. The magma intruded into a mature karst system developed in limestone of the Madison Group. Hydrothermal alteration transformed phlogopite and clinopyroxene into chlorite, cemented the cave collapse breccia matrix with carbonates, quartz, and pyrite, and formed pyrite cubes (Dahy, 1988). Circulating ground water continued karst development after the intrusion and chemically weathered the lamprophyre. The problems associated with karst are best illustrated at the west end of the dike.

The Kunisaki Tunnel

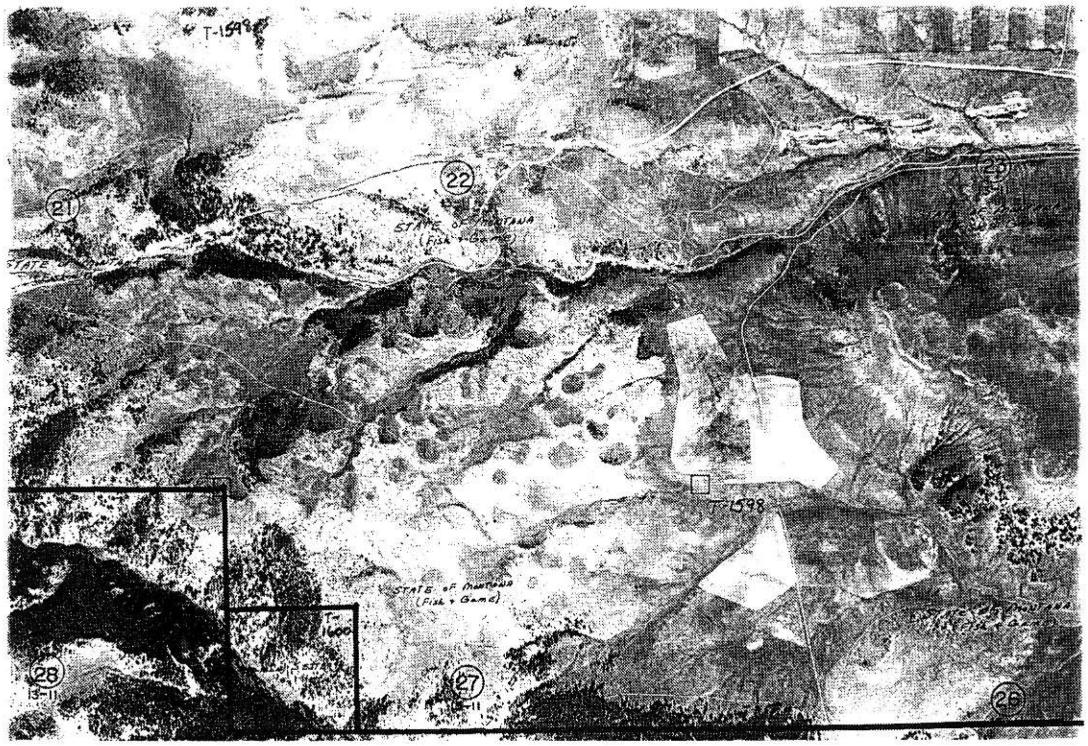
In 1972 the owner of the Yogo Mine, Chikara Kunisaki, invested $5 million to drive a 1 km long tunnel the length of the western-most of the three segments that form the Yogo dike. His intent was to mine a vertical tabular dike; however, he found geological complexity (Voynick, 1987). Aerial photos show that where the limestone forming the top of the Madison Group is exposed on the surface, the ground is "pock-marked" with collapsed sinkholes (Fig. 5), developed shortly after the limestone was deposited (*cf.* Sando and Dutro, 1979; Sando, 1988). The Yogo magma intruded into an extensive, but collapsed cave system, filled with break-down breccia (*cf.* Baker *et al.*, 1991, Fig. 57.8). As in other segments of the dike, the limestone breccia acted as a filter, trapping sapphire xenocrysts as the magma filled all cavities. The sapphire content of some pre-dike breccia exceeded the richest dike rock (Voynick, 1987, p. 158). Throughout the American-Kunisaki Mine much of the igneous rock has been altered hydrothermally and chemically weathered to a soft, crumbly mass, easily mined and washed for sapphires. Continued karst activity since the Eocene created more caverns that eventually collapsed, and collapsed the deeply weathered lamprophyre. Dahy (1988) described pieces from the overlying Kibbey and Otter formations, which had fallen down 100 m from above, and are now in the eastern end of the Kunisaki Tunnel. Considerable money and effort was spent trying to follow the main dike in areas where it simply did not exist. It had been transformed into an irregular mass of angular limestone blocks and some weathered igneous clasts embedded in a matrix of silt (washed down into the cave system from above) and clay (altered from the intrusion). In some places the only evidence of the intrusion was the presence of montmorillonite in the matrix of the breccia. The current operator, Cyprus Amax Minerals Co., is the first company in the last 50 years with the experience and resources to develop this kind of deposit. So far they have avoided the west segment of the dike.

Vortex

The small Vortex Mine has features similar to those in the nearby American-Kunisaki Mine, but is perhaps even more chaotic. A 60 m shaft and mine workings expose solid limestone, limestone with (sapphire-bearing) clay-filled joints, marble, shattered marble, open cavities (caves), breccia with clasts of limestone (and locally marble fragments) in a matrix of calcite and montmorillonite clay, and a small body of reddish-purple altered and chemically-weathered igneous rock rich in sapphire (Mychaluk, 1992; McCulloch, 1993). The breccia matrix typically contains only 3 to 5 cts/ton, whereas the "pod" of the weathered lamprophyre has a concentration of 70 cts/ton (Mychaluk, 1992). The reddish-purple ore is composed primarily of montmorillonite and calcite, with lessor amounts of phlogopite, chlorite, and iron oxides (Mychaluk, 1992).

"Exotic" clasts in the breccia include soft, bright red shale from the overlying Kibbey formation containing kaolinite, illite, and quartz in the fine-grained fraction (R.B. Berg, personal communication, 1994) and porphyry (Mychaluk, 1992). The Kibbey shale presumably fell down a sinkhole from above, whereas the porphyry was presumably carried up from an underlying intrusion. The latter may have been the distal portion of the nearby Sawmill Gulch laccolith, which was emplaced in the Cambrian-age Flathead sandstone (Dahy, 1988). Slickensides in clay-rich material are common. Some of the clay-rich zones show deformation textures as if the material had been sheared. The transitions from one kind of rock to the next are abrupt, consistent

(a)



(b)

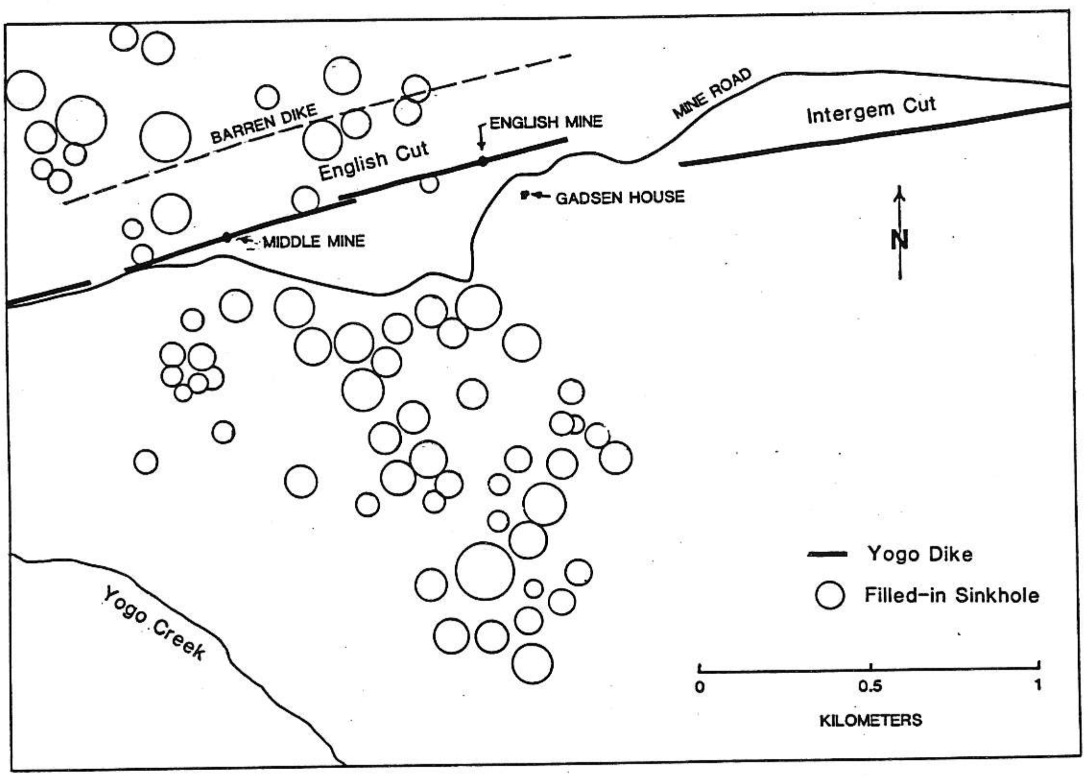


Figure 5. Aerial photo (U.S. Agricultural Stablization and Conservation Service) of the English Cut and Intergem Cut segments of the Yogo dike, showing filled-in sinkholes in the top of the limestone in the Madison group. a. Photo. b. Interpretation.

with a mechanical juxtaposition of very different rock types.

Loss of water to the Madison limestone, determined during spring run-off in 1964 by measuring stream flow upstream and downstream from the Vortex Mine, was 1.2 m^3 per second (Feltis, 1980). Features in the mine are most easily explained by collapse of a post-intrusion, dominantly vertical cave system. Dahy (1988) interpreted a small structure in Kelly Coulee, 200 m north of the Vortex Mine, as a diatreme and more recently suggested that the Vortex Mine could be a breccia pipe (Dahy, personal communication, 1993). The alternatives of cold ground water dissolution of limestone versus hot, explosive gas as the causative agent for the extensive breccia will, hopefully, be evaluated by mapping the three-dimensional structure of the deposit.

Production, Reserves and Values

Some estimates of past production and remaining reserves are listed in Table 1. The wholesale price in Montana of a fine 1 carat Yogo sapphire is currently approximately $2000 (Roncor, Inc., personal communication, 1993). Heat-treated sapphires from the placer deposits will be competing with heat-treated sapphire from Sri Lanka. The current wholesale price of a fine 1 carat Sri Lankan blue sapphire is approximately $500 (Michelsen, 1994). The higher price of the Yogos reflects higher mining costs and the persistent demand for the cornflower blue hue.

ACKNOWLEDGEMENTS

The manuscript was improved by comments of R.B. Berg. Research for this article was greatly aided by the On-Line Search services of the Montana State Library and the Lewis and Clark Library in Helena. The Inter-Library Loan services of Kathy Mora at the Great Falls Public Library and Jean Fontana and Verna Nicola at the Belt Public Library are gratefully acknowledged.

Table 1. Estimates of production and reserves of Montana sapphire rough (except where otherwise noted).

Production:		
Yogo	20 million carats	Clabaugh (1952), Brown (1982)
Missouri River	> 7.7 mill. cts.	American Gem (1994)
Dry Cottonwood Creek	> 400,000 cts.	Clabaugh (1952)
Rock Creek	190 mill. cts.	Emmett and Douthit (1994)
Reserves:		
Yogo	100 mill. cts. 800,000 cts. (finished gemstones)	Brown (1982) Voynick (1993)
Missouri River	(not available)	
Dry Cottonwood Creek	1.5 - 5 mill. cts.	American Gem (1994)
Rock Creek	> 25 mill.cts.	Emmett and Douthit (1994)

REFERENCES

Altherr, R., Okrusch, M., and Bank, H., 1981, Corundum- and kyanite-bearing anatexites from the Precambrian of Tanzania: Lithos, v. 15, pp. 191-197.

American Gem Corporation, 1994, Annual Report 1993, 1994; Management Proxy Circular: Helena.

Asen, S. and Horowitz, R.M., 1974, Apigenin 4'-0-β-D-glucoside 7-0-β-D-glucuronide: the copigment in the blue pigment of *Centaurea cyanus*: Phytochemistry, v. 13, pp. 1219-1223.

Asen, S. and Jurd, L., 1967, The constitution of a crystalline, blue cornflower pigment: Phytochemistry, v. 6, pp. 577-584.

Baker, D.W., 1991, Laramide tectonics and magmatism in the Central Montana alkalic province: *in* Baker D.W. and Berg R.B., eds., Guidebook of the Central Montana Alkalic Province, Montana Bureau of Mines and Geology Special Publication 100, pp. 128-131.

Baker, D.W., 1992, Central Montana alkalic province: critical review of Laramide plate tectonic models that extract alkalic magmas from abnormally thick Precambrian lithospheric mantle: Northwest Geology, v. 20/21, pp. 71-95.

Baker, D.W. and Berg, R.B., 1991, Guidebook of the Central Montana Alkalic Province: Montana Bureau of Mines and Geology Special Publication 100, 201 p.

Baker, D.W., McBride, G., and Dahy, J.P., 1991, Field Guide: Little Belt Mountains: *in* Baker, D.W. and Berg R.B., eds., Guidebook of the Central Montana Alkalic Province, Montana Bureau of Mines and Geology Special Publication 100, pp. 144-162.

Barron, A.A., 1982, The Yogo sapphire: *in* Eash, D.M., ed., International Gemological Symposium Proceedings, Gemological Institute of America, Santa Monica, pp. 343-347.

Bayer, E., 1958, Über den blauen Farbstoff der Kornblume: Chemische Berichte, v. 91, pp. 1115-1121.

Bayer, E., Nether, K., and Egeter, H., 1960, Synthese der blauen, im Kornblumenfarbstoff enthaltenen Chelate: Chemische Berichte, v. 93, pp. 2871-2879.

Berman, R.G., 1988, Internally consistent thermodynamic data for minerals in the system Na_2O-K_2O-CaO-MgO-FeO-Fe_2O_3-Al_2O_3-SiO_2-TiO_2-H_2O-CO_2: Journal of Petrology, v. 29, pp. 445-522.

Berezhkova, G.V., 1980, Study of anisotropy of growth rate and dissolution rate of crystals: *in* Belyaev, L.M., ed., Ruby and Sapphire, Amerind Publishing Co., New Delhi, pp. 11-15.

Billmeyer, F.W. and Saltzman, M., 1981, Principles of Color Technology, Wiley, New York, 240 p.

Bird, P., 1984, Laramide crustal thickening event in the Rocky Mountain foreland and Great Basin: Tectonics, v. 3, pp. 741-758.

Bird, P., 1988, Formation of the Rocky Mountains, western United States—a continuum model: Science, v. 239, pp. 1501-1507.

Bottinga, Y., and Weill, D.F., 1970, Densities of liquid silicate systems calculated from partial molar volumes of oxide components: American Journal of Science, v. 269, pp. 169-182.

Boynton, R.M., 1979, Human Color Vision: Holt, Rinehart & Winston, New York, 438 p.

Braile, L.W., Hinze, W.J., von Frese, R.R.B., and Keller, G.R., 1989, Seismic properties of the crust and uppermost mantle of the conterminous United States and adjacent Canada: *in* Pakiser, L.C., and Mooney, W.D., eds., Geophysical Framework of the Continental United States: Geological Society of America Memoir 172, pp. 655-680.

Brown, D., 1982, Geology of the Yogo Sapphire Deposit, Judith Basin County, Montana: Unpublished report.

Brownlow, A.H. and Komorowski, J.C., 1988, Geology and origin of the Yogo sapphire deposit, Montana: Economic Geology, v. 83, pp. 875-880.

Clabaugh, S.E., 1952, Corundum deposits of Montana: U.S. Geological Survey Bulletin 983, 100 p.

Dahy, J.P., 1988, The geology and igneous rocks of the Yogo sapphire deposit and surrounding area, Little Belt Mountains, Judith Basin Co., Montana: [M.S. Thesis] Montana College of Mineral Science and Technology, Butte, 92 p.

Dahy, J.P., 1991, Geology and igneous rocks of the Yogo sapphire deposit, Little Belt Mountains, Montana: *in* Baker, D.W. and Berg, R.B., eds., Guidebook of the Central Montana Alkalic Province, Montana Bureau of Mines and Geology Special Publication 100, pp. 45-54.

Dickinson, W.R., 1979, Cenozoic plate tectonic setting of the Cordilleran region in the United States: *in* Armentrout, J.M., Cole, M.R., and Terbest, H., eds., Cenozoic Paleogeography of the Western United States, Pacific Section, Society of Economic Paleontologists and Mineralogists, Pacific Coast Paleogeography Symposium 3, pp. 1-13.

Dudás, F.Ö., Carlson, R.W., and Eggler, D.H., 1987, Regional Middle Proterozoic enrichment of the subcontinental mantle source of igneous rocks from central Montana: Geology, v. 15, pp. 22-25.

Eggler, D.H., Meen, J.K., Welt, F., Dudás, F.Ö., Furlong, K.P., McCallum, M.E., and Carlson, R.W., 1988, Tectonomagmatism of the Wyoming Province: Colorado School of Mines Quarterly, v. 83, pp. 25-40.

Eggler, D.H. and Furlong, K.P., 1991, Petrochemical and geophysical evidence for old mantle lithosphere beneath Montana: *in* Baker, D.W. and Berg, R.B., eds., Guidebook of the Central Montana Alkalic Province, Montana Bureau of Mines and

Geology Special Publication 100, pp. 86-91.

Embry, P.A., 1987, Petrogenesis of the Yogo Peak Stock, Little Belt Mountains, Montana: [M.. Thesis] University of Montana, Missoula, 84 p.

Emmett, J.L. and Douthit, T.R., 1994, Heat treating the sapphires of Rock Creek, Montana: Gems & Gemology, v. 29, pp. 250272.

Feltis, R.D., 1980, Water resources of Judith Basin, central Montana: Montana Bureau of Mines and Geology, Hydrogeologic Map 1.

Gübelin, E.J. and Koivula, J.I., 1986, Photoatlas of Gemstone Inclusions: ABC Edition, Zurich.

Hall, A., 1987, Igneous Petrology: Longman Scientific & Technical, Essex, England, 573 p.

Hearn, B.C., 1989, Montana high-potassium igneous province: *in* Volcanism and Plutonism of Western North America, 28th Int'l Geological Congress Field Trip Guidebook T346, American Geophysical Union, pp. 1-6.

Hearn, B.C. and McGee, E.S., 1984, Garnet peridotites from Williams kimberlites, North-Central Montana, U.S.A.: *in* Kornprobst, J., ed., Kimberlites II: The Mantle and Crust-Mantle Relationships, Elsevier, Amsterdam, pp. 57-70.

Helmley, J.J., Marinenko, J.W., and Luce, R.W., 1980, Equilibria in the system Al_2O_3-SiO_2-H_2O and some general implications for alteration/mineralization processes: Economic Geology, v. 75, pp. 210-228.

Hughes, R.W., 1990, Corundum: Butterworth-Heineman, London, 314 p.

Irving, A.J. and O'Brien, H.E., 1991, Isotopic and trace-element characteristics of Late Cretaceous to Oligocene mafic volcanics from West-Central Montana: Implications for Tertiary tectonics and magma genesis: *in* Baker D.W. and Berg R.B. eds., Guidebook of the Central Montana Alkalic Province, Montana Bureau of Mines and Geology Special Publication 100, pp. 140-142.

Iyer, H.M. and Hitchcock, T., 1989, Upper-mantle velocity structure in the continental U.S. and Canada: *in* Pakiser L.C. and Mooney W.D., eds., Geophysical Framework of the Continental United States, Geological Society of America Memoir 172, pp. 681-710.

Kane, R.E. and Kammerling, R.C., 1992, Status of ruby and sapphire mining in the Mogok Stone Tract: Gems & Gemology, v. 28, pp. 152-174.

Keller, P.C., 1990, Gemstones and their Origins: Van Nostrand Reinhold, New York, 144 p.

Kelly, K.L. and Judd, D.B., 1976, COLOR: Universal Language and Dictionary of Names: National Bureau of Standards Special Publication 440.

Koivula, J.I., 1986, Carbon dioxide fluid inclusions as proof of natural-colored corundum: Gems & Gemology, v. 22, pp. 152-155.

LaTour, T.E., 1974, An examination of metamorphism and scapolite in the Skalhaho region, southern Sapphire Range, Montana: [M.S. Thesis] University of Montana, Missoula, 95 p.

Marvin, R.F., Witkind, I.J., Keefer, W.R., and Mehnert, H.H., 1973, Radiometric ages of intrusive rocks in the Little Belt Mountains, Montana: Geological Society of America Bulletin, v. 84, p. 1977-1986.

McCulloch, R., 1993, Montana mining activity: Oral presentation at Northwest Mining Association Meeting, Spokane.

Mertie, J.B., Fischer, R.P., and Hobbs S.W., 1951, Geology of the Canyon Ferry Quadrangle, Montana: U.S. Geological Survey Bulletin 972, 97 p.

Meyer, H.O.A. and Mitchell, R.H., 1988, Sapphire-bearing ultramafic lamprophyre from Yogo, Montana: a ouachitite: Canadian Mineralogist, v. 25, pp. 81-88.

Meyers, C.E., Mason, T.O., Petuskey, W.T., Halloran, J.W., and Bowen, H.K., 1980, Phase equilibria in the system Fe-Al-O: Journal of the American Ceramic Society, v. 63, pp.659-663.

Michelsen, S., 1994, Michelsen Gemstone Index: Center for the Study of Gemstone Evaluation, Bayonne, N.J.

Mychaluk, K., 1992, Geology of the Vortex Sapphire Mine, Utica, Montana: [B.S. Thesis] University of Calgary, Calgary, Alberta, 97 p.

Nassau, K., 1983, The Physics and Chemistry of Color: Wiley, New York, 454 p.

Newman, Renée, 1994, The Ruby & Sapphire Buying Guide: How to Spot and Avoid Ripoffs: International Jewelry Publications, Los Angeles, 2nd ed., 204 p.

O'Brien, H.E., Irving, A.J., and McCallum, I.S., 1991, Eocene potassic magmatism in the Highwood Mountains, Montana—petrology, geochemistry, and tectonic implications: Journal of Geophysical Research, v. 96, pp. 13,237-13,260.

Okrush, M., Bunch, T.E., and Bank, H., 1976, Paragenesis and petrogenesis of a corundum-bearing marble at Hunza (Kashmir): Mineralium Deposita, v. 11, pp. 278-297.

Philpotts, A.R., 1990, Principles of Igneous and Metamorphic Petrology: Prentice Hall, Englewood Cliffs, N.J., 498 p.

Pirsson, L.V., 1897, On the corundum-bearing rock from Yogo Gulch, Montana: American Journal of Science, 4th Series, v. 4, pp. 421-423.

Pirsson, L.V., 1900, Petrography of the igneous rocks of the Little Belt Mountains, Montana: U.S. Geological Survey 20th

Annual Report, Part 3, pp. 463-595.

Pratt, J.H., 1897, On the crystallography of the Montana sapphires: American Journal of Science, 4th Series, v. 4, pp. 424-428.

Prodehl, C. and Lipman, P.W., 1989, Crustal structure of the Rocky Mountain region: *in* Pakiser L.C. and Mooney W.D., eds., Geophysical Framework of the Continental United States, Geological Society of America Memoir 172, pp. 249-289.

Roberts, K., 1991, Gem Set Color Communication System: International Color Standard, East Greenwich, R.I.

Roedder, E., 1972, Composition of fluid inclusions: U.S. Geological Survey Professional Paper 440-JJ, 164 p.

Roedder, E., 1984, Fluid Inclusions: Reviews in Mineralogy, v. 12, Mineralogical Society of America, 644 p.

Roux, J. and Hamilton, D.L., 1976, Primary igneous analcite--an experimental study: Journal of Petrology, v. 17, pp. 244-257.

Rubin, H., 1992, Gem Dialogue Systems, Inc., 205-02 26th Ave. Bayside, NY.

Sando, W.J., 1988, Madison limestone (Mississippian) paleokarst: a geologic synthesis: *in* James, N.P. and Choquette, P.W., eds., Paleokarst, Springer Verlag, pp. 256-277.

Sando, W.J. and Dutro, J.T., 1979, Little Belt Mountains, Montana (road log): *in* Gilmour, E.H. and Dutro, J.T., eds., Carboniferous of the northern Rocky Mountains, Field Trip No. 15, 9th International Congress of Carboniferous Stratigraphy and Geology, American Geological Institute, pp. 31-37.

Schmetzer, K., 1987, Zur Deutung der Farbursache blauer Saphire—eine Diskussion. [The cause of color in blue sapphire—a discussion]: Neues Jahrbuch für Mineralogie Monatshefte, v. 1987, pp. 337-343.

Schmetzer, K. and Bank, H., 1980, Explanations of the absorption spectra of natural and synthetic Fe- and Ti-containing corundum: Neues Jahrbuch für Mineralogie Abhandlungen, v. 139, pp. 216-224.

Schmetzer, K., and Bank, H., 1981, The colour of natural corundum: Neues Jahrbuch für Mineralogie Monatshefte, v. 1981, pp. 59-68.

Schmetzer K. and Kiefert L., 1990, Spectroscopic evidence for heat treatment of blue sapphires from Sri Lanka--additional data: Journal of Gemology, v. 22, pp. 80-82.

Schmetzer K. and Medenbach O., 1988, Examination of three-phase inclusions in colorless, yellow, and blue sapphires from Sri Lanka: Gems & Gemology, v. 23, pp. 107-111.

Smith, L.E., 1964, Independent study on the commercial sapphire deposits of Yogo, Judith Basin County, and Rock Creek, Granite County, Montana: [M.S. Thesis] Eastern Michigan University, Ypsilanti, 211 p.

Stickney, M.C., 1987, Quaternary geology and seismicity of the Helena Valley, Montana: *in* Berg, R.B. and Breuninger, R.H., eds., Guidebook of the Helena Area, West-Central Montana, Montana Bureau of Mines and Geology Special Publication 95, pp. 17-26, 49-52.

Suwa, Y., 1993, Gemstone Quality and Value: Sekai-Bunka-Sha (in Japanese with English translation by Gemological Institute of America, Santa Monica, CA).

Themelis, T., 1992, The Heat Treatment of Ruby and Sapphire: Gemlab, Inc., Clearwater, FL, 235 p.

Townsend, M.G., 1968, Visible charge transfer band in blue sapphire: Solid State Communication, v. 6, pp. 81-83.

Voynick, S.M., 1987, Yogo: The Great American Sapphire: Mountain Press, Missoula, Montana, 2nd ed., 207 p.

Voynick, S.M., 1993, Montana sapphires: Compressed Air Magazine, Ingersoll-Rand Corp., December issue, pp. 18-25.

Wallace, C.A., Schmidt, R.G., Lidke, D.J., Waters, M.R., Elliott, J.E., Frendh, A.B., Whipple, J.W., Zarske, S.E., Blaskowski, M.J., Heise, B.A., Yeoman, R.A., O'Neill, J.M., Lopez, D.A., Robinson, G.D., and Klepper, M.R., 1986, Preliminary geologic map of the Butte 1° x 2° quadrangle, western Montana: U.S. Geological Survey, Open File Report OF-86-292.

Weed, W.H., 1900, Geology of the Little Belt Mountains, Montana: U.S. Geological Survey 20th Annual Report, Part 3, pp. 257-461.

Wyszecki, G. and Stiles, W.S., 1982, Color Science: Wiley, New York, 923 p.

Zeihen, L.G., 1987, The sapphire deposits of Montana: Montana Bureau of Mines and Geology Bulletin 126, pp. 28-40.

Zimmerman, E.A., 1966, Geology and ground water resources of western and southern parts of Judith Basin, Montana: Montana Bureau of Mines and Geology Bulletin 50-A, 33 p.

DIAMOND POTENTIAL IN SOUTHERN BRITISH COLUMBIA

RICHARD T. WALKER - Dynamic Exploration Ltd., 1916 - 5th Street South, Cranbrook, B.C. V1C 1K4

The possibility of diamonds contained in the alkaline igneous lithologies of southern British Columbia is generally considered to be low. The rationale for this opinion is most probably based on application of Clifford's Rule, namely that the occurrence of diamondiferous kimberlites will be restricted to rigid cratonic areas greater than 2,000 Ma (Late Proterozoic to Archean) in age. A.J.A.. Janse describes a broader definition of craton as "...composed of coherent, relatively immobile and generally low grade metamorphic blocks ... of any Precambrian, but generally Archean, age, which are surrounded by younger, generally Proterozoic, strongly deformed and high grade metamorphic mobile belts". The Canadian Cordilleran is comprised of a Phanerozoic mobile belt (defined as a long, narrow crustal region characterized by present or past tectonic activity) immediately adjacent to Proterozoic to Archean age Canadian craton. As the Cordillera is a Phanerozoic mobile belt, it is not considered to be prospective for diamonds.

However, the eastern portion of the Cordillera (Rocky and Purcell Mountains) is characterized by thin-skinned tectonics, in which sediments of the miogeocline have been thrust up and onto North American basement. Lithoprobe deep seismic studies document the presence of North American continental crust westward as "...a regionally extensive, west-facing transition (on the west side of the Monashee Mountains) from thick craton on the east to thin transitional, basinal or oceanic crust on the west..." (Cook *et al.* 1991). Furthermore, based on geochemistry, Armstrong and others interpret the presence of highly attenuated North American continental crust out as far west as the Fraser River (Varsek, 1994, pers. comm.). The interpretation of rigid, North American continental crust extending west to the western edge of the Monashee Mountains is sufficient for the arguments presented in this discussion.

Recent studies in the Alberta Basin (Villeneuve *et al.* 1993, Ross *et al.* 1991) have correlated exposures of Canadian Shield in Saskatchewan into the sub-surface of the Alberta Basin using aeromagnetic data. Correlations were confirmed through geochronological dating of monazite and zircon separates recovered from 90 basement penetrating drill-holes in the Alberta Basin (Ross *et al.* 1991). The domains thus defined can be traced with certainty from Saskatchewan through the Alberta Basin to the eastern edge of the Purcell Mountains (west of the Rocky Mountain Trench). Furthermore, the southernmost portion of the Alberta Basin is underlain by Archean basement correlated to the Hearn Province of the Canadian Shield (Wyoming Craton in the United States). Ages determined for this domain range from 1.779 (Late Proterozoic) to 3.278 (Archean) billion years (Villeneuve *et al.* 1993, Ross *et al.* 1991). This predominantly Archean domain underlies the Rocky Mountain Alkaline Belt (RMAB), from the International Border to the area of the Columbia Icefields northeast of Golden.

There are a wide variety of mafic to ultramafic, alkaline occurrences in the Rocky and Purcell Mountains, ranging from kimberlite to lamprophyre and including olivine melilitites, lamproites, kimberlites, alkaline to basaltic lamprophyres, alnoites, aillikites. There is only one widely acknowledged kimberlite occurring in the RMAB, the Cross Kimberlite in Crossing Creek. The Joff occurrence has been defined by separate authors as an alkaline lamprophyre (olivine melilitite) (Pell 1987) or a hypabyssal facies phlogopite-bearing kimberlite (Fipke *et al.* 1989).

The Cross kimberlite has an emplacement age between 240 and 250 Ma (Grieve 1982, Smith *et al.* 1988). Rb-Sr dating on phlogopite separates from the HP pipe (ultramafic lamprophyre) northeast of Golden returned dates of 348±7 Ma and 391±12 Ma and a K-Ar date of 396±10Ma. A Rb-Sr phlogopite age for the Larry occurrence northeast of Golden returned an age of 334±7 and 348±7 Ma (MINFILE 083C 001). A kimberlite in the Purcell Mountains returned a Rb-Sr phlogopite-apatite mineral pair age of 245±2.4 Ma (Pope and Thirlwall 1991). These emplacement ages are supported on the basis of stratigraphic relationships. Helmstaedt *et al.* (1988) interpret most of the diatremes in the RMAB to have been intruded prior to deposition of Devonian strata. Several of the diatremes are in sharp contact with the basal Devonian unconformity, indicating deposition prior to the unconformity. At least three distinct intrusive events have been interpreted for the RMAB, namely during the Ordovician, Ordoviclan-Silurian and Permian-Triassic (Helmstaedt *et al.* 1988). The emplacement ages determined to date support Permian-Triassic alkaline intrusive activity, before displacement associated with the Late Cretaceous - Early Tertiary deformation.

Parrish and Reichenbach (1991) sampled seven of the pipes in the southern Rocky Mountains (Jack, Mark, Mike, HP, Cross, Blackfoot and Joff) in an attempt to determine the age of emplacement of these diatremes. U-Pb dating of zircon separates failed to establish a magmatic age of intrusion for the diatremes, however a wide range of dates were obtained ranging from Archean (2.7-2.6 Ga) to Paleozoic (440 Ma). Possible sources proposed for xenocrystic zircons include gneisses of the western Canadian Shield, basement beneath the Alberta Basin, basement rocks exposed in the Cordillera and intrusives associated with orogeny and/or rifting.

Diatremes of the southern Cordillera were apparently intruded predominantly in the Paleozoic and contain zircon xenocrysts of Archean to early Paleozoic age. Crustal-type xenoliths are documented from many of the occurrences (i.e. Blackfoot, Summer, Cross, and HP) as well as mantle-derived xenoliths including peridotites, chromitites, pyroxenites, hornblendites and eclogites. In addition to abundant indicator minerals, at least 10 micro-diamonds have been reported and at least three recovered from the Jack and Mark occurrences northeast of Golden (BC Assessment Report 15289). Dia Met Minerals Ltd. further reports recovery of two macro-diamonds from each of the Mark and Jack occurrences (BC Assessment Reports 15289 and 16195).

In summary, diamonds have been recovered from two diatremes located in a Paleozoic mobile belt (Jack and Mark claims northeast of Golden). The mobile belt consists of thrust sheets tectonically emplaced upon composite Archean and Proterozoic basement, interpreted based on several complimentary lines of evidence. Ross *et al.* (1991) have correlated gneisses of the Canadian Shield to basement gneisses of the Alberta Basin using aeromagnetic data while others (Armstrong *et al.* 1991, McDonough and Parrish 1991, Murphy *et al.* 1991, Parkinson 1991) have proposed a Canadian Shield origin for basement gneiss exposures in the Cordillera. Therefore, it is a valid conclusion that Canadian Shield gneisses extend into the subsurface of the Alberta Basin and extend beneath the Cordillera to at least the western edge of the Monashee Mountains, where transitional crust is interpreted to occur.

Therefore, alkaline diatremes of the southern Cordillera overlie a composite block of the Archean Hearn Province. Zircons recovered from diatremes in the Rocky Mountains document ages similar to those recovered from basement in the Alberta Basin, including Archean ages. Many of these same diatremes contain mantle xenoliths (peridotitic and eclogitic) and indicator minerals, xenoliths and diamonds. It is therefore reasonable to expect the presence of diamonds in alkaline occurrences having mantle-derived peridotitic/eclogitic nodules and/or zircons of Archean age.

REFERENCES

Armstrong, R.L., Parrish, R.R., van der Hayden, P., Scott, K., Runkle, D. and Brown, R.L. 1991. Early Proterozoic basement exposures in the southern Canadian Cordillera: core gneiss of Frenchman's Cap, Unit I of the Grand Forks Gneiss, and the Vaseaux Formation. Canadian Journal of Earth Science, vol. 28, pp. 1169-1201.

Cook, F.A., Varsek, J.L. and Clarke, E.A. 1991. Proterozoic craton to basin craton transition in western Canada and its influence on the evolution of the Cordillera, Canadian Journal of Earth Science, vol. 28, pp. 1148-1158.

Grieve, D.A. 1985. 1980 - Petrology and chemistry of the Cross kimberlite (82J/2). British Columbia Ministry of Energy, Mines and Petroleum Resources Geology in British Columbia, 1977-1981, pp. 34-41.

Helmstaedt, H.H., Mott, J.A., Hall, D.C., Schulze, D.J. and Dixon, J.M. 1988. Stratigraphic and structural setting of intrusive breccia diatremes in the White River - Bull River area, southeastern British Columbia. in British Columbia Ministry of Energy, Mines and Petroleum Resources Geological Fieldwork 1987, Paper 1988-1, pp. 363-368.

Janse, A.J.A. 1984. Kimberlites - where and when. In J.E. Glover and P.G. Harris. eds., Kimberlite Occurrence and Origin: A Basis for Conceptual Models in Exploration, Department of Geology and University Extension, University of western Australia, Publication No. 8, pp. 19-61.

McDonough, M.R. and Parrish, R.R. 1991. Proterozoic gneisses of the Malton Complex, near Valemount, British Columbia: U-Pb ages and Nd isotopic signatures, Canadian Journal of Earth Science, vol. 28, pp. 1202-1216.

Murphy, D.C., Walker, R.T. and Parrish, R.R. 1991. Age and setting of the Gold Creek gneiss, crystalline basement of the Windermere Supergroup, Cariboo Mountains, B.C., Canadian Journal of Earth Science, vol. 28, pp. 1217-1231.

Parkinson, D. 1991. Age and isotopic character of Early Proterozoic basement gneisses in the southern Monashee Complex, southeastern British Columbia, Canadian Journal of Earth Science, vol. 28, pp. 1159-1238.

Parrish, R.R. and Reichenbach, I. 1991. Age of xenocrystic zircon from diatremes of western Canada, Canadian Journal of Earth Science, vol. 28, pp. 1232-1238.

Pell, J. 1987. Alkaline Ultrabasic Rocks in British Columbia: Carbonatites, Nepheline Syenites, Kimberlites, Ultramafic Lamprophyres and Related Rocks. British Columbia Ministry of Energy, Mines and Petroleum Resources Open File 1987-17, 109 p.

Pope, A.J. and Thirlwall, M.F. 1991. Tectonic setting, age, and regional correlation of ultrabasic-ultrapotassic dykes in the northern Purcell Mountains, southeast British Columbia, Canadian Journal of Earth Science, vol. 29, pp. 523-530.

Ross, G.M., Parrish, R.R., Villeneuve, M.E. and Bowring, S.A. 1991. Geophysics and geochronology of the crystalline basement of the Alberta Basin, western Canada, Canadian Journal of Earth Science, vol. 28, pp. 512-522.

Smith, C.B., Colgan, E.A., Hawthorne, J.B. and Hutchinson, G. 1988. Emplacement age of the Cross kimberlite, southeastern British Columbia, by the Rb-Sr phlogopite method, Canadian Journal of Earth Science, vol. 25, pp. 790-792.

Villeneuve, M.E., Ross, G.M., Theriault, R.J., Miles, W., Parrish, R.R. and Broome, J. 1993. Tectonic subdivision and U-Pb geochronology of the crystalline basement of the Alberta Basin, western Canada, Geological Survey of Canada Bulletin 447, 86p.

THRUST ROTATION OF THE BELT BASIN, CANADA AND UNITED STATES

JAMES W. SEARS - Department of Geology, University of Montana, Missoula, MT 59812

ABSTRACT

The Purcell anticlinorium represents a structurally inverted part of the Belt basin in the northern Rocky Mountain thrust system. Marginal facies from the northeast side of the basin now occupy the northeast limb of the anticlinorium. Prior to Cretaceous thrusting, the marginal facies are proposed to have occupied a position southwest of the Lewis and Clark line, which now overlies the ramp that formed essentially at the base of the marginal sequence. Restoration of the marginal facies to their proposed southwest-facing footwall cutoff along the Lewis and Clark line indicates that the basin underwent 25-30 degrees of clockwise rotation about an Euler pole in west-central Montana. Emplacement of the Belt basin onto the North American craton was accommodated by horizontal shortening of adjacent shelf, platform, and foreland basin strata. Foreland basin deposits indicate that the thrust rotation began in Campanian and continued through Paleocene time.

INTRODUCTION

The Middle Proterozoic Belt Supergroup is a thick sedimentary deposit that is widely exposed in the Purcell anticlinorium of the northern Rocky Mountains (Fig. 1). Price (1981) showed that the anticlinorium represents the structurally inverted form of part of the original Belt depositional basin. The inversion resulted from thrust emplacement of the thick sedimentary fill of part of the basin onto the relatively flat surface of the adjacent North American craton (Price, 1981).

The purpose of this paper is to show that the Lewis and Clark line of Montana and Idaho marks the original location of the northeastern margin of the Belt basin, and thus provides a palinspastic guide for restoration of the Purcell anticlinorium. The restoration suggests that the anticlinorium underwent approximately 25 to 30 degrees of clockwise rotation during thrust emplacement.

STRUCTURALLY INVERTED BELT BASIN

The broad central area of the Purcell anticlinorium represents the Belt basin's inverted axial trough, where the Belt Supergroup is >15 km thick (Harrison, 1972). The lower division of the Belt Supergroup (Prichard Formation and equivalents) is >6 km thick in this area, and is dominated by basinal turbidites (Cressman, 1989; Hoy, 1993). Syndepositional mafic sills with a cumulative thickness of >1.5 km characterize the lower section (Turner *et al.*, 1993).

The wide, gently dipping, northeastern limb of the Purcell anticlinorium (Fig. 2) defines the structurally inverted northeastern margin of the Belt basin (Price, 1964; Aitken and McMechan, 1991). The lower division of the Belt Supergroup thins eastward across the northeastern limb to <1 km thick, and the facies change from basinal turbidites to platformal stromatolitic dolomites and trough-crossbedded quartz arenites (Price, 1964; Harrison, 1972; Mudge, 1970; Reynolds, 1984; Aitken and McMechan, 1991; Whipple *et al.*, 1984).

Northwest Geology, v. 23, p. 81-91, 1994

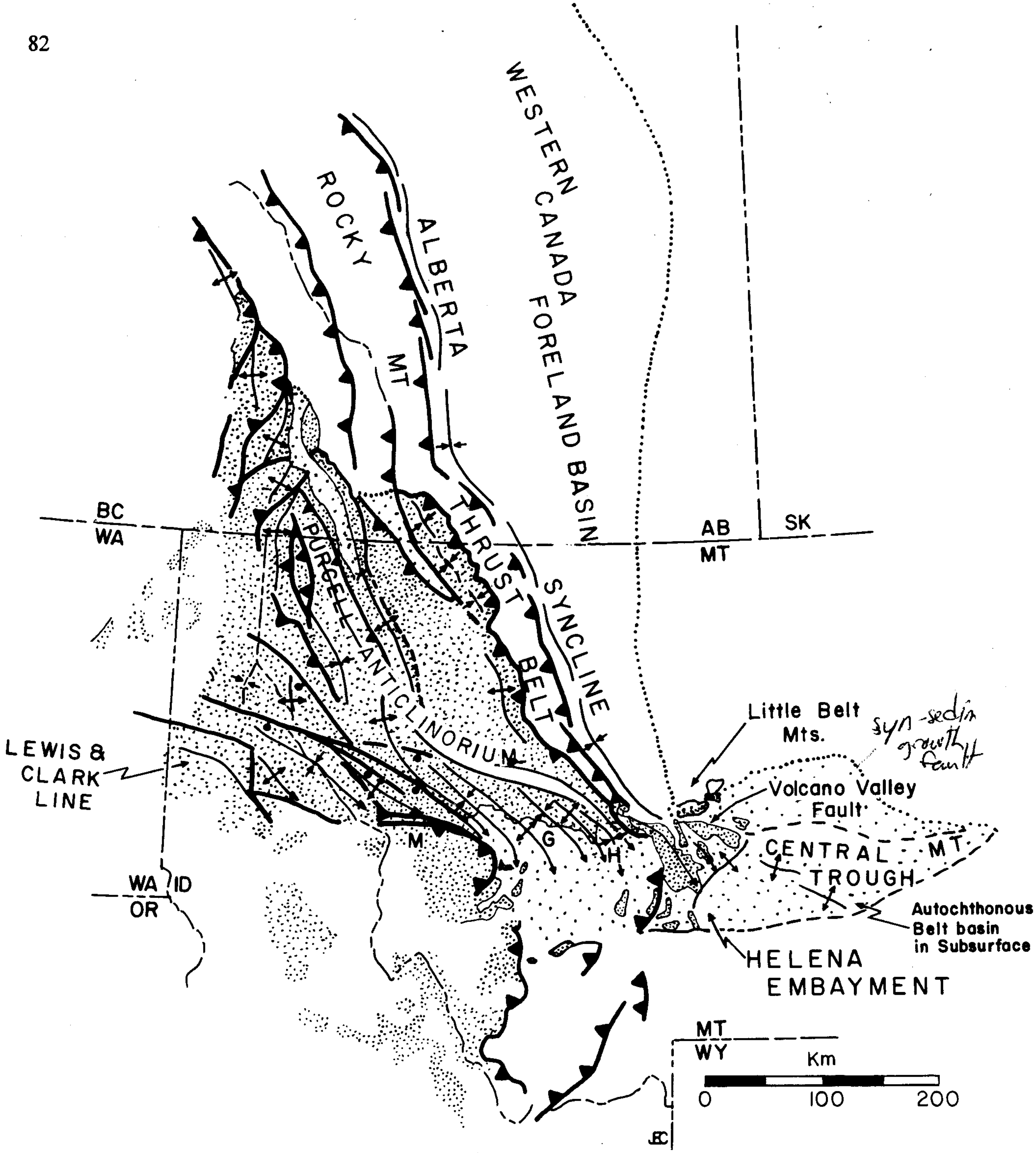
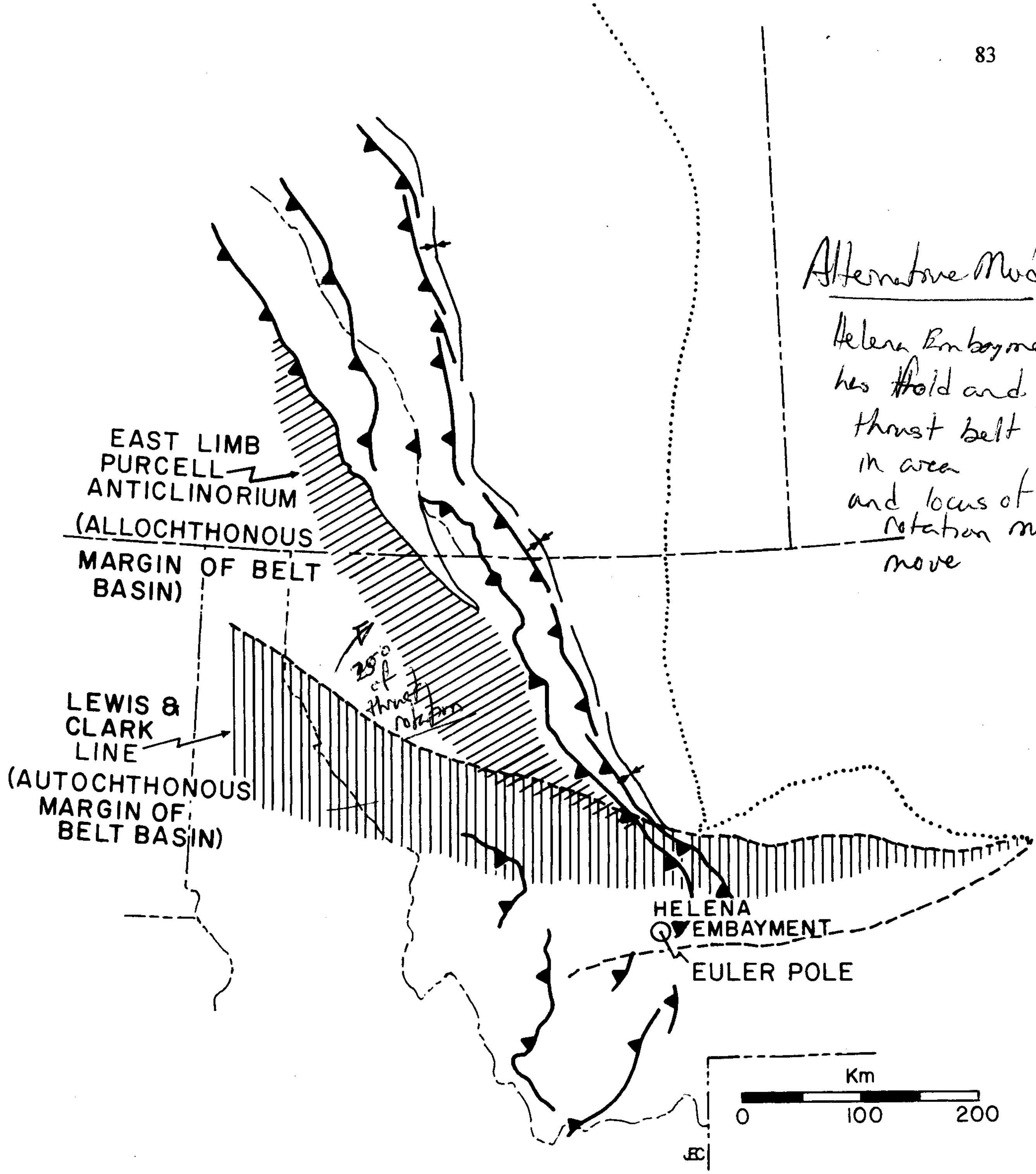


Figure 1. Tectonic map of Purcell anticlinorium and adjacent part of the Rocky Mountain thrust system and western Canada foreland basin. Simplified from Link *et al.* (1993), Wheeler and McFeely (1991), and Mallory (1972). Dense stipple, areas mostly underlain by Belt Supergroup; light stipple, Belt Supergroup in subsurface. M, Missoula; G, Garrison, H, Helena.

– related structures converge in Helena Embayment area
– increasing displacement of oldest margin of Belt-Purcell Basin to N



Alternative Model

Helena Embayment has fold and thrust belt in area and locus of rotation must move

Figure 2. Convergence of east limb of Purcell anticlinorium and Lewis and Clark line toward Helena embayment, interpreted to represent clockwise rotation of Purcell anticlinorium about Euler pole during tectonic inversion of Belt basin. Note convergence of Rocky Mountain thrust belt and western Canada foreland basin toward Euler pole.

East of Missoula, Montana, the sub-Cambrian unconformity provides a datum that further defines the geometry of the inverted basin. Thin Middle Cambrian platform strata unconformably overlap the Belt Supergroup across the Purcell anticlinorium (Mudge *et al.*, 1982; Wallace *et al.*, 1987). In the central area of the anticlinorium, the Cambrian overlies a >15 km thick section of Belt Supergroup, and rests on the uppermost Missoula Group (Harrison *et al.*, 1986). On the northeast limb, across a width of 50-100 km, the Cambrian overlaps successively older and thinner Belt units to the northeast (Mudge *et al.*, 1982); about 15 km of section are traversed across the limb. The Cambrian rests on autochthonous pre-Belt crystalline rocks beneath the foreland fold-thrust belt and foreland basin, as revealed by seismic and borehole data (Price and Mountjoy, 1970; Mudge *et al.*, 1982). Restoration of the Cambrian to a level datum plane requires restoration of the Purcell anticlinorium into a >15-km-deep depression.

AUTOCHTHONOUS BASIN MARGIN

A major, linear, structural depression that can be traced for hundreds of kms across Montana and Idaho appears to represent a deformed remnant of the original depositional basin of the Belt Supergroup. The eastern part of the depression is the central Montana trough, a long, narrow feature that crosses central Montana, evident in isopach maps of Phanerozoic stratigraphic units (Mallory, 1972). Seismic and borehole data show that Belt rocks occupy the trough at least as far east as Billings (Mallory, 1972).

The central part of the depression is the Helena embayment of the Belt basin (Harrison, 1972). In the Little Belt Mountains, Belt strata and pre-Belt crystalline basement rocks crop out beneath the Cambrian in a series of exposures that trend obliquely across a segment of the north margin of the Helena embayment. In the northern exposures, in the Little Belt Mountains, the Cambrian rests directly on pre-Belt crystalline basement (Weed, 1899). Progressively southward, the Cambrian overlaps a 1-km-thick platformal section of lower Belt rocks. The Belt rocks, in turn, rest unconformably on the crystalline basement. The platformal sequence is truncated on the south by the Volcano Valley fault, a WNW-trending, SW-dipping fault that had major normal displacement during Belt sedimentation, and minor oblique reverse displacement during formation of the Rocky Mountains (Godlewski and Zieg, 1984). South of the fault, the Cambrian overlies a thick, basinal section of Belt rocks. Rock types include bedded massive sulfide deposits, turbidites, and mafic sills. Mass-flow deposits contain stromatolitic limestone olistoliths, indicating a shallow-water source area across the Volcano Valley fault to the north (Zieg and Leitch, 1994). The Volcano Valley fault is interpreted as a growth fault with a few kilometers of displacement that formed during Belt deposition along the north margin of the Helena embayment (Zieg, 1986).

Thick basinal deposits like those of the southern Little Belt Mountains re-emerge in several anticlines and thrust plates farther west in the Helena embayment (Zieg, 1986). Along the southern margin of the Helena embayment, the lower Belt contains the 500 m thick LaHood conglomerate, a mass-flow deposit with olistoliths of crystalline basement rocks with distinctive lithologies that match rocks from the Archean terranes of southwest Montana (McMannis, 1963).

The Helena embayment plunges west beneath allochthonous rocks of the Cordillera, and is continuous with a deep synclinorium that crosses Montana and Idaho (Fig. 1). The synclinorium contains the Boulder batholith and its thick volcanic cover (Hamilton and Meyers, 1974), 6-km thick Cretaceous foreland basin deposits (Gwinn and Mutch, 1965), the Sapphire allochthon, and the Bitterroot lobe of the Idaho batholith (Hyndman, 1980).

The Lewis and Clark line forms the northern margin of the synclinorium. It is a broad structural zone trending approximately S70E from northern Idaho to west-central Montana. It is best known for its many major oblique-slip faults (Wallace *et al.*, 1990) and en echelon folds, but it is also a complex SW-facing homocline across which the structural level of the thrust system drops by >15 km. The homocline apparently overlies a major SW-facing ramp in the basal decollement beneath the fold-thrust belt. The ramp formed essentially at the base of the Belt marginal facies at the top of the crystalline basement and appears to be the counterpart to the inverted basin margin represented by the northeast limb of the Purcell anticlinorium (Sears, 1988).

The mean dip of bedding along the homocline is about 20-25 degrees toward the southwest (Sears, 1988). The mean width of the homocline is about 35-40 km. Geometric constructions thus suggest that the mean height of the homocline and the underlying thrust ramp is about 16+/-4 km.

The regional dip of the base of the Cambrian provides independent control on the thrust ramp geometry. The base of the Cambrian is folded over the Purcell anticlinorium, then dips down the homocline into the synclinorium. In the synclinorium near Garrison, the base of the Cambrian underlies 6 km of Mesozoic and 2 km of Paleozoic rocks (Gwinn and Mutch, 1965), and is thus 8 km beneath the surface. That depth corresponds to its projected regional position (Sears, 1988). Garrison overlies the lower flat.

The homocline along the Lewis and Clark line is also the locus of a set of dextral, oblique-slip faults with major down-to-the-south normal displacements (Wallace *et al.*, 1990). These Tertiary faults post-date thrusting and associated cleavage development, and appear to sole into the subsurface thrust ramp (Sears, 1988). They may provide a secondary guide to the location of the ramp.

RESTORATION OF PURCELL ANTI-LINORIUM

The above analysis suggests that the homocline along the Lewis and Clark line drapes a thrust ramp whose height, width and facing direction are consistent with the ramp forming along the floor of the northeastern limb of the Purcell anticlinorium. The homocline trends southeastward directly into the exposed margin of the Belt basin in the autochthonous Helena embayment. The Lewis and Clark line thus very likely overlies the pre-Cretaceous continuation of the autochthonous Belt basin margin across western Montana and Idaho. As such, it provides palinspastic control for restoration of part of the Belt basin; the map distance between the Lewis and Clark line and the east limb of the Purcell anticlinorium measures the displacement of the northeast margin of the Belt basin from its original site of deposition (Fig. 2). The basin margin and the Lewis and Clark line intersect near Helena, where marginal Belt rocks are exposed in the autochthonous Helena embayment.

Figure 3 restores the Purcell anticlinorium to the Lewis and Clark line, by approximately 25 degrees of rotation about an Euler pole south of Helena. Point A is fixed in the autochthonous part of the Helena embayment. Point B is independently constrained by a balanced cross-section that indicates approximately 250 km of horizontal displacement for the northern end of the Purcell anticlinorium (Price, 1981). The restored trace of the Purcell anticlinorium between Points A and B approximately coincides with the Lewis and Clark line. The restoration is consistent with other published cross-section lines between points A and B (Fig. 3).

Figure 4 shows that isopachs of syn-depositional mafic sills of the lower Belt Supergroup, and Prichard Formation member G turbidite sands, restored according to Fig. 3, define a WNW-plunging basin whose axis follows the base of the proposed ramp. The basin is covered on the west by the Cordilleran miogeocline.

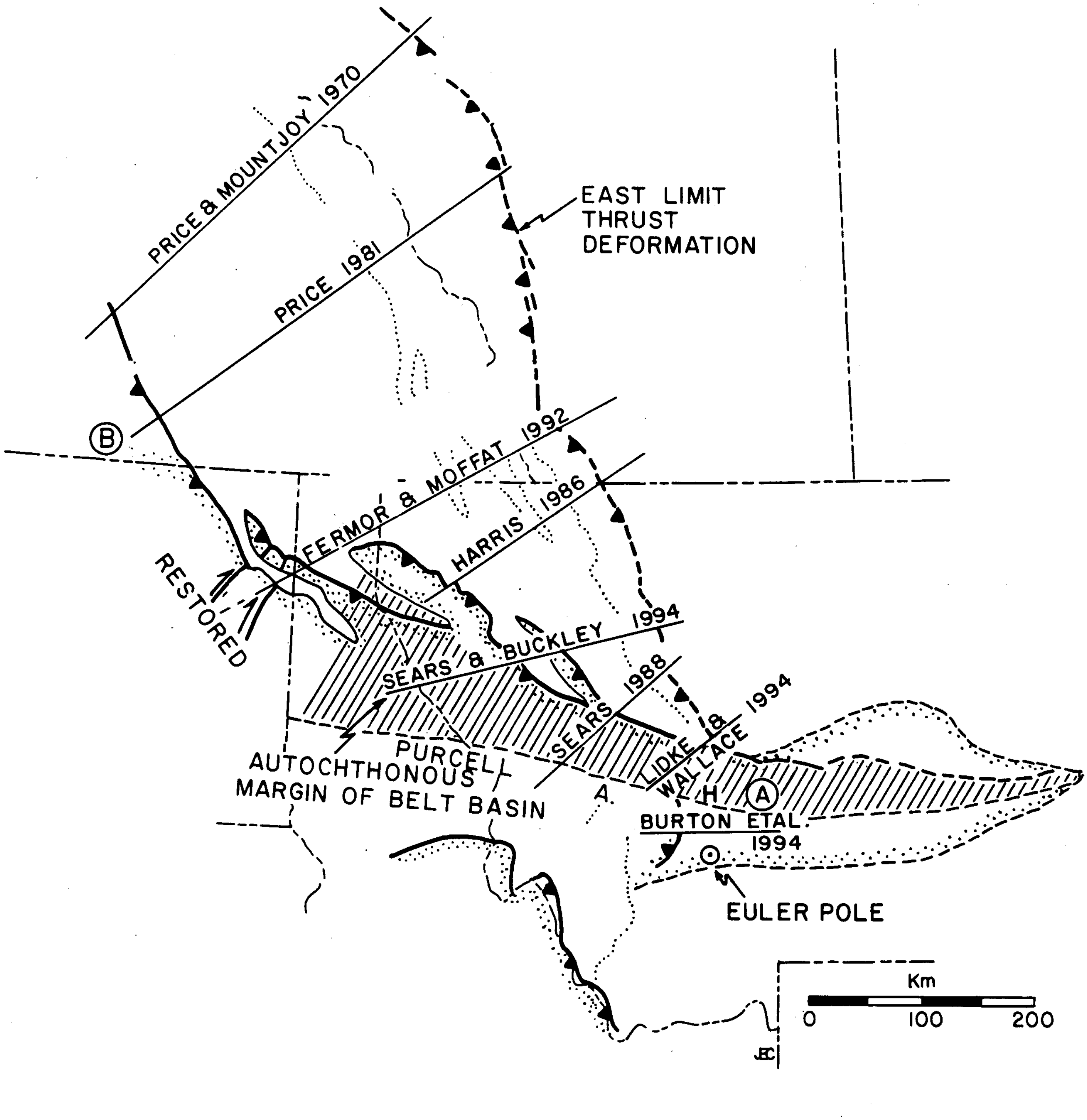


Figure 3. Purcell anticlinorium restored to Lewis and Clark line (hatchured zone). The loci of several balanced cross-section lines are shown for reference. They are consistent with the proposed restoration.

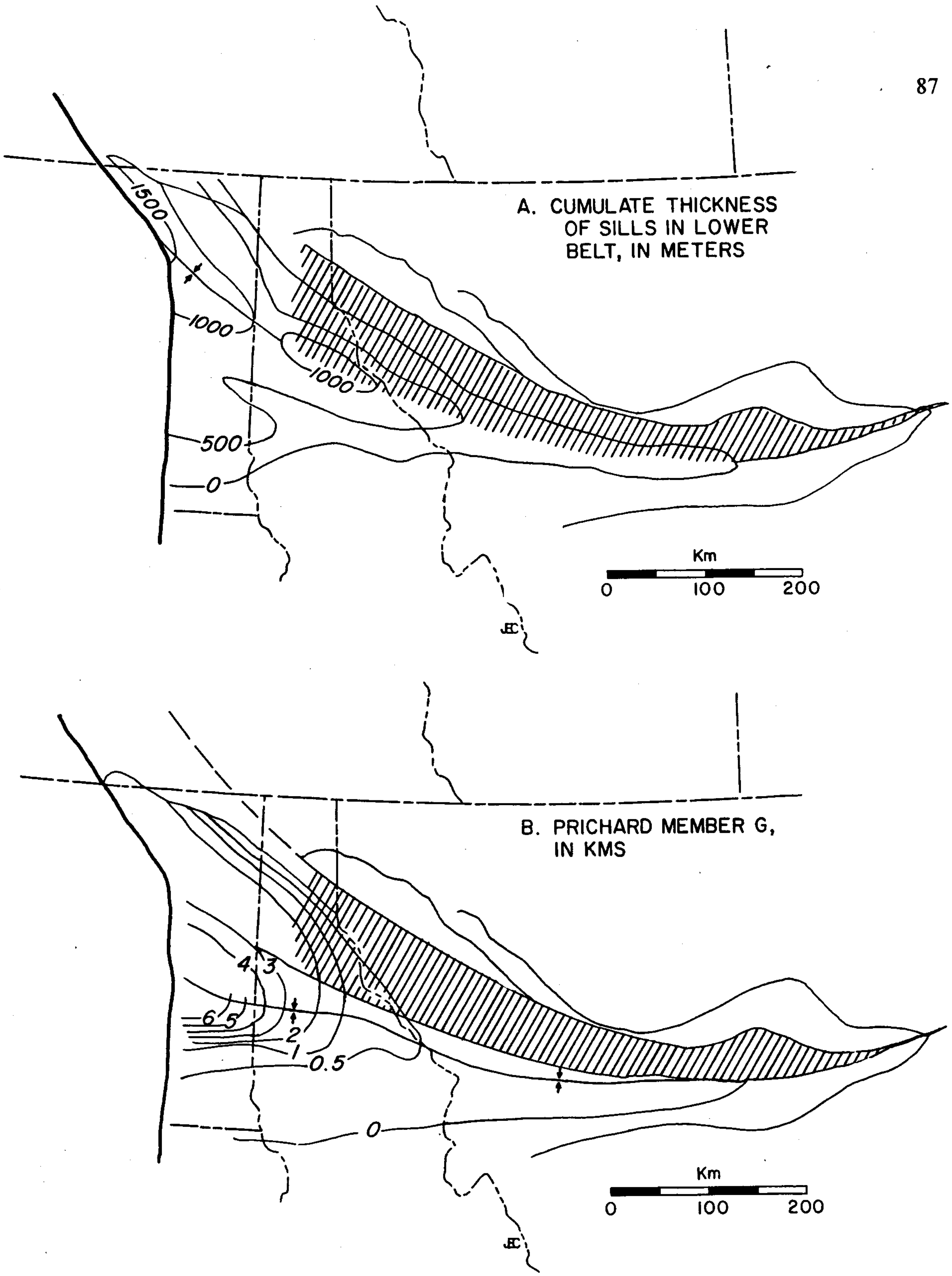


Figure 4. Isopachs and facies of selected units in Belt basin restored according to present model. A. Syndepositional mafic sills, composite thickness, modified after Turner, written communication, 1993; B. Prichard member G, Prichard quartzite member, middle Aldridge Formation and quartzite unit of Greyson Formation, modified after Cressman, 1989, Zieg, 1986.

ROCKY MOUNTAIN FORELAND

The proposed thrust rotation of the Belt basin is reflected in the decreasing width of the adjacent Rocky Mountain thrust belt and western Canada foreland basin toward the southeast (Fig. 1). The thrust belt on the east side of the Purcell anticlinorium is 150 km wide at the northern end of the anticlinorium, and tapers toward the proposed Euler pole. The width of the thrust belt is proportional to the cumulative horizontal displacement across the thrust belt, as reflected in restored cross sections (Fig. 3). A large part of the transport of the anticlinorium was accommodated by regular imbrication of the adjacent foreland; the anticlinorium acted as a master thrust plate above a growing imbricate fan (Boyer and Elliott, 1982). Greater displacement of the anticlinorium with increased radius from the Euler pole resulted in a matching increase in width of the thrust belt.

The Alberta syncline of the western Canada foreland basin also tapers toward the southeast. It is 225 km wide in the north, and narrows to zero near the proposed Euler pole, as defined by deformation of the base of the Fish Scale Zone, a Cretaceous marker bed (Leckie and Smith, 1992). The syncline plunges northwest at 6 to 8 m/km, as shown by depth to crystalline basement on a series of cross-sections along the western edge of the basin. The top of the crystalline basement is near sea level in the Little Belt Mountains east of Helena and deepens to about 5 km below sea level along Price and Mountjoy's 1970 section (Fig. 3). The deepening is mostly a function of thickening of the foreland basin deposits and is reflected in Late Cretaceous and Paleocene isopachs (Cook and Bally, 1975).

The narrowing and thinning of the Alberta syncline is consistent with decreased loading of the craton by thrusts as the displacement across the foreland fold-thrust belt decreases toward the Euler pole near Helena.

EVIDENCE FOR ROTATION

White (1978) determined horizontal movement vectors for the northern part of the Montana foreland thrust belt between Glacier National Park and Augusta, based on statistical analysis of folded bedding, thrust surfaces, and thrust striations. The vectors are approximately parallel to small circles about an Euler pole near Helena (see White, 1978, Fig. 7).

Little definitive paleomagnetic data that could test the rotation model exists for the Purcell anticlinorium. Apparent poles for the Belt Supergroup are of unproven age and have no independent reference on the craton (Irving and Wynne, 1991; Link *et al.*, 1992; O'Bradovich *et al.*, 1984). Data from five sites where Early Cretaceous Kootenai Formation overlies Belt Supergroup in Montana show clockwise rotations of 15 to 35 degrees, but one site records counterclockwise rotation of 29 degrees (Eldredge and Van der Voo, 1988).

TIMING OF EMPLACEMENT

Timing of emplacement of the Purcell anticlinorium is constrained by the history of the foreland basin. Deposits older than Campanian appear to cross the restored eastern Purcell anticlinorium without interruption, while Campanian and younger deposits reflect emergence of the anticlinorium.

In Late Jurassic through Early Cretaceous time (150-124 Ma), the western edge of the restored Belt basin was internally deformed, metamorphosed, and intruded as part of the Omineca crystalline belt (Archibald *et al.*, 1983). The area was uplifted and denuded of 8-10 km of cover before 122-98 Ma ago, when shallowly emplaced, undeformed granitic plutons cross-cut all penetrative fabrics and many mappable thrusts within the western part of what is now the Purcell anticlinorium (Hoy and Vanderheyden, 1988; Archibald *et al.*, 1983).

The crustal thickening in the Omineca crystalline belt coincided with subsidence of a wide foreland basin on the craton (Stockmal *et al.*, 1992). Foreland basin deposits of Late Jurassic and Early Cretaceous age thicken and coarsen toward the west, and record unroofing of the deformed western miogeocline and its granitic plutons. First clasts from the Omineca crystalline belt occur in late Oxfordian beds of the Fernie Formation, 152-144 Ma old (Leckie and Smith, 1992), and stratigraphically overlie the eastern part of the Belt Supergroup and its Paleozoic and early Mesozoic platform cover, without angular unconformity, in the Fernie basin (Price, 1962). Many rock units extend east from the Fernie basin unto the craton, and southeast across the restored Belt/Purcell basin. The Aptian-aged (114 Ma) Cadomin chert and quartzite pebble conglomerate of the Fernie basin extends as the basal Kootenai conglomerate of western Montana (Mallory, 1972). A distinctive set of lacustrine limestones with abundant ostracodes and gastropods extends from the middle Blairmore to the middle Kootenai units, crossing the restored Belt basin from the Canadian platform on the north to the Wyoming province on the south (see Mallory, 1972). These relationships indicate that the southeastern part of the Purcell anticlinorium formed after Early Cretaceous time.

No angular unconformities are known within the foreland basin deposits in the region under discussion, and teetonic inversion could not have occurred until at least after early Campanian; deposits of that age are cut by the Lewis thrust (Price and Mountjoy, 1970). In Montana, a set of Late Cretaceous andesite sills occur in Cretaceous strata on either side of the Purcell anticlinorium (Mudge *et al.*, 1982; Wallaee *et al.*, 1987). They range in age from 97 to 80 Ma (Wallace *et al.*, 1987), and may represent the youngest units that had continuity across the eastern part of the presently inverted basin. They intruded undisturbed strata and shared all subsequent deformation.

The tectonic inversion of the eastern part of the Belt basin appears to have occurred in Campanian through Paleocene time. The oldest known occurrences of Belt clasts in the foreland basin are in Campanian conglomerates (Gwinn and Mutch, 1965; Veile, 1960). While pre-Campanian strata appear to have had continuity across the presently inverted basin, major differences occur in the stratigraphic units of Campanian and younger ages between the northeastern and southwestern sides of the inverted basin (Wallace *et al.*, 1990). Noteably, the Campanian Two Medicine Formation, a broad alluvial plain deposit (Veile, 1960), occurs on the northeast, while the correlative Golden Spike Formation, a thick basinal deposit (Gwinn and Mutch, 1965), crops out on the southwest.

CONCLUSION

The basin-inversion model simplifies restoration of the Belt basin in northwestern Montana. Restoration of the Belt basin based entirely on rotating marginal facies counterclockwise back to a position along the Lewis and Clark line suggests displacement consistent with independent estimates of the minimum cumulative shortening in the adjacent thrust system. The model suggests that this part of the Belt basin experienced an average clockwise rotation of up to 25-30 degrees.

ACKNOWLEDGEMENTS

I thank Karl Kellogg for a careful review, and Ray Price for helpful discussions. The research was partially supported by a grant from the NASA JOVE program.

REFERENCES

Aitken, J.D., and McMechan, M.E., 1991. Middle Proterozoic assemblages, Chapter 5 in Geology of the Cordilleran orogen in Canada, H. Gabrielse and C.J. Yorath, eds.: Geological Survey of Canada, no. 4, p. 97-124.

Archibald, D.A., Glover, J.K., Price, R.A., Farrar, E., and Carmichael, D.M., 1983, Geochronology and tectonic implications of magmatism and metamorphism, southern

Kootenay arc and neighboring regions, southeastern British Columbia. Part I: Jurassic to mid-Cretaceous: Canadian Journal of Earth Sciences, v. 20, p. 1891-1913.

Boyer, S.E. and Elliott, D., 1982, Thrust systems: American Association of Petroleum Geologists, Bulletin v. 66, p. 1196-1230.

Burton, B.R., Ballard, D.W., Lageson, D.R., Perkins, M., Schmidt, C.J., Warne, J.R., 1994, Deep drilling results and new interpretations of the Lombard thrust, southwestern Montana: Belt Symposium III, Proceedings, Whitefish, Montana, Special Publication of the Montana Bureau of Mines and Geology, in press.

Cook, T.D., and Bally, A.W., 1975, Stratigraphic atlas of North and Central America: Princeton University Press, Princeton, NJ, 272 p.

Cressman, E.R., 1989, Reconnaissance stratigraphy of the Prichard Formation (middle Proterozoic) and the early development of the Belt Basin, Washington, Idaho, and Montana: U.S. Geological Survey Professional Paper 1490, 80 p.

Eldredge, S., and Van der Voo, R., 1988, Paleomagnetic study of thrust sheet rotations in the Helena and Wyoming salients of the northern Rocky Mountains: in Schmidt, C.J., and Perry, W.J., eds., Interaction of the Rocky Mountain foreland and the Cordilleran thrust belt: Geological Society of America, Memoir 171, Ch 11, p. 319-332.

Fermor, P.R., and Moffat, I.W., 1992, Tectonics and structure of the western Canada foreland basin, in Macqueen, R.W., and Leckie, D.A., eds., Foreland basins and fold belts: American Association of Petroleum Geologists Memoir 55, p. 81-105.

Godlewski, D.W. and Zieg, G.A., 1984, Stratigraphy and depositional seting of the Precambrian Newland Limestone: in Hobbs, S.W., ed., The Belt: Montana Bureau of Mines and Geology, Special Publication 90, p. 2-4.

Gwinn, V.E., and Mutch, T.A., 1965, Intertongued Upper Cretaceous volcaniuc and non-volcanic rocks, central-western Montana: Geological Society of America, Bulletin, v. 71, p. 1125-1144.

Hamilton, W., and Meyers, W.B., 1974, Nature of the boulder batholith of Montana: Geological Society of America Bulletin, v. 85, p. 365-378.

Harris, D.W., 1985, Crustal structure of northwestern Montana: MS thesis, University of Montana, Missoula, 63 p.

Harrison, J.E., 1972, Precambrian Belt basin of northwestern United States-Its geometry, sedimentation, and copper occurrences: Geological Society of America, Bulletin, v. 83, p. 1215-1240.

Harrison, J.E., Griggs, A.B., and Wells, J.D., 1986, Geologic and structure maps of the Wallace 1X2 degree Quadrangle, Montana and Idaho: U.S. Geological Survey, Miscellaneous Investigations Map I-1509, scale 1:250,000.

Hoy, T., 1993, Geology of the Purcell Supergroup in the Fernie West-half map-area, southeastern British Columbia: BC Min. Energy, Mines and Petroleum, Bull. 84, 157 p.

Hoy, T., and Vanderheyden, P., 1988, Geochemistry, geochronology and tectonic implications of two quart monzonite intrusions, Purcell Mountains, southeastern British Columbia: Canadian Journal of Earth Sciences, v. 25, p. 106-115.

Hyndman, D.W., 1980, Bitterroot dome-Sapphire block, an example of a plutonic-core gneiss-dome complex with its detached suprastructure, in Crittenden, M.D., Coney, P.J., and Davis, G.H., eds., Cordilleran core complexes: Geological Society of America, Memoir 153, p. 427-443.

Irving, E., and Wynne, P.J., 1991, Paleomagnetism: review and tectonic implications, Chapter 3 in Geology of the Cordilleran orogen in Canada, H. Gabrielse and C.J. Yorath, eds.: Geological Survey of Canada, no. 4, p. 61-86.

Leckie, D.A., and Smith, D.G., 1992, Regional setting, evolution, and depositional cycles of the western Canada foreland basin, in Macqueen, R.W., and Leckie, D.A., eds., Foreland basins and fold belts: American Association of Petroleum Geologists Memoir 55, p. 9-46.

Lidke, D.J., and Wallace, C.A., 1994, Structure sections across south-central Belt terrane, west-central Montana: Belt Symposium III, Proceedings, Whitefish, Montana, Special Publication of the Montana Bureau of Mines and Geology, in press.

Link, P.K., and 12 others, 1992, Middle and Late Proterozoic stratified rocks of the western U.S. Cordillera, Colorado Plateau, and Basin and Range province, in Reed, J.C., and 6 others, eds., Precambrian: Conterminous U.S.: Boulder, Colorado, Geological Society of America, The Geology of North America, v. C-2, p. 463-595.

Mallory, W.W., 1972, Geologic atlas of the Rocky Mountain region: Rocky Mountain Association of Geologists, Hirschfeld Press, Denver, Colorado, 331 p.

McMannis, W.J., 1963, LaHood Formation - A coarse facies of the Belt Series in southwestern Montana: Geological Society of America, Bull., v. 74, p. 407-436.

Mudge, M.R., 1970, Origin of the disturbed belt in northwestern Montana: Geological Society of America, Bulletin, v. 81, p. 377-392.

Mudge, M.R., Earhart, R.L., Whipple, J.W., and Harrison, J.E., 1982, Geologic and structure maps of the Choteau 1X2 degree Quadrangle, western Montana: U.S. Geological Survey, Map I-1300, scale 1:250,000.

O'Bradovich, J.D., Zartman, R.E., and Peterman, Z.E., 1984, Update of the geochronology of the Belt Supergroup: in Hobbs, S.W., ed., The Belt: Montana Bureau of Mines and Geology, Special Publication 90, p. 47-48.

Price, R.A., 1962, Fernie map-area, east half Alberta and British Columbia: Geological Survey of Canada, Paper 61-24.

Price, R.A., 1964, The Precambrian Purcell system in the Rocky Mountains of southern Alberta and British Columbia: Bulletin of Canadian Petroleum Geology, v. 12, p. 399-426.

Price, R.A., 1981, The Cordilleran foreland thrust and fold belt in the southern Canadian Rocky Mountains: Geological Society of London Special Publication 9, p. 427-488.

Price, R.A., and Mountjoy, E.W., 1970, Geologic structure of the Canadian Rocky Mountains between Bow and Athabasca Rivers - a progress report, in Wheeler, J.O., ed., Structure of the southern Canadian Cordillera: Geological Association of Canada Special Paper No. 6, p. 7-25.

Reynolds, M.W., 1984, Tectonic setting and development of the Belt basin, northwestern United States: in Hobbs, S.W., ed., The Belt: Montana Bureau of Mines and Geology, Special Publication 90, p. 47-48.

Sears, J.W., 1988, Two major thrust slabs in the west-central Montana Cordillera, in Schmidt, C.J., and Perry, W.J., eds., Interaction of the Rocky Mountain foreland and the Cordilleran thrust belt: Geological Society of America, Memoir 171, Ch 11, p. 165-170.

Sears, J.W., and Buckley, S.N., 1994, Cross-section of the Rocky Mountain thrust belt from Choteau to Plains, Montana: Implications for the eastern margin of the Belt Basin: Belt Symposium III, Proceedings, Whitefish, Montana, Special Publication of the Montana Bureau of Mines and Geology, in press.

Stockmal, G.S., Cant, D.J., and Bell, J.S., 1992, Relationship of the stratigraphy of the western Canada foreland basin to Cordilleran tectonics : Insights from geodynamic models, in Macqueen, R.W., and Leckie, D.A., eds., Foreland basins and fold belts: American Association of Petroleum Geologists Memoir 55, p. 107-124.

Turner, R.J.W., Hoy, T., Leitch, C.H.B., Anderson, D., and Ramsom, P., 1993, Guide to the Geological setting of the Middle Proterozoic Sullivan sediment-hosted Pb-Zn Deposit, southeastern British Columbia: Belt Association Guidebook, Link, P.K., ed., p. 53-94.

Viele, G.W., 1960, The geology of the Flat Creek area, Lewis and Clark County, Montana: PhD thesis, University of Utah, Salt Lake City, 213 p.

Wallace, C.A., and 14 others, 1987, Preliminary geological map of the Butte 1X2 degree quadrangle: U.S. Geological Survey, Open-File report 86-292, scale 1:250,000.

Wallace, C.A., Lidke, D.J., and Schmidt, R.G., 1990, Faults of the central Lewis and Clark line and fragmentation of the Late Cretaceous foreland basin in west-central Montana: Geological Society of America, Bulletin, v. 102, p. 1021-1037.

Weed, W.H., 1899, Little Belt Mountains, Montana: U.S. Geological Survey Folio 56.

Wheeler, J.O., and McFeely P., 1991, Tectonic assemblage map of the Canadian Cordillera and adjacent parts of the United States of America: Geological Survey of Canada, Map 1712A, scale 1:2,000,000.

White, B.G., 1978, Thrusting in the northwestern Montana overthrust belt: PhD thesis, University of Montana, Missoula, 103 p.

Whipple, J.W., Connor, J.J., Raup, O.B., and McGimsey, R.G., 1984, Preliminary report on the stratigraphy of the Belt Supergroup, Glacier National Park, and adjacent Whitefish Range, Montana: in Montana Geological Society Guidebook, 1984 Field Conference Northwestern Montana: Empire Press, Billings, p. 33-50.

Zieg, G.A., 1986, Stratigraphy and sedimentolgy of the Middle Proterozoic upper Newland Limestone, in Roberts, S.M., ed., Belt Supergroup: Montana Bureau of Mines and Geology, Special Publication 94, p. 125-142.

Zeig, G.A., and Leitch, C.H.B., 1994, The geology of the Sheep Creek copper deposits, Meagher County, Montana: Belt Symposium III, Proceedings, Whitefish, Montana, Special Publication of the Montana Bureau of Mines and Geology, in press.

GAS BUBBLE AND EXPANSION CRACK ORIGIN OF "MOLAR TOOTH" CALCITE STRUCTURES IN THE MIDDLE PROTEROZOIC BELT SUPERGROUP, WESTERN MONTANA

GEORGE FURNISS - Department of Geology, University of Montana, Missoula, MT 59801
JOHN F. RITTEL - Department of Geology, University of Montana, Missoula, MT 59801
DON WINSTON - Department of Geology, University of Montana, Missoula, MT 59801

ABSTRACT

"Molar tooth" are sedimentary structures of complex interconnecting, thin sheets and small spheroids, composed of uniform, blocky calcite crystals 5-15 μm in diameter. The structures, which resemble the surface of an elephant's tooth (Bauerman, 1884), are common in dolomitic argillite of the Middle Proterozoic Belt and Purcell Supergroups of the northern Rocky Mountains. Molar tooth are particularly common in the Helena and Wallace Formations in western Montana (Fig. 1). Experiments that closely replicate molar tooth structures suggest that molar tooth formed as gas-generated voids which were filled with blocky calcite before the sediments were lithified. Experimental models using mud, yeast and sugar in glass aquaria produced gas bubbles and gas expansion cracks closely mimicking the shapes of molar tooth structures. Finely crystalline blocky calcite similar to the molar tooth calcite formed rapidly when solutions of $CaCl_2$ and Na_2CO_3 were mixed in the laboratory suggesting that the uniform, blocky mosaic texture of the "molar tooth" calcite precipitated rapidly. Therefore we interpret molar tooth structures to have developed after gas generated in Belt sediments formed bubbles and cracks, which were filled by pore water that became supersaturated with respect to calcite.

Calcite precipitated as a result of gas and pore water interactions. Suggested gasses include CO_2, H_2S, and CH_4. Based on diagenetic sequences described from modern sediments (Neev and Emery, 1967), CO_2 was probably the first gas produced by oxidation of organic matter. After available oxygen was consumed, H_2S may have been generated by sulfate reduction. Pyrite found within molar tooth structures supports sulfate reduction a short distance below the depositional interface. The absence of gypsum, but the presence of halite casts, in the Helena Formation may also point to sulfate reduction of gypsum similar to that inferred from the Dead Sea sediments (Neev and Emery, 1967). Organic films within molar tooth calcite may indicate methane generation as the final gas-producing process. All three gasses promote precipitation of calcite. In Dead Sea sediments, gypsum near the surface disappeared at depth, perhaps resulting from bacterial sulfate reduction which produces H_2S and CO_2 gases. We suggest a similar origin for the molar tooth structures and the calcite which preserved them. Molar tooth structures in the Belt commonly form repeated sequences of shapes in fining-upward siliciclastic-to-dolomite cycles (O'Connor, 1967). Escaping H_2S and CO_2 gases in the presence of CH_4 gas may have induced supersaturation with respect to calcite in pore water during regression and contraction of Belt waters. The precipitation of ample calcite preserved the original gas-formed structures in the Proterozoic sediments.

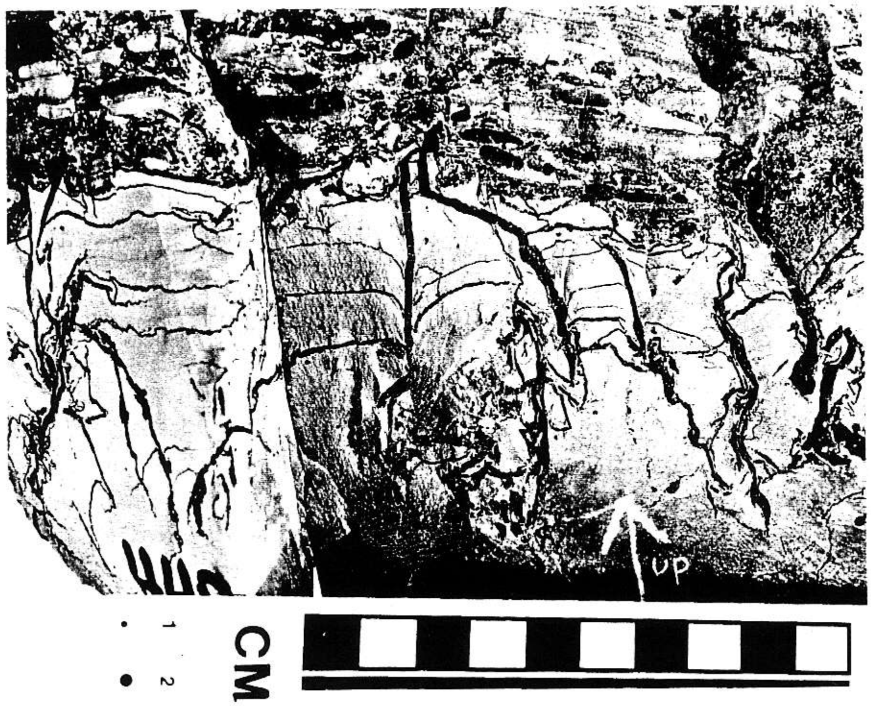


Figure 1. Network of "molar tooth" ribbon calcite (black) in Helena Dolomite near Hungry Horse Dam, protruding through scoured surface into conglomerate bed. Scale bar divisions indicate centimeters. Arrow shows stratigraphic-up direction.

BIBLIOGRAPHY

Adshead, J. D., 1963, Petrology of the carbonate rocks of the Siyeh Formation, S. W. Alberta [unpubl. M. S. thesis]: University of Alberta.

Bauerman, H., 1884, Report on the geology of the country near the forty-ninth parallel of the North latitude West of the Rocky Mountains: Canada Geol. Survey Rept. Prog. 1882-84, pt. B, p. 1-42.

Bell, Richard T., 1966, Precambrian rocks of the Tuchodi Lakes map-area, northeastern British Columbia, Canada [unpubl. Ph.D. thesis]: Princeton University, 138 p. (University Microfilms).

Boyce, R. L., 1975, Depositional systems in the LaHood Formation, Belt Supergroup, Precambrian, south-western Montana [unpubl. Ph. D. thesis]: University of Texas at Austin.

Burst, J. F., 1965, Subaqueously formed shrinkage cracks in clay: Jour. Sed. Petrology, v. 35, p. 348-353.

Cayeux, L., 1935, Les Roches Sedimentaires de France: Roches Carbonatees, Masson, Paris, [A translation of this work by A. V. Carozzi (1970) in English]: Hafner Publishing Co., Inc., Darien, Conn., 463 p.

Calver, Clive R., and Baillie, Peter W., 1990, Early Diagenetic Concretions Associated with Intrastratal Shrinkage Cracks in an Upper Proterozoic Dolomite, Tasmania, Australia: Jour. Sed. Petrology, v 60, p. 293-305.

Daly, R. A., 1912, Geology of the North American Cordillera at the forty-ninth parallel: Canada Geol. Survey Mem. 38, pts. 1,2, and 3.

Eby, D. E., 1975, Carbonate sedimentation under elevated salinities and implications for the origin of "molar-tooth" structures in the middle Belt carbonate interval (late Precambrian), Northwestern Montana (abst.): Geol. Soc. Amer. Ann. Meet. Salt Lake City. Abstr. with Programs v. 7, p. 1063.

Eby, D. E., 1977, Sedimentation and early diagenesis within eastern portions of the "Middle Belt Carbonate Interval" (Helena Formation), Belt Supergroup (Precambrian Y), western Montana [unpubl. Ph.D. thesis]: State University of New York, Stony Brook, 712 p.

Eby, D. E., and Yeo, G. M., 1977, Proterozoic "molar-tooth" structures of North America: Geol. Soc. America, Abstr. Programs, v. 9, no. 7, p. 962.

Fenton, C. L., and Fenton, M. A., 1936, Walcott's "Precambrian Algonkian algal flora" and associated animals: Geol. Soc. America Bull., v. 47, p. 609-620.

Fenton, C. L., and Fenton, M. A., 1937a, Studies of fossil calcareous algae (abs.): Geol Soc. America Proc., 1936, p 354.

Fenton, C. L., and Fenton, M. A., 1937b, Belt series of the north: stratigraphy, sedimentation, paleontology: Geol. Soc. America Bull., v. 48, p. 1873-1969.

Furniss, G., 1990, Gass bubble and expansion crack origin of "molar tooth" calcite sturctures in the Middle Proterozoic Belt Supergroup, Western Montana (unpubl. M.S. thesis):

Gillson, J. L., 1929, Contact metamorphism of the rocks in the Pend Oreille district, northern Idaho: U.S. Geol. Survey Prof. Paper 158-F, p. 111-121.

Grotzinger, J. P., 1981, The stratigraphy and sedimentation of the Wallace Formation, northwestern Montana and northern Idaho [M. S. thesis]: University of Montana, Missoula, 153 p.

Grotzinger, J. P., 1986, Shallowing-upward cycles of the Wallace Formation, Belt Supergroup, northwestern Montana and northern Idaho: in S. Roberts, (ed.)., Belt Supergroup: A Guide to Proterozoic Rocks of Western Montana and Adjacent Areas: Montana Bureau of Mines and Geology Special Publication 94, p. 143-160.

Horodyski, R. J., 1976, Stromatolites of the upper Siyeh Limestone (Middle Proterozoic), Belt Supergroup, Glacier National Park, Montana: Precambrian Research, v. 3, p. 517-536.

Horodyski, R. J., 1985, Stromatolites of the Middle Proterozoic Belt Supergroup, Glacier National Park, Montana: a summary and a comment on the relationship between their morphology and paleoenvironment, in Toomey, D. F. Nitecki, M. H., (eds.), Paleoalgeology: Contemporary Research and Applications: Springer Verlag, Berlin, p. 34-39.

JCPDS, 1974, Mineral powder diffraction file data book, Joint Committee on Powder Diffraction Standards, Swarthmore, Penn., file # 5-586, p. 145.

Neev, D. and Emery, K. O., 1967, The Dead Sea depositional processes and environments of evaporites: Israel Geol. Survey Bull. 41, 147 p.

O'Connor, M. P., 1965, "Molar-tooth" structure--a probable Precambrian alga. (abs.): Geol. Soc. America Spec. Paper 82, p. 340.

O'Connor, M. P., 1967, Stratigraphy and petrology across the Precambrian Piegan Group-Missoula Group boundary, southern Mission and Swan ranges, Montana [unpubl. Ph.D. thesis]: University of Montana, Missoula, 269 p.

O'Connor, M. P., 1972, Classification and environmental interpretation of the cryptalgal organo sedimentary "molar-tooth" structure from the Late Precambrian Belt-Purcell Supergroup: J. Geol., v. 80, no. 5, p. 592-610.

Peterson, J. A., 1971, Precambrian cyclic carbonate facies, western Montana: Rocky Mountain Section American Assoc. Petrol. Geologists Bull, v. 55, p. 540.

Pflug, Hans D., 1968, Gesteinbildende Organismen aus dem Molar

Tooth Limestone der Beltserie (Praekambrium): Palaeontographica ser. B, V. 121, nos. 4-6, p. 134-141.

Pratt, Brian R., 1992, Shrinkage features ("molar tooth" structure) in Proterozoic limestones; new model for their origin through synsedimentary earthquake-induced dewatering: Geol. Soc. America, Abstr. Programs, v. 24, no. 7, p. 53.

Rezak, R., 1957, Stromatolites of the Belt Series in Glacier National Park, and vicinity, Montana: U.S. Geol. Survey Prof. Paper 294-D, p. 127-154.

Ross, C. P., 1959, The geology of Glacier National Park and the Flathead region, northwestern Montana: U.S. Geol. Survey Prof. Paper 296, 125 p.

Ross, C. P., and Rezak, R., 1959, The rocks and fossils of Glacier National Park; The story of their origin and history: U. S. Geol. Survey Prof. Paper 294-K, p. 401-439.

Ross, C. P., 1963, The Belt Series of Montana: U. S. Geol. Survey Prof. Paper 346, 122 p.

Schieber, J., 1985, The relationship between basin evolution and genesis of stratiform sulfide horizons in Mid-Proterozoic sediments of central Montana (Belt Supergroup). [unpubl. Ph.D. thesis]: University of Oregon, at Eugene, 811 p.

Schieber, J., 1986a. The possible role of benthonic microbial mats during the formation of carbonaceous shales in shallow Mid-Proterozoic basins: Sedimentology, v. 33, p. 521-536.

Schieber, J., 1989, Pyrite mineralization in microbial mats from the mid-Proterozoic Newland Formation, Belt Supergroup, Montana, U.S.A.: Sedimentary Geology, b. 64, p. 79-90.

Smith, A. G., and Barns, W. C., 1966, Correlation of and facies changes in the carbonaceous, calcareous, and dolomitic formations of the Precambrian Belt-Purcell Supergroup: Geol. Soc. America Bulletin, v. 77, no. 12, p.1399-1426.

Smith, A. Gilbert., 1968, The origin and deformation of some "molar-tooth" structures in the Precambrian Belt-Purcell Supergroup: J. Geol., v. 76, no. 4, p. 426-443.

Walcott, C. D., 1914, Precambrian Algonkian algal flora in Cambrian geology and paleontology, pt. 3: Smithsonian Misc. Colln., b. 64, p. 77-156.

Winston. D., Horodyski, R. J., and Whipple, J. W., 1989, Middle Proterozoic Belt Supergroup, Western Montana: American Geophysical Union 28th International Geological Congress: Field Trip Guidebook T334, July 20-28, p. 13.

Yeo, G. M., 1978, "Molar tooth" structure and penecontemporaneous dolomitization of shallow marine, Proterozoic carbonates: Geol. Soc. America, Abstr. Programs, v. 10, no. 7, p. 520.

Young, G. M., and Long, D. G. F., 1977, Carbonate sedimentation in a late Precambrian shelf sea, Victoria Island, Canadian Arctic Archipelago: J. Sediment. Petrol. v. 47, no. 3, 943-955 p.

TROY John Balla Ph.D. Tim Hayes

~60' thick

$250 / capital costs

134,535,000 0.74% Cu 1.93 oz/Ag

Montanore – Noranda

"ore rod" – dewatering features

– throughout section

– roll front – graben to east

– early to mid diagenetic mineralization

Rockore

same horizon as Montanore

– not much sign – green malachite, iron oxide staining

– intense oxidation leaches copper and leaves silver

12000' 27000?' 27' thick

"Blue Haze"

lead zone 2.5% Pb 0.29% Cu 0.20 opt Ag

MINING UNDER WILDERNESS AREA

in 200' section to

THREE-DIMENSIONAL STRUCTURE OF THE ROCKY MOUNTAIN TRENCH REVEALED BY SEISMIC REFLECTION PROFILING

ARIE J. VAN DER VELDEN - Department of Geology and Geophysics, University of Calgary, Calgary, Alberta, T2N 1N4
FREDERICK A. COOK - Department of Geology and Geophysics, University of Calgary, Calgary, Alberta, T2N 1N4

ABSTRACT

More that 800 km of industry seismic reflection data provide six crossings of the Rocky Mountain Trench Fault system between 49°N and 50°30'N. Prominent reflections from the mid-Proterozoic Moyie sills can be tied to well data and outcrop on both sides of the Trench. At 49°N, a direct match between these reflections in the hanging wall and footwall of the Rocky Mountain Trench Fault indicates ~10 km of extension. Displacement on the Rocky Mountain Trench Fault decreases to ~4-6 km at 50°30'N, accompanied by a decrease in rollover in the hanging wall of the fault. The near-basement reflections can be tied to data extending to the foreland, and can be followed across the Trench without offset on all six Trench crossing. The Rocky Mountain Basal Detachment follows the near-basement reflections, and changes abruptly from gently to steeply west-dipping (1°-4° to >15°) on the west side of the Trench. The Rocky Mountain Trench Fault flattens into the Basal Detachment near this hinge, which may have been reactivated and appears to control the location of the fault.

Retrodeformation of the interpreted data along 49°N combined with data extending to the foreland places a pronounced increase in thickness and a facies change in lower Belt-Purcell strata near a change in crustal thickness (~36 km to the west and ≥42 km to the east) and a change in slope of the top of the North American Basement. This transition may represent past of a margin that was established initially in early Proterozoic time, on which a thick sequence of sediments was deposited that was compressed in Mesozoic time and was subject to regional extension in Eocene time.

- 3.5 km hole in Moyie Anticlinorium by Denver Energy
- correlate sills across listric RMF in trench which flattens at basement
- 20 km of rock between well head at Moyie drill head and basement stratigraphy
- basement slice - explains Bouguer gravity high, seismic signature,
- another basement slice under Creston anticline

Northwest Geology, v. 23, p. 97, 1994

- can see moho at 36 km near St. Mary's lake and it appears to deepen to east at RM Trench
- basement dip flat east of trench, steepens to west and then flattens

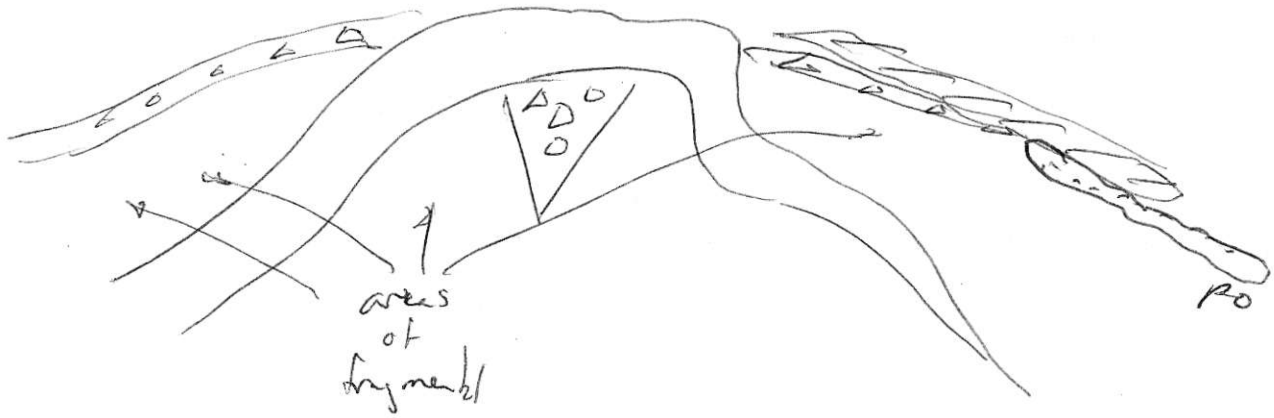
- ancient basement ramp existed in this area
- RM Fault following this basement ramp

T. Höy

- subsidence related to sill emplacement (S. Buckley)
- ? Premier L., Mt Fisher? near Kootenay King, not SEDEX
- Bull River – Cu-Pb-Zn – Aldridge + Moyie Sills
- garnets in Goldstream & Sullivan – "coticules"
- mud volcanoes
- Kootenay King – St Joe horizon?

B. Turner

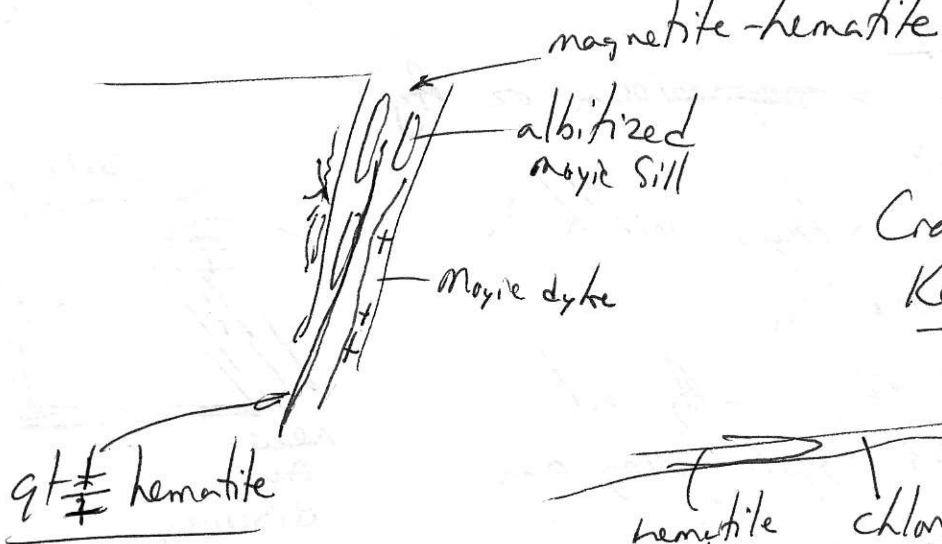
- Sullivan Arch sill has 80%? core of granophyre which is unusual? same as Fors



- replacement body of pyrrhotite
- Stemwinder – fragmental – some conformable?
- three hydrothermal centres in corridor
- albite alteration related to sills (red-herring) – no direct link with ore formation
- chl-py associated with HW albite at Sullivan – Mg-rich seawater interacting with Na-rich fluid that formed albitite

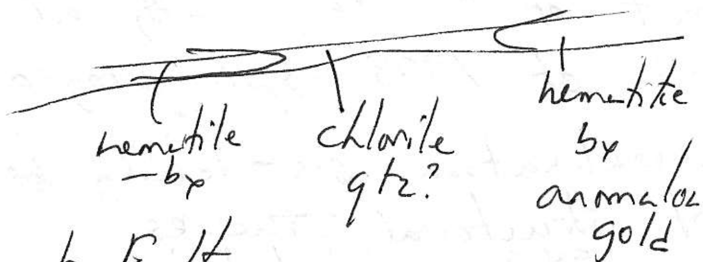
IRON RANGE

SCHEMATIC MODEL



Craig Kennedy

Peter Klewchuk



- Cranbrook Fault
- St Marys Fault
- correlation with placer gold

Leg

Western dolomite — breccia

- Karst or tectonic?

- volcanic unit with some sills into dolomite ? how Derek?

minor galena
Kootenay Arc
Shallower!
Irish Deposits
epigenetic and syngenetic

dolomite
halo

- event - same age as Nicol Creek lavas
- lower in strata

quartzite
volcanic

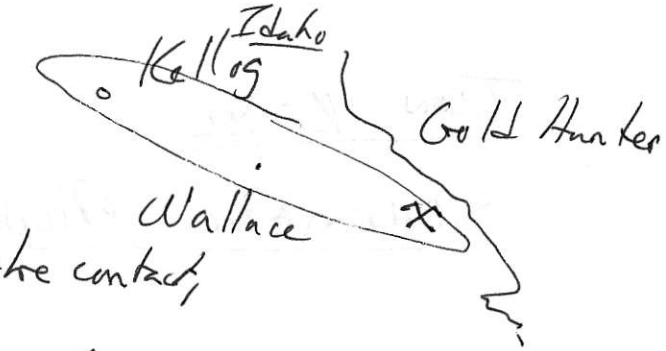
803 CRANBROOK ST., CRANBROOK, B.C. V1C 3S2. PHONE (604) 489-4301, 1-800-663-2708. FAX (604) 489-5758.

- dolomites are magnetic!

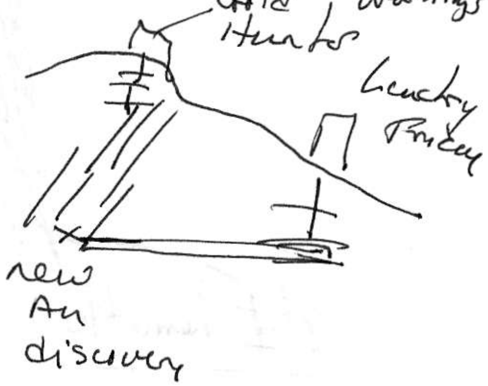
Gold Hunter Deposit

Terry Devoe
Hecla Mining Company

- accessed from Lucky Friday deposit in Coeur d'Alene



- Au - zone unusual
- most production from middle Prichard, Pritchard/Burke contact, and St. Regis
- 6.2 mt - ~~100,000~~ 100 m oz Ag
- 1885 to 1998
- mineralization stays within Wallace Formation



$\frac{1}{2}$ mt of 5% Pb, — Ag at end of 1993 expl. program

- mineralization post folding because of structural studies

py, gal, sphal, tedrahedrite, bournite, cassiterite, barite, ankerite

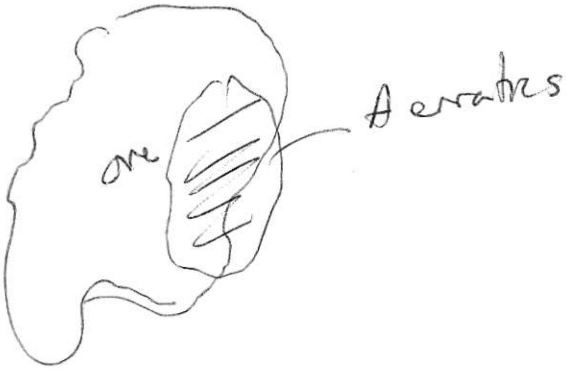
- fragmented ore textures? - galena cement common
- used chemistry and didn't show any zoning
- Siderite close to veins in wallrock (< 10's of feet) into ankerite, into carbonate — in Wallace Carbonate
- correlation with quartzite
- isotopic Sulphur

Coeur d'Alene { 0 — 15

only possible as Sedex vent facies

Paul Ransom

- Sullivan sub-basin
- aw are of multiple sand volcanoes in basin apparently
- roots beneath ore body "chaotic breccia" – N-S trending
- slump sheet of barren, well-laminated pyrrhotite – folded laminations near base with isoclines with SW vergence ~derived from distal orebody to NE
- separates from waste beds & debris flows with large lens of pyrrhotite (laminated) and other clasts



Po-source
no distal
sulphides
at "A"? horizon

- "thickened turbidites"
- I, H, U = concentrator hill horizon
- "injected turbidites" – into oozes
- asymmetric, tapering depression
- meta-sediment between gabbro sills at Sullivan – not granophyre

NORTHWEST GEOLOGY

Montana State University

Volume 23

October 1994

Contents