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THE ORIGIN OF MINERALIZED FRACTURES AT THE BLUEBELL MINE SITE, RIONDEL, BRITISH COLUMBIA

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Abstract

The Bluebell Pb-Zn deposit is located on the east side of Kootenay Lake in southeastern British Columbia. It is a fracture-controlled replacement deposit hosted in Lower Cambrian marble of the Badshot and Mohican formations. The orebodies trend west-northwest, parallel to a prominent set of fractures that controlled mineralization. These fractures are interpreted to have formed as part of a conjugate set. Contrary to some earlier suggestions, folding played no role in fracture formation. The youngest generation of folds are reinterpreted as having formed through development of shear bands during extension of the layering prior to fracture development. Formation of the fractures also postdated motion on a number of early Tertiary normal faults that are associated with regional extension and alkaline magmatism. West-northwest–trending fractures, which formed as part of a conjugate set, were reactivated and controlled the intrusion of mafic dikes, mineralizing fluids, and minor late faulting. Fracturing, faulting, mafic dike intrusion, and mineralization may be associated with a long-lived basement structure that runs below the area, which could have facilitated the ascent of deep crustal or mantle-derived magma and mineralizing fluids. The results of this study provide a new perspective on exploration strategies for hydrothermal ore deposits in this part of southeastern British Columbia.

Introduction

The Bluebell Pb-Zn deposit is located in Riondel, on the eastern shore of Kootenay Lake in southeastern British Columbia, western Canada (Fig. 1). It is hosted by Lower Cambrian marble of the Badshot and Mohican formations. The Badshot Formation and along-strike equivalents form a prominent marker unit in the Kootenay Arc, a narrow, elongate region of the Canadian Cordillera that extends in an eastward-convex arch from near Revelstoke, British Columbia, into northern Washington (Fig. 1).

The Kootenay Arc contains numerous carbonate-hosted Pb-Zn deposits of varying age and mode of formation. Fyles (1970) recognized three types of deposit:

1. "Metaline-type" deposits, named after the district in Washington, are developed in brecciated, dolomitized limestone and bear many similarities to classic Mississippi Valley-type deposits. They are lenticular, more or less stratiform deposits in relatively undeformed Middle Cambrian carbonate rocks of the Nelway and Metaline formations (Fig. 1).
2. "Salmo-type" deposits are found in dolomitized zones of Lower Cambrian (Badshot Formation and equivalent) limestone. They comprise lenticular disseminations of pyrite, sphalerite, and galena in rocks that have undergone polyphase deformation. The sulfide is typically localized by folds and has been penetratively deformed. This type includes the Jersey, H.B., and Reeves McDonald deposits around Salmo and the Duncan deposit in the Lardeau district (Fig. 1).
3. The third type of ore deposit, known as the "Bluebell type," consists of massive or disseminated sulfides in limestone and marble adjacent to fractures in a number of different stratigraphic units. These deposits are generally richer in silver, have a more complex mineralogy and are higher grade than Salmo-type deposits (Fyles, 1970). Examples of this type include Bluebell, the Florence-Lakeshore deposit near Ainsworth, and the Lucky Jim deposit north of Sandon (Fig. 1; Fyles, 1967).

The Bluebell-type deposits form a subset of vein and replacement-type Ag-Pb-Zn-Au deposits that are found around the Nelson batholith and adjacent areas (Fig. 1). Beaudoin et al. (1992a) studied the isotopic characteristics of this suite of deposits and, based on Pb isotopes, identified four groups: the Kokanee, Sandon, Ainsworth, and Bluebell groups. They interpreted three sources of Pb—depleted mantle, lower crust, and upper crust. The Bluebell group (including the Bluebell deposit and some deposits in the northern part of the Ainsworth camp) is characterized by Pb interpreted to be from depleted mantle and lower crust, whereas Pb in the Ainsworth deposits was considered to be derived from the upper and lower crust. The Sandon and Kokanee Groups contain Pb interpreted to be derived exclusively (Kokanee group) or dominantly (Sandon group) from the upper crust. A mantle source for CO_2 was also suggested by Beaudoin et al. (1991).

This suite of fracture-controlled deposits formed after regional metamorphism and ductile deformation associated with Mesozoic contractional Cordilleran orogenesis, probably during the Eocene (Fyles, 1967; Beaudoin et al., 1992b). The subject of this paper is the formation of the fracture network that facilitated mineralization at Bluebell. In particular, we focus on the hypothesis that fractures that controlled mineralization formed during, and as a result of, folding. We present data on ductile and brittle structures around the deposit,

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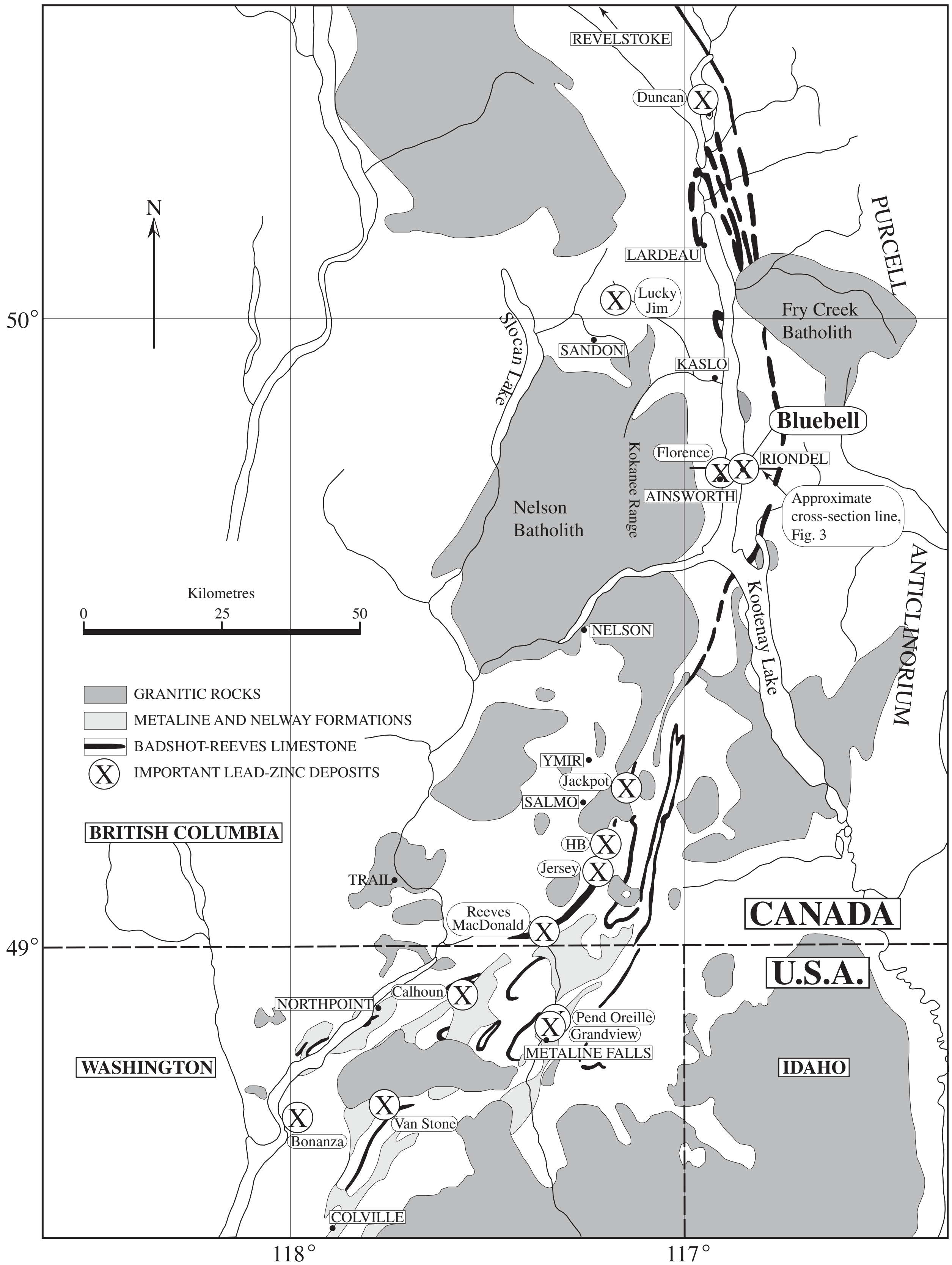

FIG. 1. Map of the Kootenay Arc showing the location of important lead-zinc deposits, igneous rocks, and major host carbonate units. The approximate line of section from Figure 3 is also shown. The name "Kootenay Arc" refers to the arcuate shape of the belt rather than to a magmatic arc. After Fyles (1970).

propose a new interpretation of the youngest fold generation, and argue that folding preceded fracture formation. The deposit formed above a multiply reactivated Paleoproterozoic basement structure that may have provided a conduit for deep mineralizing fluids.

Regional Geology

The Kootenay Arc is a narrow, curvilinear, metamorphosed, and polydeformed region that forms part of the interior of the Canadian Cordilleran orogen in southeastern British Columbia. It is located to the west of the foreland fold and thrust belt, on the west flank of the Purcell anticlinorium, a large north-plunging structure cored by Mesoproterozoic rocks of the Belt-Purcell Supergroup (Price, 1981). The Kootenay Arc straddles the boundary between rocks that formed on the ancestral North American continental margin in the Proterozoic-early Paleozoic and those that formed in oceanic and back-arc environments to the west of ancestral North America during the late Paleozoic-early Mesozoic (Klepacki, 1985; Colpron and Price, 1995; Warren, 1997; for an alternative view see Thompson et al., 2007).

Rocks of the Kootenay Arc were deformed and regionally metamorphosed in the Middle Jurassic-Early Cretaceous during shortening associated with formation of the Canadian Cordillera (Archibald et al., 1983; Leclair et al., 1993). They were intruded by major plutonic suites in the Middle Jurassic, and again in the mid-Cretaceous (115–95 Ma). The mid-Cretaceous plutons mostly postdate regional deformation and metamorphism but there was further localized deformation and minor magmatism associated with early Tertiary extension (Fyles, 1967; Fyles et al., 1973; Archibald et al., 1984; Sevigny and Theriault, 2003; Moynihan and Pattison, 2008). Early Tertiary extensional structures are widely developed across southeastern British Columbia, although estimates of the magnitude of extension vary significantly (e.g., Parrish et al., 1988; Johnson and Brown, 1996; Glombick et al., 2006).

Stratigraphy

The tight folds of the Kootenay Arc are developed in rocks of Neoproterozoic-Mesozoic age (Fig. 2; Fyles and Eastwood, 1962; Fyles, 1964, 1967; Hoy, 1977). The Neoproterozoic-Lower Cambrian Hamill Group (Colpron et al., 2002) is dominated by quartz-rich metasedimentary rocks with minor amphibolite and calc-silicate. A regional unconformity is developed in the Hamill Group (Devlin and Bond, 1988; Warren, 1997), separating units that were deposited in fault-bounded basins during rifting from an upper part distinguished by laterally continuous units deposited in a shallow-marine setting. This unconformity is interpreted to record the change from active continental rifting to thermal subsidence on a passive margin between 549 and 520 Ma (Devlin and Bond, 1988; Warren, 1997).

The Upper Hamill Group is conformably overlain by the Mohican and Badshot Formations (Fig. 2; Fyles and Eastwood, 1962; Fyles, 1964). The Mohican Formation is a transitional unit comprising interlayered silicilastic and carbonate

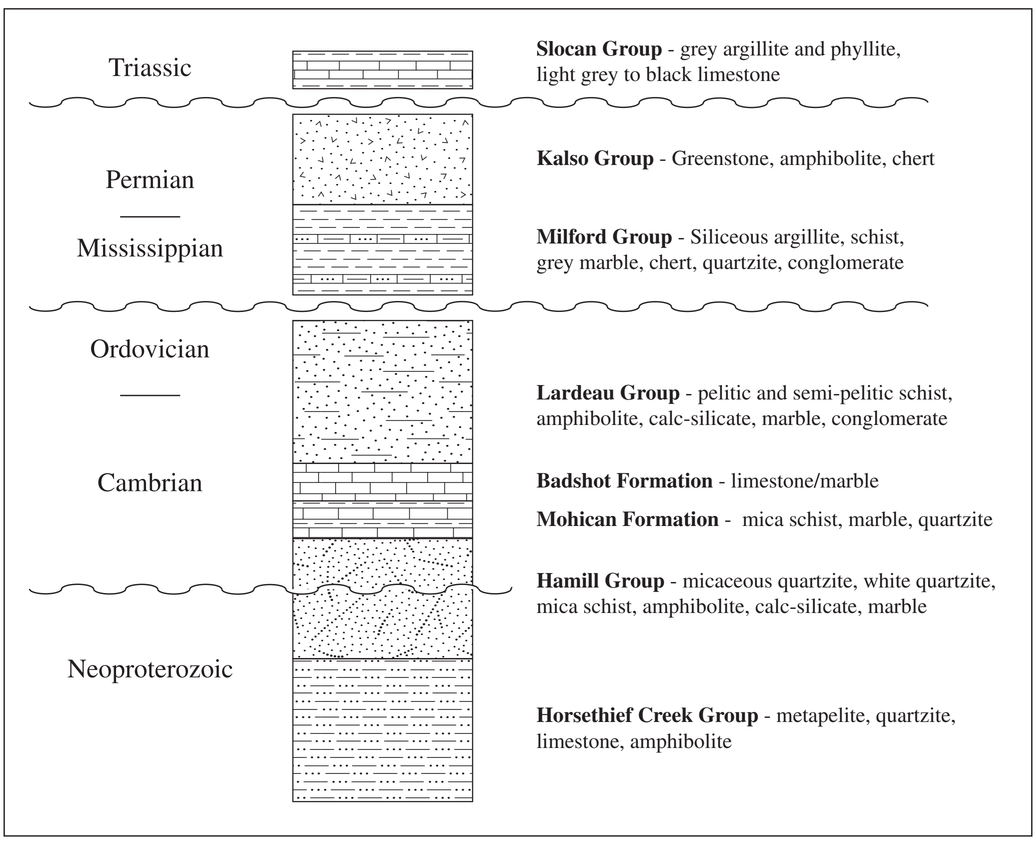

FIG. 2. Summary of the stratigraphy of the Kootenay Arc, based on Fyles and Eastwood (1962), Fyles (1964, 1967), Reesor (1973), Hoy (1980), Klepacki (1985), Devlin and Bond (1988), Leclair (1988), and Warren (1997).

metasedimentary rocks. It is overlain by Archaeocyathid-bearing calcite and dolomite marble of the late Lower Cambrian Badshot Formation. The Badshot Formation forms a laterally continuous marker unit and is interpreted to have been deposited on a tectonically quiescent shallow-marine shelf (Warren, 1997). The Badshot Formation is followed in conformable succession by the lower Paleozoic Lardeau Group, a varied sequence comprising siliclastic metasedimentary rocks, mafic metavolcanic rocks, and carbonate and calc-silicate rocks (Fyles and Eastwood, 1962; Fyles, 1964; Hoy, 1977; Colpron and Price, 1995, and references therein).

The lowest part of the Lardeau Group is a fine-grained black metapelite that records deposition under deep water, anoxic conditions. Its contact with the Badshot Formation is interpreted to mark the point when the rate of carbonate production could no longer keep pace with subsidence (Warren, 1997). A return to active rifting is recorded by metavolcanic rocks and coarse grits of upper parts of the Lardeau Group. This post-Cambrian extension on the western margin of ancestral North America is interpreted to be responsible for differences between the Lardeau Group and age-equivalent strata to the east (Colpron and Price, 1995).

The Lardeau Group is unconformably overlain by a sequence comprising upper Paleozoic-Mesozoic rocks of the Milford, Kaslo, and Slocan groups (Fig. 2). These rocks, which include metamorphosed limestone, argillite, sandstone, conglomerate, and mafic volcanic rocks, are generally interpreted to record deposition in back-arc environments to the west of ancestral North America, prior to Cordilleran shortening (Fyles, 1967; Klepacki, 1985; Roback et al. 1994; Hoy and Dunne, 1997).

Igneous rocks

Metasedimentary and metavolcanic rocks of the Kootenay Arc host numerous Middle Jurassic (ca. 165 Ma) and mid-Cretaceous granitic plutons and minor intrusive bodies (Archibald et al. 1983, 1984; Logan, 2002). In the central Kootenay Arc, the Middle Jurassic plutonic suite is represented by the calc-alkaline Nelson batholith and associated minor bodies. The Nelson batholith is an 1800 km^2 body intruded during the interval 159 to 173 Ma (Ghosh, 1995). It ranges in composition from diorite to granite, but is dominated by porphyritic hornblende granodiorite. The second major plutonic suite was intruded during the interval 117 to 95 Ma. Rock types include hornblende and biotite granodiorite, biotite granite, and two-mica granite, which are interpreted to have been derived from crustal anatexis (Brandon and Lambert, 1993). Examples include the Bayonne, Fry Creek, and White Creek batholiths.

The youngest igneous rocks in the Kootenay Arc are early Tertiary mafic djkes and small intrusions, some of which are lamprophyre (Fyles, 1967; Leclair, 1988; Beaudoin et al., 1992b; Sevigny and Theriault, 2003). Beaudoin et al. (1992b) reported K-Ar dates in the range of 26 to 30 Ma for whole-rock analyses of altered gabbroic dikes from Bluebell; however, these dates were interpreted as resulting from alteration and are not thought to represent the crystallization age of the dikes.

Deformation and metamorphism

The outcrop pattern in the central Kootenay Arc is dominated by two generations of gently plunging folds (Fig. 3; Fyles, 1964, 1967; Hoy, 1977, 1980; Leclair, 1988). The earliest folds are a series of high amplitude isoclines with an axial-planar schistosity. The most clearly defined F_1 folds in the central Kootenay Arc are westward-closing recumbent anticlines cored by the Hamill Group. The largest of these, the Riondel nappe (Hoy, 1977) (equivalent to the Meadow Creek anticline in the Duncan Lake area; Fyles, 1964) has a 20-km-long overturned lower limb (Fig. 3). F_1 isoclines were coaxially refolded around gently plunging F_2 axes, giving rise to a type 3 interference pattern (Ramsay, 1967). F_2 axial planes generally dip gently to moderately steeply to the west, but steepen toward higher structural levels to the east (Hoy, 1980). Mineral lineations (L_1 and L_2) generally plunge gently north or south, parallel to fold axes.

Deformation took place in the Kootenay Arc between deposition of the Cambrian-Ordovician Lardeau Group and the Mississippian Milford Group, and again in Late Permian-Middle Triassic (Read and Wheeler, 1976; Klepacki, 1985). However, Cordilleran (Middle Jurassic-Lower Cretaceous) deformation is responsible for the dominant structures and fabrics in the central Kootenay Arc (Archibald et al., 1983; Leclair et al., 1993; Warren, 1997).

Cordilleran deformation was accompanied by Barrovian regional metamorphism. A narrow, elongate region of anomalously high metamorphic grade runs parallel to Kootenay Lake (Crosby, 1968; Livingstone, 1968; Reesor, 1973; Hoy, 1976; Archibald et al., 1983; Pattison et al., 2010). Metamorphic grade ranges from the biotite zone on the flanks of this

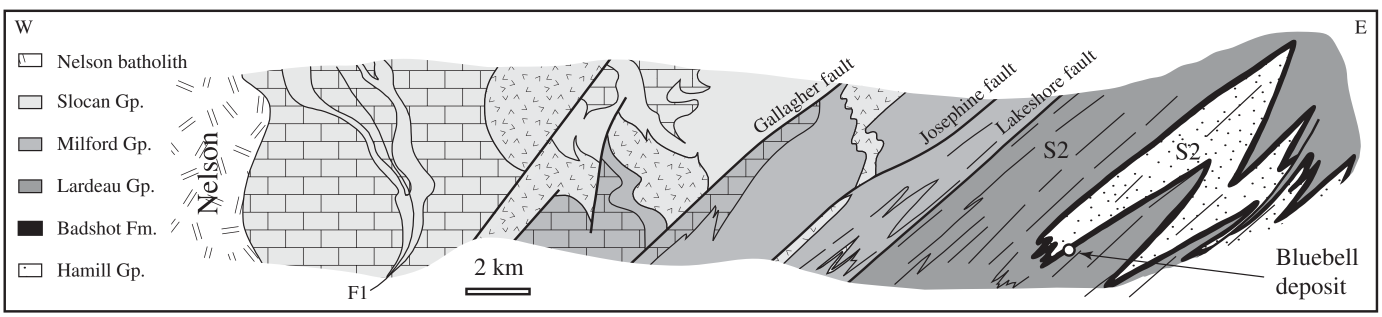

FIG. 3. Cross section across the central Kootenay Arc, showing the location of the Bluebell mine on the overturned limb of a refolded fold nappe. The approximate line of section is marked on Figure 1. The normal faults around Ainsworth, on the west side of Kootenay Lake, are labeled. From Fyles and Read (1981), based on Fyles (1967) and Hoy (1980).

high to the sillimanite zone in its core. The Bluebell deposit lies in the sillimanite zone, in the center of this metamorphic high. Rocks in the center of this belt were metamorphosed under conditions approximating 650° to 700°C at 20 to 25 km depth (Hoy 1977, Archibald et al., 1983; Moynihan and Pattison, 2008). The metamorphic high is bounded on the west side by a series of normal faults that accommodated differential exhumation during early Tertiary extension. From east to west, they are the Lakeshore, Josephine, and Gallagher faults (Fig. 3). The Gallagher fault marks the western boundary of the amphibolites facies belt and coincides with a change in $^{40}Ar/^{39}Ar$ mica cooling ages from Jurassic-Early Cretaceous in the hanging wall to early Tertiary in the footwall (Mathews, 1983; Archibald et al., 1984; Moynihan and Pattison, 2008). Outside of this amphibolites facies belt, greenschist and transitional greenschist-amphibolite facies regional metamorphic assemblages are ubiquitous, but are locally overprinted in low-pressure contact-metamorphic aureoles (Archibald et al., 1983; Pattison and Vogl, 2005).

Basement controls on sedimentation, deformation and mineralization in the Kootenay Arc and Purcell anticlinorium

The craton to the east of and below the southeastern Canadian Cordillera comprises a number of Archaean and Palaeoproterozoic domains that were assembled during the Paleoproterozoic (Ross, 1991; Price and Sears, 2000, and references therein). Tectonic boundaries between these domains trend northeast-southwest, and intermittent reactivation of these basement structures affected patterns of sedimentation, deformation, and mineralization in overlying rocks.

A prominent northeast-southwest–trending zone referred to as the Vulcan structure projects from the North American craton across the Purcell anticlinorium and southern Kootenay Arc (Price, 2000; Price and Sears, 2000). Syndepositional normal faults that formed above this basement structure account for large variations in the thicknesses and facies distribution of Mesoproterozoic, Neoproterozoic, and Paleozoic rocks (Lis and Price, 1976; Price, 2000; Price and Sears, 2000). The zone also exerted a control on the location of clastic-hosted (Mesoproterozoic) and carbonate-hosted (Lower-Middle Cambrian) Pb-Zn deposits (Kanasewich, 1968; Hoy, 1982; Hoy et al. 2000b). Reactivation of this zone during Cordilleran deformation led to right-lateral reverse faulting and localized granitic intrusions along the same trend (Price, 2000; Price and Sears, 2000).

Another multiply-reactivated northeast-southwest–trending basement structure recognized by McMechan (2010) crosses the central Kootenay Arc around the latitude of Riondel-Ainsworth. Features associated with this zone include anomalous northeast-trending faults in the Rocky and Purcell mountains, a cluster of Ordovician-Early Devonian diatreme pipes, Pb-Zn showings, and sedimentary facies and thickness changes in Mesoproterozoic and Paleozoic rocks. It may also have acted as a conduit for the mid-Cretaceous White Creek batholith, which is intruded along one of the major transverse faults in the Purcell anticlinorium. When palinspastically restored to account for Cordilleran deformation, these features line up parallel to the Red Deer zone, a geophysically imaged boundary that marks the northwest margin of the Archean Hearne province in the cratonic basement (McMechan, 2010).

The Bluebell Deposit

The Bluebell deposit was staked in 1882 and between 1885 and 1927, 540,000 tons (t) of ore, having an average grade of 6.5 percent Pb and 8.2 percent Zn, was produced (Hoy, 1980). Renewed production during the period 1952–1971 yielded 4,777,000 t, grading 5.1 percent Pb, 6.1 percent Zn, 1–2 oz/t Ag, 0.1 percent Cu, and 0.0003 percent Cd (Changkakoti et al., 1988). There are three sulfide-rich zones; from north to south these are the Comfort, Bluebell, and Kootenay Chief zones (Fig. 4).

Structural setting of the Bluebell deposit

The Riondel peninsula is underlain by a west-dipping panel of penetratively deformed rocks belonging to the Hamill Group, the Mohican and Badshot formations, and the Lardeau Group (Hoy, 1980). The deposit is hosted in carbonate strata of the Lower Cambrian Badshot and Mohican formations. Stratigraphy of the mine site is outlined in Table 1.

These units lie in inverted stratigraphic sequence due to their position on the overturned lower limb of the Riondel nappe, a recumbent isoclinal F_1 fold with an amplitude of >20 km (Fig. 3). The F_1 Riondel nappe is refolded by a series of tight-isoclinal F_2 folds that plunge gently north and are overturned to the east. The penetrative S-L fabrics developed in the rocks on the Riondel peninsula are associated with the development of these F_2 folds. S_1 is only rarely preserved in quartz-rich layers in the hinges of some minor F_2 folds.

TABLE 1. Stratigraphy Around the Bluebell Mine Site (after Shannon, 1970).

Thickness (m)	Stratigraphic unit	Lithology
120	Hamill Group-Mohican Formation	White quartzite, brown mica schist and pegmatite; a marble lens (“upper limestone” max. thickness of 5.5 m) 24 m above the Badshot Fm. contact marks the base of the Mohican Formation
30-45	Badshot Formation	Mostly white, alternating fine- and coarse-grained calcite marble with some gray bands and phlogopite partings; some coarse-grained gray layers are dolomitic
215	Lardeau Group	Graphitic gray and black schistose argillite or argillaceous quartzite; beds are partly calcareous and somewhat feldspathic; locally, lenses of impure marble
150	Lardeau Group	Hornblende schist with some interlayered quartz mica schist and limy schist
>490	Lardeau Group	Quartz-calc-silicate schist, feldspathic and calcareous in places, with amphibolite and pegmatite sills, locally lenses of impure marble

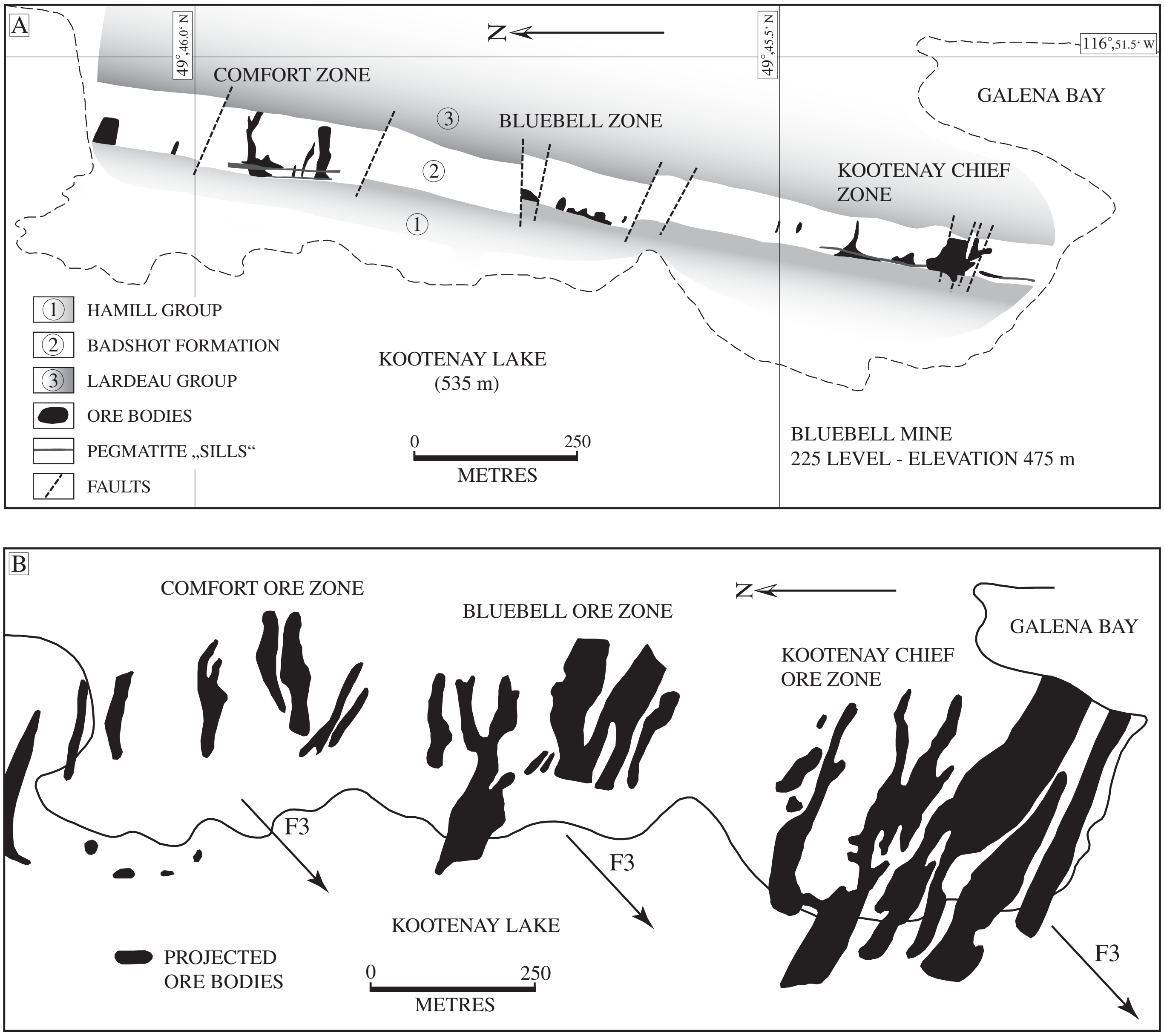

FIG. 4. A) Plan of the 225 level of the Bluebell mine. The outline of the Riondel peninsula is also shown. Adapted from Irvine (1957). B) Plan of the Bluebell mine showing the projection of orebodies onto the surface, from Shannon (1970).

S_2 dips 20° to 60° to the west and is generally subparallel to compositional layering (S_0) except in the hinges of minor folds (Figs. 5A, 6). Due to the convex-eastward curvature of the Kootenay Arc, average S_2 swings from south-southwest south of the peninsula to south-southeast to the north (Hoy, 1977, 1980). Minor F_2 folds are common; they plunge at low angles to the north, are tight to isoclinal, and have rectilinear hinges. Minor F_2 fold axes and intersections of S_2 with S_0 are invariably colinear with a stretching lineation, L_2 (Figs. 5A, 6). This is defined by stretched quartz crystals and quartz-feldspar aggregates in metasedimentary rocks, veins, and pegmatites; aligned phyllosilicates and sillimanite-rich nodules in micaceous schists, and aligned tourmaline crystals in quartz veins. It lies orthogonal to the necks of calc-silicate boudins in quartzite on S_2 (Fig. 5B). Internal boudinage is also developed within schistose layers and locally gives rise to discontinuous undulations of S_2. The alignment of sillimanite crystals and sillimanite-rich nodules with L_2 means that peak metamorphism preceded or was synchronous with D_2 deformation.

The youngest folds in the area (F_3) are open deflections of S_2/S_0 that plunge southwest (Fig. 6; Fyles, 1967; Livingstone, 1968; Hoy, 1980; this study). These folds and later structures are described in detail in a later section.

Nature and geometry of the deposit

The Bluebell orebodies, first described in detail by Irvine (1957), comprised three zones separated by barren intervals approximately 300 m wide (Fig. 4). The bodies were massive sulfide replacements that developed in marble along steeply dipping fractures (Irvine, 1957). The ore formed tabular

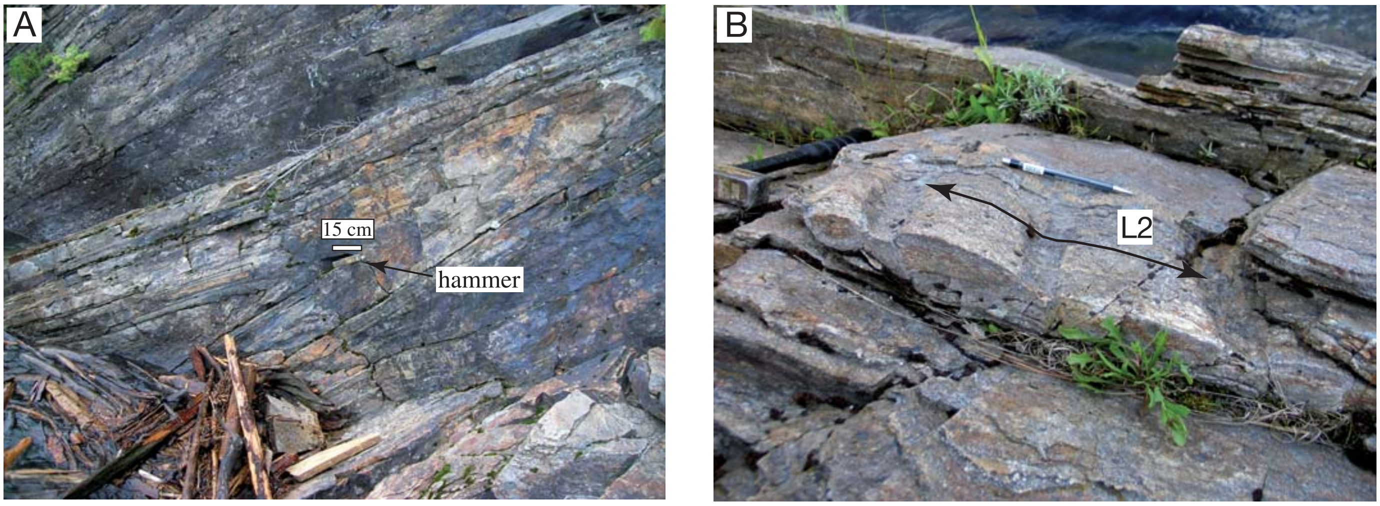

FIG. 5. D_2 (Lower Cretaceous) structures at Riondel. A) Example of an isoclinal fold with thickened hinge zone. The dominant west-dipping schistosity in Riondel (S_2) is axial-planar to these folds, and the stretching lineation is parallel to the gently plunging fold axes. Hammer head for scale. B) Boudinage of a calc-silicate layer. The boudin neck runs approximately down the dip of the layer, perpendicular to the shallow-plunging stretching lineation (L_2). Pencil for scale.

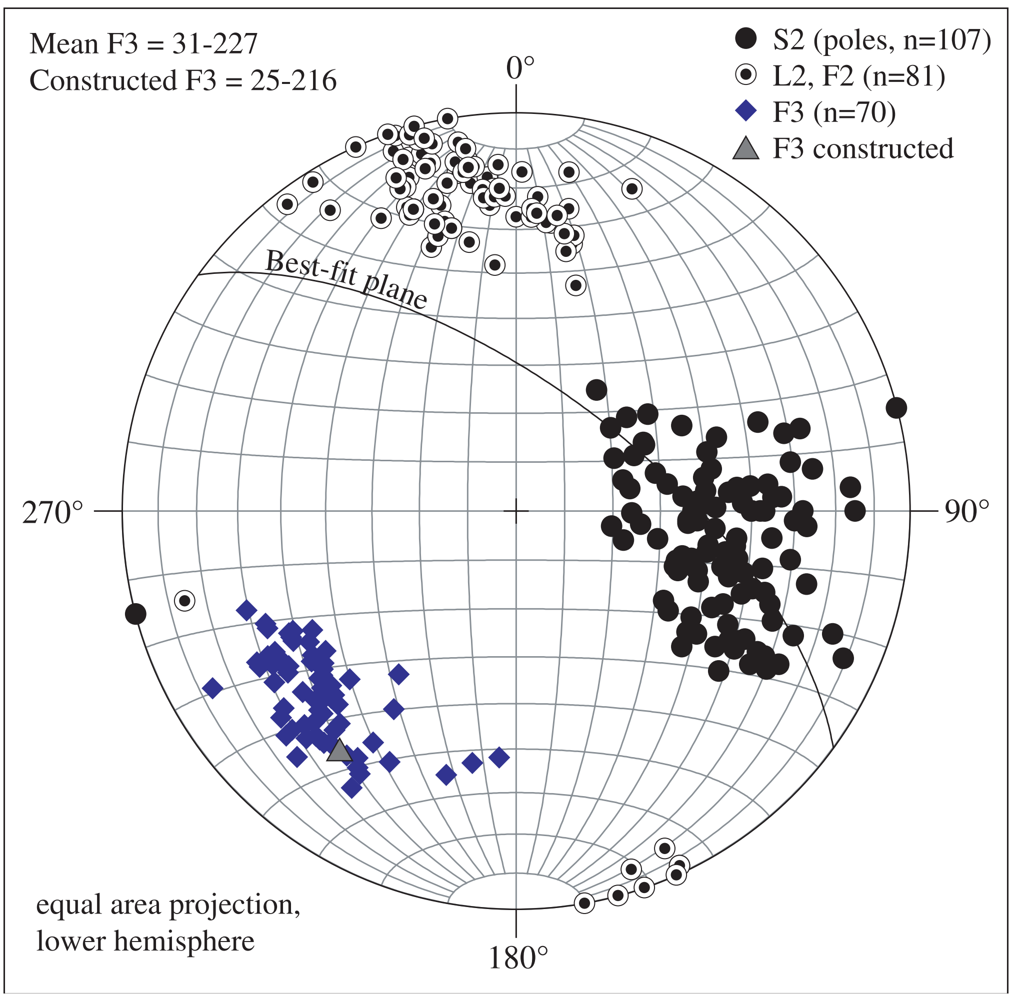

FIG. 6. Equal area lower hemisphere stereonet showing the orientation of ductile structures on the Riondel peninsula. S_2 and L_2 are gently folded around southwest-plunging F_3 folds. "L_2" includes minor F_2 fold axes, S_2/S_0 intersection lineations and L_2 mineral lineations as these are invariably parallel to one another in this area.

bodies with irregular outlines that were controlled by variation in the extent to which it spread along layers intersected by the fractures (Irvine, 1957). Ore typically extended 1.5 to 3 m outward from fractures; however, where fractures intersected favorable horizons, ore extended up to 30 m along the layer. Larger bodies up to 30 m wide were also produced in places where ore from adjacent, closely spaced fractures coalesced. The orebodies extended downdip as much as 500 m (Irvine, 1957).

The sulfide bodies plunged west-northwest, parallel to the intersection of fractures with stratigraphic layering. The mineralized fractures strike west-northwest and dip steeply north. The attitudes reported by Irvine (1957) for each of the ore zones are reproduced in Table 2; the small differences in fracture orientations between the ore zones result from abrupt changes across zones of brecciated rock that are up to 15 m wide (Irvine, 1957). The breccia zones are planar, approximately parallel to the mineralized fractures.

In map view, the sulfide bodies formed elongate bodies trending west-northwest (Fig. 4). In cross section, Ransom (1977) described the idealized average body as being mushroom shaped, with crosscutting keels 1 to 30 m wide, widening upward into a cap up to 6 m thick that extended as much as 50 m from the keel zone. Some lateral shoots were also developed close to the base of the marble.

Ore was most extensively developed at the upper contact of the Badshot Formation, at the contact between fine- and coarse-grained marble within the formation, below schist within the formation, and on the underside of granitic pegmatite and mafic dikes (Fig. 7; Irvine, 1957; Shannon, 1970; Ransom, 1977; Hoy and Ransom, 1981). Depressions on the footwall and antiforms on the hanging wall were also favorable sites for ore accumulation (Ransom, 1977). Ore accumulated in these locations because they each presented an impediment to upward movement of mineralizing fluids (Ransom, 1977).

Pegmatite dikes are deformed and are typically approximately parallel to stratigraphic layering, whereas mafic dikes crosscut the west-dipping foliation (S_2). Two kinds of mafic dikes have been recognized. Deformed brown dikes (Shannon, 1970), referred to as lamprophyre by Ransom (1977), strike north-south, dip east, and display normal-sense offsets parallel to west-dipping S_2. Younger dikes, which have been referred to as diabase (Ransom, 1977), lamprophyre, and greenstone (Shannon, 1970) typically trend west-northwest and postdate ductile deformation and regional metamorphism. Altered gabbroic dikes of this type comprise partly altered plagioclase phenocrysts sitting in a plagioclase-rich matrix that has been partly altered to white mica, carbonate, and opaques (Beaudoin et al., 1992a). These dikes probably belong to the Eocene suite developed regionally but this has not been confirmed.

TABLE 2. Orientation of the Mineralized Fractures in Each of the Three Ore Zones (from Irvine, 1957).

Ore zone	Strike	Dip
Comfort	N72°W (288°)	83°N
Bluebell	N76W (285°)	82°S
Kootenay Chief	N63W (298°)	85°N

A number of faults transect the Badshot Formation (Fig. 4A). These faults have strike separations of up to 10 m and are generally oriented parallel to mineralized fractures (Irvine, 1957). In the Kootenay Chief and Bluebell ore zones these faults are concentrated in the central part of the ore zone (Irvine, 1957). Shannon (1970) described mineralization along faults, suggesting faulting predated mineralization. Elsewhere, faults appear to truncate ore (Fig. 4A) and deformed galena crystals observed during the current study indicate that there was local deformation after crystallisation of the ore.

The mineralization at Bluebell is dominated by massive replacement of marble. These coarse-grained masses consist mainly of pyrrhotite, sphalerite, galena, knebelite (Fe-Mn olivine), quartz, and calcite (Westervelt, 1960; Ohmoto and Rye, 1970). Pyrite is present as a replacement product of pyrrhotite but some may be primary (Shannon, 1970). Small amounts of arsenopyrite, chalcopyrite, siderite, and rhodochrosite are also present. Approximately 10 percent of the sulfide mineralization is contained in partially filled vugs, where pyrrhotite, sphalerite, galena, and minor chalcopyrite are intergrown with quartz and calcite (Ohmoto and Rye, 1970).

Ohmoto and Rye (1970) identified three stages in the mineralization, starting with formation of knebelite (period I), followed by massive sulfide-quartz-carbonate ores (period II), and finally, formation of crystals in vugs (period III). The boundary between periods II and III is arbitrary, with a gradual increase in the abundance of calcite and quartz relative to sulfides. Based on mineral assemblages and fluid inclusions, Ohmoto and Rye (1970) estimated temperatures of 320° to 450°C during deposition of period III crystals, at a depth of approximately 6 ± 2 km; earlier stages formed at temperatures >450°C.

The vuggy, space-filling nature of some of the ore, its association with late crosscutting fractures, its development in marble belonging to different stratigraphic units, and the relationship with unmetamorphosed mafic dikes collectively provide compelling evidence that Bluebell is a late, fracture-controlled replacement deposit (Irvine, 1957; Fyles, 1967; Ohmoto and Rye, 1970; Shannon, 1970; Ransom, 1977; Hoy, 1980; Hoy et al., 2000a). Nelson (1991) and Hoy et al. (2000a) noted the similarity of Bluebell to manto-type deposits.

Although the control exerted by the west-northwest–trending fractures in localizing the deposit has been well established, the relative age and origin of the fractures is less clear. Some previous authors have suggested fractures formed during and as a result of folding (Irvine, 1957; Shannon, 1970; Hoy, 1980; Hoy et al. 2000a). In the following section we give a detailed account of late structures on the Riondel peninsula and suggest that these interpretations are incompatible with the observed geometry. Instead, we suggest that the folds and fractures formed sequentially.

Late Folds and Fractures on the Riondel Peninsula

F_3 folds

Southwest-plunging (F_3) folds on the Riondel peninsula were reported by Fyles (1967), Livingstone (1968), and Hoy (1980). F_3 folds are symmetric to weakly asymmetric gentle to open southwest-plunging folds with steep axial planes (Figs.

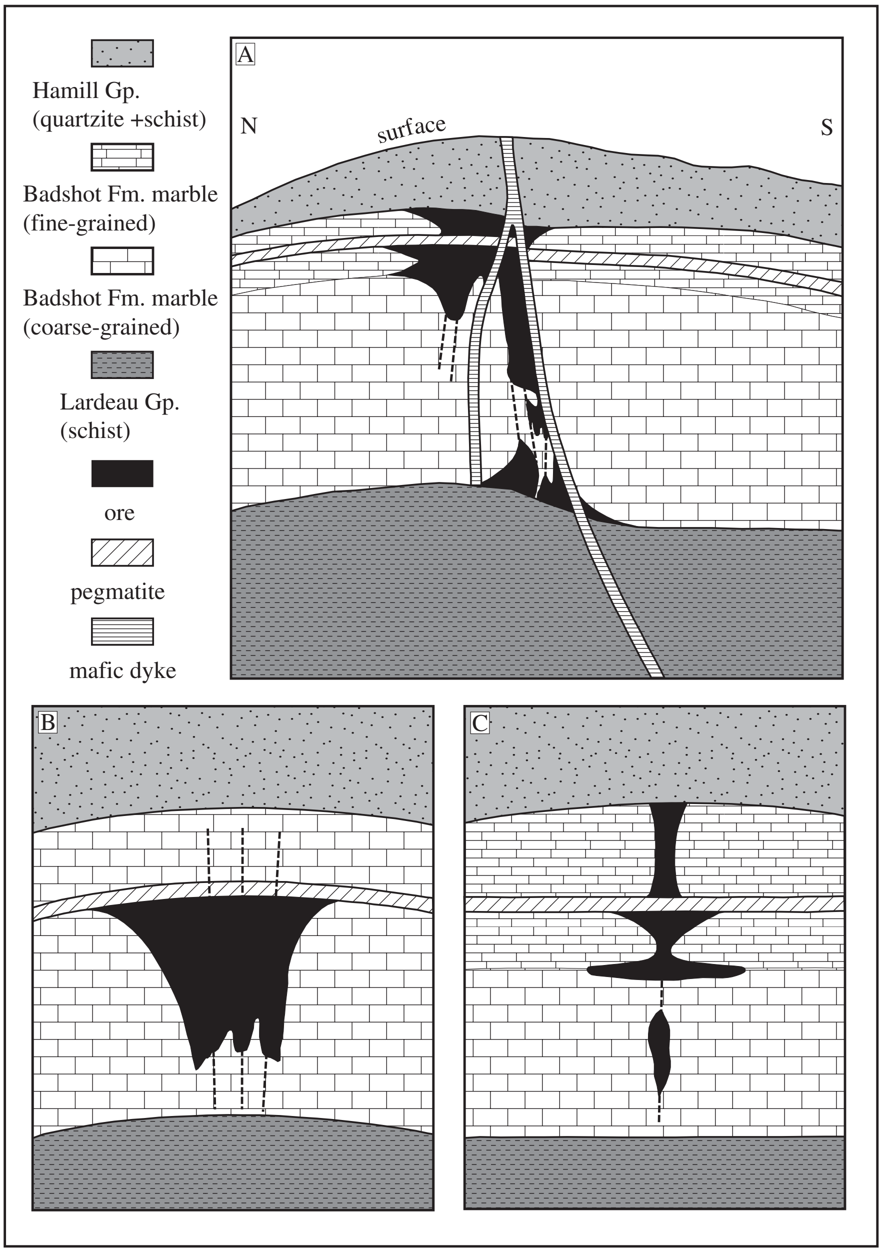

FIG. 7. Examples of longitudinal profiles through Bluebell orebodies. These show the influence on ore location exerted by fractures (A), mafic dikes (A), pegmatite (B, C) and lithology (A, C). From Shannon (1970).

6, 8A). Where asymmetry is discernible, folds display S-shapes when viewed down plunge. Folds are prominent at the meter and centimeter scale; the only map-scale fold with this orientation is the Sherraden Creek fold to the south of the Riondel peninsula (Hoy, 1980). Livingstone (1968) noted that L_2 lineations around individual F_3 folds fall on small circles and suggested these folds formed by flexural slip. Hoy also proposed this mechanism and noted the S asymmetry, interpreting this to reflect a "sinistral shear sense" (Hoy 1980, p. 67). This interpretation of the asymmetry may be valid if the folds are modified buckles or drag folds, but a different origin is suggested here.

A notable characteristic of F_3 folds is that they are highly discontinuous. Rather than forming elongate linear features, they impart an irregular wavy character to folded surfaces (Fig. 8B). In addition to being laterally discontinuous, folds are discontinuous normal to S_2/S_0, leading to disharmony between adjacent layers. On suitably exposed surfaces the folds appear as slightly elongate, scoop-shaped undulations that commonly have length/width ratios of only 2 or 3:1. The discontinuous nature of these folds can make precise measurement of individual fold hinges difficult.

The key to the interpretation of F_3 folds presented here is the recognition of southwest-trending centimeter-scale shear bands that transect S_2 in schistose rocks of the Hamill Group (Fig. 8C). These centimeter-scale shear bands are discontinuous in each direction and commonly merge with S_2 at their terminations. They dip northwest and cause discontinuous

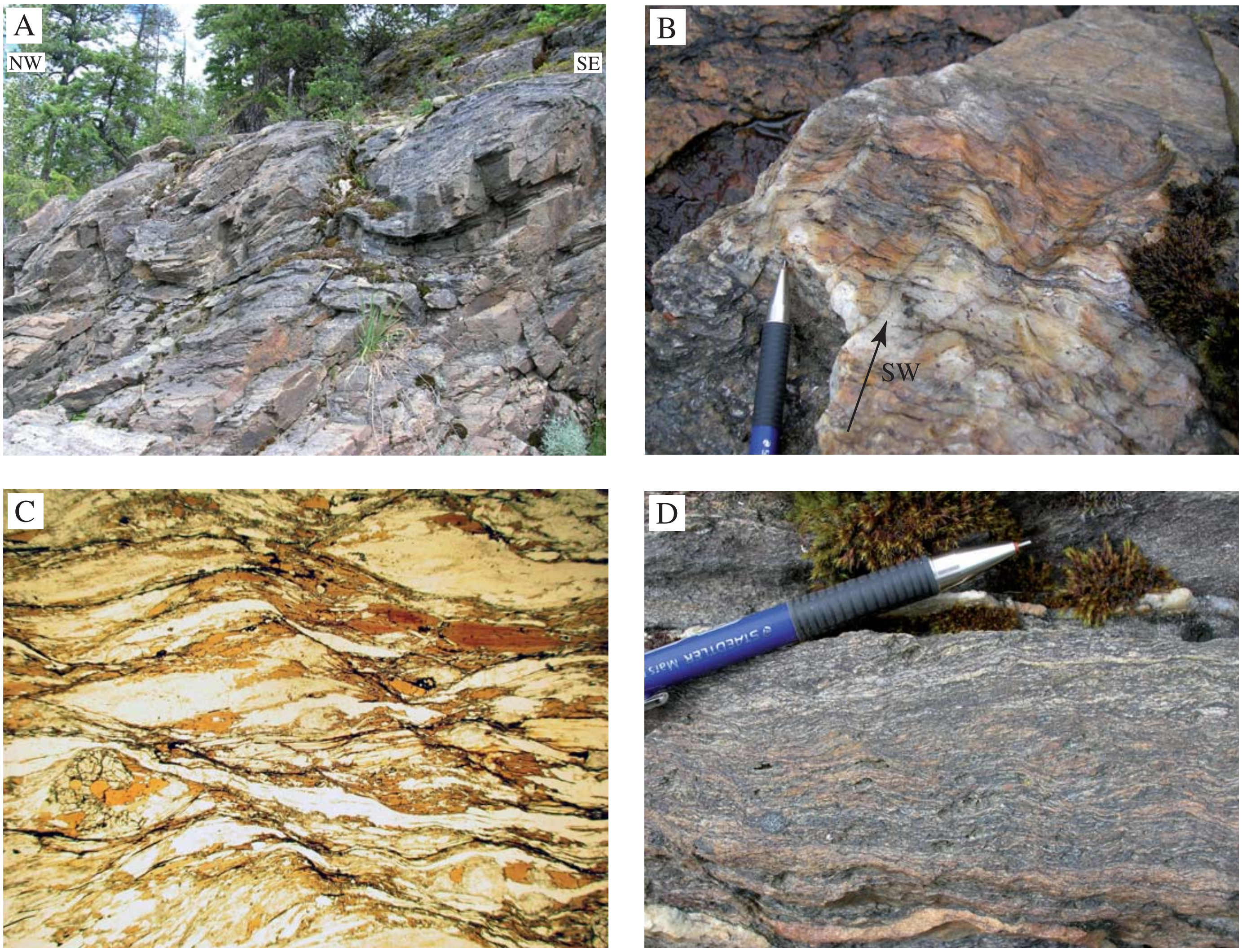

FIG. 8. F_3 structures on the Riondel peninsula, developed at a range of scales. A) Open, sigmoidal southwest-plunging F_3 folds with steep axial planes. This photograph is taken looking up the plunge of the folds. Hammer for scale. B) Irregular undulations on a quartz vein caused by F_3. C) Photomicrograph of shear bands in garnet-sillimanite schist of the Hamill Group. The field of view is 12 mm. Viewed normal to the axes of F_3 folds. D) Incipient shear bands developed in Hamill Group semipelite. The viewing direction is close to the axes of F_3 folds.

sigmoidal deflections of S_2 that define open folds with southwest-plunging axes (Fig. 8C, D).

Shear bands are variably developed. In some instances, they form narrow shear zones with relatively discrete offset of layers and deflection of S_2. Elsewhere the shear bands are broad and diffuse (Fig. 8D); in these cases, they are manifested in the thinning of the northern limbs of slightly asymmetric sigmoidal antiforms. In each case, slightly asymmetric discontinuous folds with southwest-plunging axes were produced as a result of shear-band development. Meter-scale folds exposed on the shoreline have a similar geometry to those observed at the centimeter scale. These are diffuse shear bands developed on a larger scale.

This geometry is interpreted to result from heterogeneous development of shear band cleavage (Platt and Vissers, 1980; White et al., 1980; Dennis and Secor, 1987; Williams and Price, 1990), and indicates reactivation of, and a component of extension along, S_2. The orientation and asymmetry of the shear bands and associated folds indicate oblique dextral-normal sense shearing during D_3, and the discontinuous nature of the folds observed at Riondel is characteristic of these structures (Passchier and Trouw, 2005). Southwest-plunging F_3 folds at all scales (Fig. 8) are interpreted to have formed in a similar manner, based on the reasoning that folds over a range of scales with common orientations and characteristics most likely formed at the same time, by the same mechanism. There is no evidence for shortening of S_2 during or subsequent to D_3; buckle folds are not developed.

Shear band cleavage fabrics and associated folds are widely developed in schistose rocks in the footwall of the Gallagher fault, the structurally highest of the early Tertiary normal faults on the west side of Kootenay Lake (Fig. 3; Fyles, 1967). These structures are best developed in the immediate footwall of the fault and decrease in prominence eastward (Moynihan and Pattison, 2008). An eastward decrease in the intensity of F_3 folding was also noted by Livingstone (1968).

As F_3 structures are restricted to the footwall of the fault, they must be at least as old as the normal faulting. In thin section, shear bands at Riondel are characterized by concentrations of opaque material; however, elsewhere in the footwall of the Gallagher fault, similar shear bands are rich in secondary chlorite, suggesting development during retrograde metamorphism following peak metamorphism and D_2 deformation.

The absolute age of phase III folding is poorly constrained. It postdates Early to mid Cretaceous peak metamorphism and D_2 deformation (Leclair et al., 1993; Moynihan and Pattison, unpub. data), but is older than mafic dikes of presumed, but unconfirmed, Eocene age. Livingstone (1968) reported transection of F_3 folds by medium- to fine-grained leucrocratic dikes with chilled margins. The age of these dikes was not established, but granitic intrusions younger than approximately 75 Ma (Campanian) have not been recorded in the Kootenay Arc (Logan, 2002, and references therein). Early Tertiary muscovite (55±3, 59±3 Ma) and biotite (50 ± 3 Ma) K-Ar ages from the Riondel peninsula (Beaudoin et al., 1992a) are similar to those elsewhere in the footwall of the Gallagher fault and record cooling that was probably coincident with regional extension (Archibald et al., 1984).

Late subvertical shear zones

S_2 is overprinted by a number of meter-scale subvertical ductile shear zones trending approximately 340°. Deflection of S_2/S_0 into these shear zones indicates west side-down displacement. It is not clear whether these shear zones formed before, during or after F_3 folding; however, the shear zones predate fracturing, mafic dike intrusion, and mineralization.

Brittle fractures

Quartzite of the Hamill Group is extensively fractured. Based on surface measurements taken west of the Comfort and Bluebell ore zones, there are two dominant sets of fractures; a third is also locally developed (Figs. 9, 10). Fractures belonging to each of the two main sets are generally visible together in outcrop (Fig. 11A), with a typical spacing of centimeter-decimeter. They have smooth segments but are laterally discontinuous, with numerous jogs between subparallel

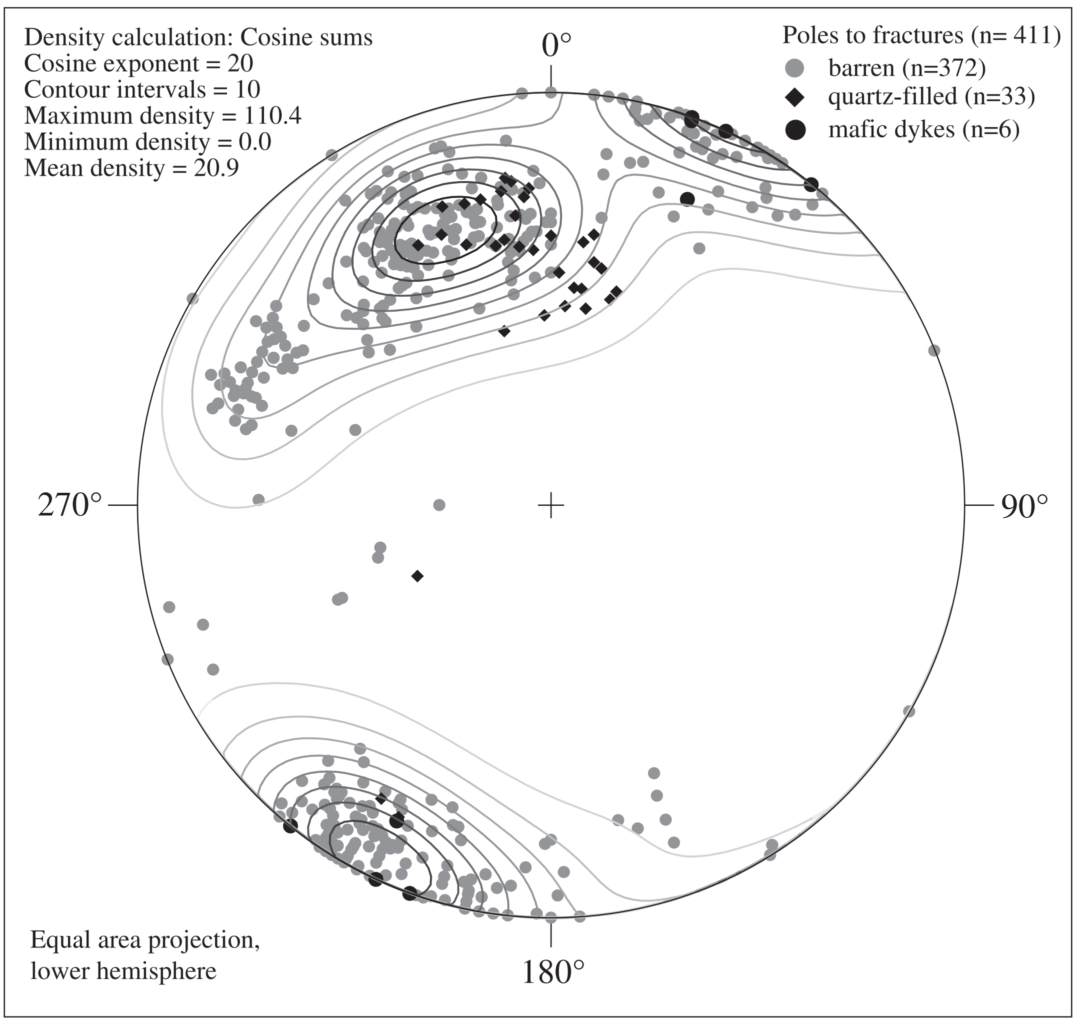

FIG 9. Equal area lower hemisphere stereonet showing contoured poles to brittle fractures in the Hamill Group (hanging wall), directly to the west of the Comfort and Bluebell ore zones. There are two dominant orientations. The fracture set trending west-northwest is mineralized and hosts mafic dikes. Fractures trending east-northeast are commonly filled with quartz. A third set of fractures, striking north-northeast, is locally developed.

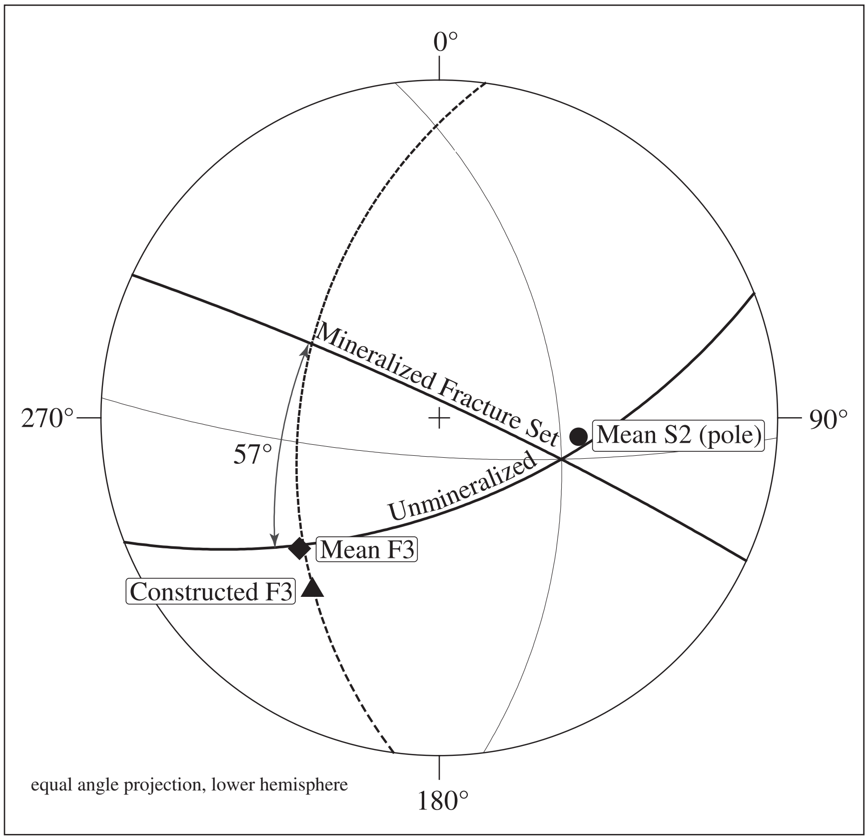

FIG. 10. Synoptic equal angle lower hemisphere stereonet showing the geometric relationship between F_3 folds and brittle fractures. The orientations of the two dominant fracture sets are plotted using the values with the greatest concentration of field measurements. The acute and obtuse bisectors of these fracture sets are also shown. The acute dihedral angle between the two sets of fractures is 57°. Mean S_2 is marked with a black circle. Mean measured F_3 is shown using a diamond shape. The triangle indicates the orientation of the F_3 axis constructed from all S_2 measurements.

surfaces and slight angular changes in the orientation of individual surfaces. The dominant fractures commonly belong to the set clustered around an attitude of 295°/85° N (Figs. 9, 10). This is the orientation of the fractures that host mineralization in each of the three ore zones (Irvine, 1957; Fyles, 1967; Table 2). A small percentage of fractures with this orientation

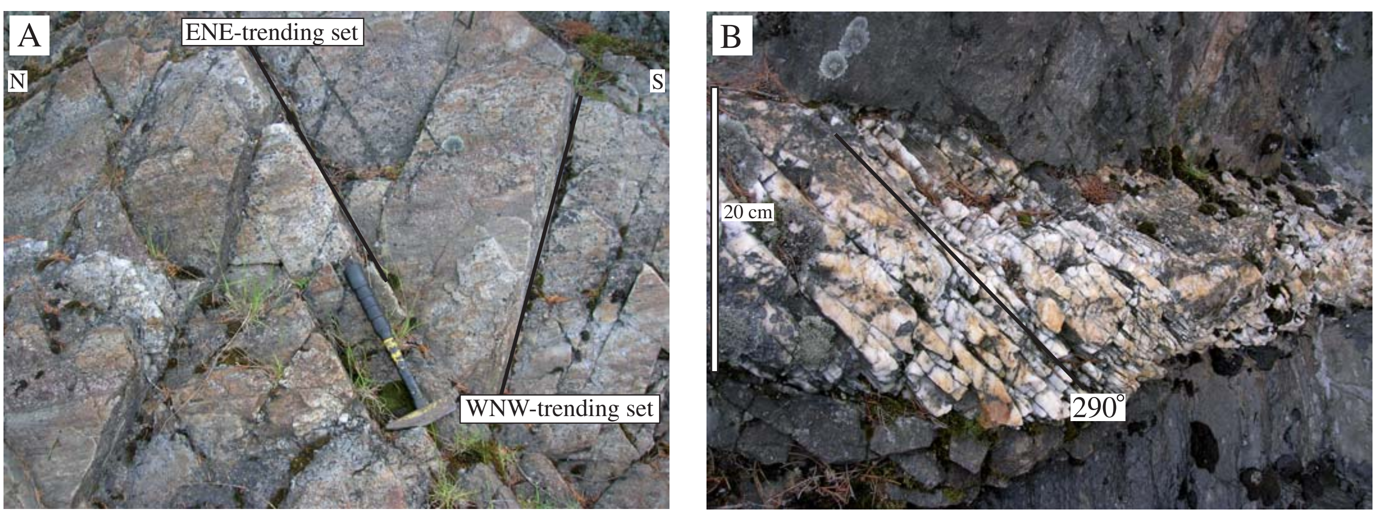

FIG. 11. A) Photograph of the two dominant sets of fractures in quartzite of the Hamill Group on the Riondel peninsula. Looking east onto S_2/S_0. Hammer for scale. B) Quartz vein occupying southeast-dipping fracture cut by west-northwest–trending fractures.

have macroscopic quartz fillings. The quartz in these fractures retains delicate growth textures, indicating growth into open space with no subsequent recrystallization.

The second prominent cluster of fractures is concentrated around a strike of 68°/60° S (Figs. 9, 10). Quartz fillings are much more common in these fractures compared with those trending west-northwest. Mineralization along fractures with this orientation has not been reported, but at UTM 11U 0509865 5512241, coarse-grained galena is intergrown with quartz in a vein with this orientation. Space-filling textures are preserved and the mode of occurrence suggests this local concentration of galena formed through remobilization from the main ore deposit. There is a particularly high concentration of thick quartz veins with this orientation directly above the downdip projection of the Bluebell ore zone. Some of the quartz veins filling fractures with this orientation are cut by fractures belonging to the west-northwest–trending (mineralized) set (Fig. 11B).

All of the mafic dikes observed on the shoreline of the Riondel peninsula are oriented parallel to the west-northwest–trending (mineralized) fracture set. They cut obliquely across F_3 folds and truncate north-northwest–trending subvertical west side-down shear zones. At UTM 11U 0509849 5512370, a west-northwest–trending dike jogs onto a fracture belonging to the east-northeast–trending (unmineralized) set for a short segment. These relationships suggest that dikes were intruded along fractures that formed before or during emplacement. One of the dikes studied is cut by an array of fractures belonging to the east-northeast–trending (unmineralized) set.

There is generally little or no displacement visible across the measured fractures, but they locally have offsets of up to a few centimeters. At the location of intersections between the two sets of fractures, various relationships are apparent, including simple crossing with no measurable offset, deflection of one fracture set close to the intersection line, termination of one fracture against another, and apparent offset of one of the main throughgoing fractures, sometimes with shorter parallel fractures developed around the intersection line. The two main fracture sets are interpreted as having formed as a conjugate set of shear fractures based on mutually crosscutting relationships and small spatially coincident variations in the orientation of each set. If so, the ore-hosting set must have been preferentially opened before or during mineralization; these fractures also localized late faulting.

Discussion

Did mineralized fractures form during and as a result of folding?

Irvine (1957) proposed that formation of the mineralized fractures at Bluebell accompanied folding. According to this interpretation (Irvine 1957, p. 103), north-south shortening produced “cross-warps of the strata,” including the “gentle syncline surrounding the Bluebell mine” with a wavelength of 5 km. The “hanging-wall quartzite, acting as a unit with the underlying limestone, broke into segments, each of which...lifted as a gentle anticlinal arch.” As a consequence, “a series of anticlinal arches formed within the confines of a synclinal fold.” Arching produced “tension fractures....oriented parallel to the fold axes.” This compressional period was then followed by “one of relaxation, during which the anticlinal arches sustained some degree of collapse along gravity faults which formed from some of the tension fractures” (Irvine, 1957, p. 103).

There are a number of problems with this interpretation. There is no evidence that there was north-south shortening. Mapping of the region around the mine site has confirmed the presence of a gradual change in the strike of S_2, from the south-southwest south of Riondel to south-southeast to the north (Crosby, 1968; Hoy 1977, 1980); however, this “fold” is not a local feature with a wavelength of just 5 km; it reflects the curvature of the Kootenay Arc, a structural salient with a wavelength of >275 km and an amplitude of 80 km (Fig. 1). Varsek and Cook (1994) attributed formation of the arcuate salient to the presence of a cratonic embayment to the east, whereas Thompson et al. (2007) suggested that the shape reflects the subsurface geometry of a basement high to the west. This salient formed during Middle Jurassic-Lower Cretaceous deformation, when rocks were undergoing north-south extension at middle to upper amphibolite facies conditions. A later fold generation with westerly plunging axes has not been recognized in the area (Crosby, 1968; Livingstone, 1968; Hoy, 1977, 1980) and the map of the deposit does not show each of the deposits located on antiformal crests (Fig. 4A). Buckling and outer arc extension would not be expected in a weak carbonate layer surrounded by quartz-rich rocks. The interpretation does not account for the observation that “fractures are spaced uniformly along the limestone formation (Irvine, 1957, p. 103), nor does it explain the second set of fractures that form an angle of approximately 60° with the mineralized set.

Shannon (1970) adopted the interpretation proposed by Irvine (1957) and presented a block diagram to illustrate the geometry of the deposit (Fig. 12); this was also reproduced in Nelson (1991). In this illustration, the ore zones are shown occupying the crestal region of three symmetric antiforms plunging west-southwest, parallel to the strike of fractures and mafic dikes. A similar diagram in Ohmoto and Rye (1970) and Ohmoto (1971) shows fanning of fractures around antiformal arches centered on each of the ore zones. These diagrams depict the overall form of the deposit adequately, but details are misleading. Fractures and dikes trend west-northwest, not west-southwest. Mineralized fractures consistently dip north rather than fanning around ore zones. The folds depicted are not evident on the map, and the southwest-plunging F_3 folds are not represented in any way. The interpretation implicit in each of the diagrams that outer-arc extension led to formation of the fractures in the marble is not supported by the data presented above.

An alternative proposal regarding the timing of fracture formation was made by Hoy (1980, p. 85; also Hoy et al., 2000a; Hoy and Lefubre, 2003), who suggested that “the fractures and related mineralization may have developed during the Phase 3 deformation.” However, the fractures cut across and show no variation between F_3 hinge zones, F_3 limbs, and areas unaffected by F_3 folding. They do not exhibit any of the geometric relationships with respect to fold axes that typically develop during simultaneous folding and fracturing (Price and Cosgrove, 1990). According to the interpretation presented here, the formation of F_3 folds involved extension and

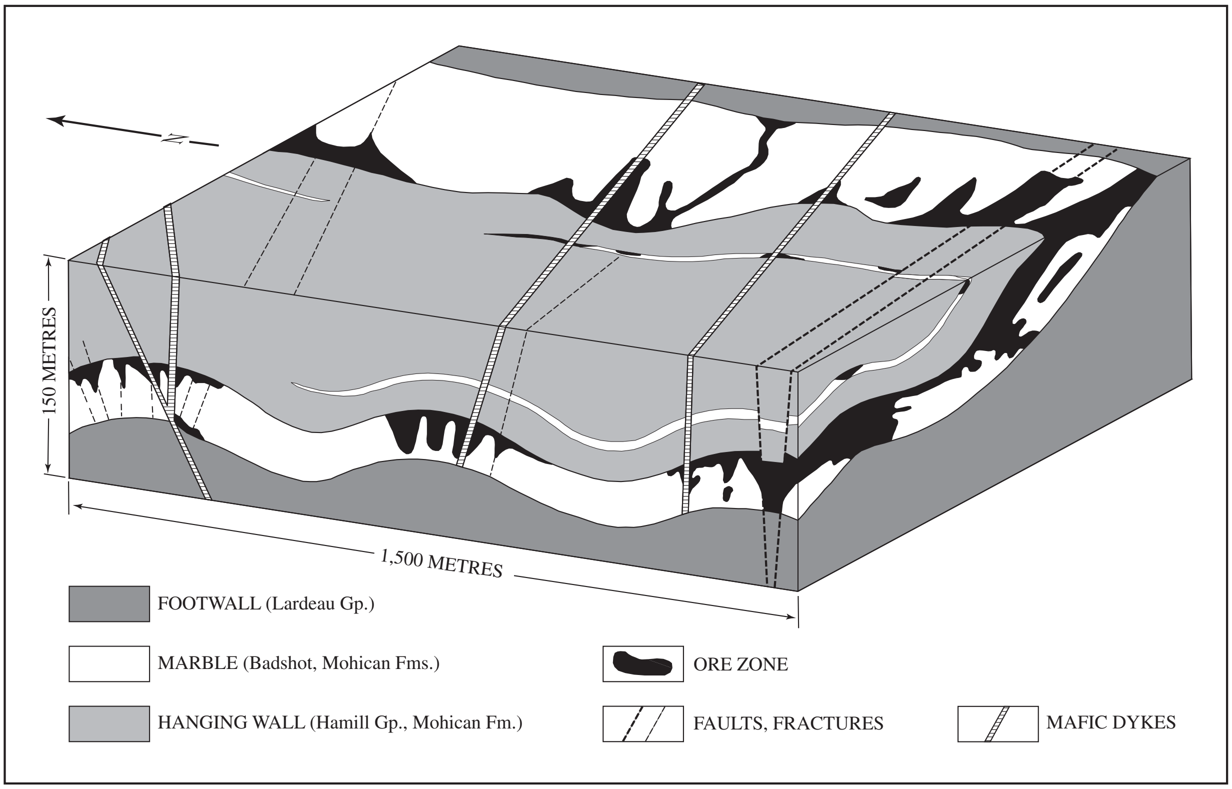

FIG. 12. Block diagram of the Bluebell mine area, reproduced from Shannon (1970). Similar diagrams appear in Ohmoto and Rye (1970), Ohmoto (1971), and Nelson (1991). Three antiforms are shown, with mineralization developed in the outer arcs of antiforms, parallel to fractures. Note the vertical exaggeration. This geometry is not supported by data presented here. See discussion in text.

slip along S_2, and the formation of diffuse shear zones; this deformation must have taken place prior to formation of fractures. We interpret the brittle fractures as having originated as a conjugate set of shear fractures, one of which was preferentially opened and mineralized. Irrespective of whether or not this is the case, we suggest that each of the fracture sets formed after F_3 folding. This result is identical to that of Fyles (1967), who refuted the suggestion that mineralized fractures on the Kootenay-Lakeshore properties (opposite the Riondel peninsula) formed in response to folding.

The interpretation that folding played no role in formation of the fractures does not contradict observations made by Irvine (1957), Shannon (1970), and Ransom (1977) that undulations, particularly in the hanging wall, were favorable sites for ore formation. Preexisting folds of impermeable surfaces formed structural traps, passively inhibiting the passage of fluids, thereby localizing ore formation. It is likely that local antiformal structures that host ore (Irvine, 1957; Shannon, 1970) reflect undulations in layering resulting from a combination of D_2 symmetric boudinage and particularly F_3 asymmetric warping. The scale and discontinuous nature of F_3 is reflected in the "dimpled" (Shannon, 1970, p. 117) character of layering and the absence of large deflections of stratigraphic contacts on the Riondel peninsula (Fig. 4A).

Implications for formation of the deposit

At Bluebell, some or all of the mineralization postdated intrusion of undeformed mafic dikes. However, in the broader Ainsworth-Riondel area there is temporal overlap of vein-type mineralization and of mafic dikes-sills intrusion (Fyles, 1967). In the Ainsworth area, on the west side of Kootenay Lake, lamprophyre dikes lie parallel to foliation or in the same steeply dipping fractures as mineralized veins. Sills and dikes are locally mineralized and are fractured and offset by the vein faults (Fyles, 1967). As stated by Fyles (1970, p. 53), "mafic dikes in the Bluebell and Ainsworth areas that follow the same pattern of fracturing as the veins show that this mineralizing process is closely associated with magmatic activity." Abundant systematic fractures in the rocks at Bluebell and elsewhere provided conduits for fluid movement into the geochemical trap provided by the carbonate rocks. The high spatial density of deposits and early Tertiary intrusions in the Ainsworth-Bluebell area presumably reflects enhanced development of fracture networks in the area. Development of this fracture network took place independently of earlier folding processes.

Beaudoin and co-workers (Beaudoin et al., 1991, 1992a) linked alkaline magmatism and vein-hosted Ag-Pb-Zn-Au mineralization with early Tertiary extension in southeastern British Columbia. According to this interpretation, mantle upwelling during extension led to adiabatic partial melting of subcontinental mantle, generating lamprophyre and gabbro dikes at approximately the same time as mineralization. Beaudoin et al. (1991) noted the requirement for a large transcrustal fault zone to connect deep-seated Pb, C, and possibly O reservoirs with upper crustal levels and suggested this role was played by the Eocene Slocan Lake normal fault. The normal faults west of Ainsworth were also invoked as conduits for mineralizing fluids.

A complicating factor is that recent interpretations (Cook and Van der Velden, 1995; Carr and Simony, 2006) suggest the Slocan Lake fault does not penetrate the crust, as proposed by, for example, Cook et al. (1992). In additions, veins, fractures, and faults that host ore deposits in the Ainsworth area are younger than some, and probably all of the Ainsworth normal faults. The Josephine fault is crosscut by fractures of the Highland vein system (Fyles, 1967). The Gallagher fault zone displays a combination of brittle and ductile structures that must have formed prior to cooling of the rocks and formation of open transverse fractures. It is likely that the Lakeshore fault, which is represented by a zone of "highly sheared" rock (Fyles, 1967) is also older than late, open fractures. Whatever the role played by normal faulting in generating conditions that led to formation of ore deposits, timing relationships suggest the Ainsworth faults were not direct fluid pathways during mineralization.

Another potential route for passage of Pb and CO_2 from deep levels is along the multiply reactivated transverse basement structure that crosses Kootenay Lake around the latitude of Riondel/Ainsworth (McMechan, 2010). The suggestion (McMechan, 2010) that this deep structure affected mineralization in the overlying Bluebell and Lakeshore deposits is supported by the finding of Beaudoin et al. (1992a) that Pb in these deposits came from a deeper source when compared with other vein-type deposits in areas to the northwest of the basement structure. This zone may have facilitated passage of mafic magma and hydrothermal fluids from deep reservoirs. If so, similar fracture networks developed elsewhere above this basement zone are prospective sites for hydrothermal mineralization.

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