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3 **Styles, Textural Evolution, and Sulfur Isotope Systematics of Cu-rich**
4 **Sulfides from the Cambrian Whalesback Volcanogenic Massive Sulfide**
5 **(VMS) Deposit, Central Newfoundland, Canada**

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Abstract

The Whalesback Cu-rich VMS deposit in the Newfoundland Appalachians is a highly deformed deposit found on a steep limb of a closed and boudinaged overturned fold. The deposit was intensely deformed at low temperature but medium pressure (>175 MPa) during the accretion of the composite Lushs Bight Oceanic Tract-Dashwoods terrane onto the Humber margin at ca. 480 Ma.

The ore mineralogy consists of chalcopyrite, pyrrhotite and pyrite with lesser sphalerite and trace Ag-, Bi- and Hg-tellurides. Four styles of sulfide mineralization are present: 1) disseminated (5%); 2) vein (50%); 3) breccia (25%); and 4) semi-massive to massive (20%). Independent of mineralization style, massive pyrite and pyrrhotite (and some chalcopyrite) are commonly parallel to main S_2 schistosity in the deposit, whereas late chalcopyrite piercement veins occur at a high angle to S_2 . The progressive increase in pressure and temperature produced a remobilization sequence wherein sphalerite was the first sulfide phase to cross the brittle-ductile boundary, followed by pyrrhotite and finally chalcopyrite. Maximum temperature was not high enough for the pyrite to cross the brittle-ductile boundary. Instead, pyrite grains were incorporated and transported by pyrrhotite and chalcopyrite during the ductile remobilization events, rounding and fracturing them. Remobilization of the sulfides occurred mainly by plastic flow, but some solution transport and re-precipitation is locally observed.

In situ secondary ion mass spectrometry (SIMS) sulfur isotopes geochemistry of sulfides yielded values of $\delta^{34}S$ ranging from +2.7‰ to 4.7‰ for pyrite, +2.1‰ to 4.0‰ for pyrrhotite and +1.3‰ to 4.7‰ for chalcopyrite. Sulfur isotopic modeling suggests

that at least 60% of the sulfur was derived from leaching of igneous rocks (i.e., basalts), with the reminder derived from thermochemical sulfate reduction (TSR) of seawater sulfate during alteration of the basalts by seawater. At the deposit scale, sulfur isotopes retained their original signature and did not re-equilibrate during the secondary deformation and remobilization events.

Introduction

The Cu-rich Whalesback volcanic massive sulfide (VMS) deposit was discovered by the Betts Cove mining company in 1879 (Martin, 1983; Kean et al., 1995). In 1880, the property was sold to the Newfoundland Consolidated Copper Mining Company who dug numerous trenches and emplaced a 18m deep shaft in the hanging wall of the deposit, but no mineralization was encountered and exploration of the property stopped until mining rights were granted to the British Newfoundland Exploration Company (BRINEX) in 1957 (MacLean, 1947; Kean et al., 1995). Between 1960-1962, extensive exploration programs were jointly undertaken by BRINEX and the Anglo-American Corporation, successfully delineating a 2.7 million tonnes resource at 1.8% Cu. The Whalesback deposit entered production in 1963 and produce 3.8 million tonnes at 1% Cu over its 9 year mine life. In July 1972, the production ceased abruptly due a major cave-in breaching the surface and low Cu prices (Kean et al., 1995). More recently, an increase in the price of copper triggered a third exploration cycle at the Whalesback and adjacent Little Deer properties by Thundermin Resources Inc. To date, the Whalesback deposit has an indicated resource of 797,000 tonnes at 1.67% Cu (CIM/NI43-101; Brown, 2012)

The Whalesback deposit is a type example of an Appalachian stringer-dominated

Cyprus-type VMS deposit. Globally, Cyprus-type (also known as mafic-type) VMS deposits are commonly Cu-rich stratabound to stratiform, syngenetic deposits that form on or near the seafloor by precipitation from hydrothermal fluids at temperatures between 200-350°C (e.g., Large, 1977; Franklin et al., 1981, 2005; Lydon, 1984, 1988; Hannington, 2014). The ore mineralogy of Cyprus-type commonly consists of pyrite, chalcopyrite, pyrrhotite and sphalerite with minor amounts of galena, tetrahedrite, tennantite, arsenopyrite, bornite, and magnetite (e.g., Large, 1977, 1992; Eldridge et al., 1983; Lydon, 1988; Ohmoto, 1996; Franklin et al., 2005). They often have metal zoning patterns that were driven by temperature-dependent metal solubility differences with low temperature Zn-(Pb) deposition followed by higher temperature Cu deposition; the latter often leading to zone refining of earlier formed Zn-(Pb) sulfides (Ohmoto, 1996). The Cu-rich sulfides in Cyprus-type deposits, and like those at Whalesback, generally precipitate from hotter fluids often adjacent or within the footwall feeder conduit or at the base of the sulfide mound), whereas Zn- and Pb-rich sulfides precipitate from cooler hydrothermal fluids at the top and outer margins of the deposits.

Cyprus-type deposits, like all VMS deposits generally form within extensional geodynamic regimes, with Cyprus-type systems generally forming at mid-ocean ridges, back-arc basins, and intra-oceanic arc rifts (e.g., Swinden, 1991a; Piercey, 2010, 2011; Hannington, 2014). In ancient environments, the extensional stage of tectonic activity is commonly followed by uplift, basin inversion, compressional deformation, and metamorphism of the sequence hosting the massive sulfide deposits often due to post-VMS formation accretionary tectonics (e.g., McClay, 1995; Nelson, 1997). During this accretionary activity, rheological differences between sulfides and more competent

silicate minerals in the host sequence can lead to significant remobilization of the sulfides during deformation, creating distinct deformation and metamorphic textures such as *durchbewegung* (Cox, 1987; Marshall and Gilligan, 1987, 1989, 1993). *Durchbewegung* texture, as defined by Marshall and Gilligan (1989), consists of a mixture of secondary tectonic origin composed of angular to rounded clasts of competent materials (e.g., silicates) within a matrix of predominantly less competent material (e.g., sulfides) where the competent clasts are generally contorted and disoriented.

Despite a sound theoretical understanding of sulfide deformation processes, only a few studies have documented the effects of metamorphism and deformation on sulfide textures, and the primary sulfur isotope distribution in VMS deposits (e.g. Bachinski, 1977b; Craig and Vokes, 1993; Cook and Hoefs, 1997; Wagner et al., 2004; Barrie et al., 2010a; Barrie et al., 2010b). The purpose of this investigation is to document the mineralogy, textural relationships and sulfur isotope composition of the major sulfide phases present at the Cyprus-type Whalesback VMS deposit. This deposit is an outstanding natural laboratory to test the effects of deformation on sulfide textures and their potential influence (or lack thereof) on sulfur isotope systematics at low temperature and medium pressure.

Geology

Regional Geology

The Whalesback VMS deposit is located 11km northeast of the community of Springdale and is situated within the Notre Dame subzone of the Dunnage Zone of the Newfoundland Appalachians, Canada (Fig. 1). The Dunnage Zone is bounded to the west

by the Humber Zone and to the east by the Gander Zone (Williams, 1979; Williams et al., 1988; Hibbard et al., 2004). It represents the deformed vestiges of arcs, back arcs, and ophiolite complexes assembled during the closure of the Cambrian to Ordovician Iapetus Ocean (Fig. 1; Swinden et al., 1989; Swinden, 1991b; Kean et al., 1995; van Staal and Colman-Sadd, 1997; Evans and Kean, 2002; Rogers and van Staal, 2002; Rogers et al., 2006; van Staal, 2007; Zagorevski et al., 2010). The Dunnage Zone is divided into the western Notre Dame subzone, which formed near the Laurentian equatorial margin, and the eastern Exploits subzone, which formed on the edge of Gondwana and related microcontinents at mid- to high-southerly latitudes (e.g. Zagorevski et al., 2006; van Staal et al., 2007; Cocks and Torsvik, 2002). The Notre Dame subzone is subdivided into five zones which are, from oldest to youngest: 1) the Late Neoproterozoic–Cambrian ribbon-shaped Dashwoods microcontinent; 2) the 510-501 Ma mafic to ultramafic ophiolitic rocks of the Lushs Bight oceanic tract (LBOT); 3) the 489-477 Ma mafic to ultramafic ophiolitic rocks of the Baie Verte oceanic tract (BVOT); 4) the 488-453 Ma granodioritic to gabbroic Notre Dame magmatic arc; and 5) the 481-460 Ma ophiolite-arc-back arc tectonic collage of the Annieopsquotch accretionary tract (AAT) (Dunning and Krogh, 1985; Elliot et al., 1991; Szybinski, 1995; Cawood et al., 1996; Swinden et al., 1997; Waldron and van Staal, 2001; Zagorevski et al., 2006; van Staal, 2007; van Staal et al., 2007; Skulski et al., 2010; van Staal and Barr, 2012).

The Whalesback VMS deposit occurs within the LBOT sequence. Obduction of the LBOT sequence onto the western margin of the Dashwoods microcontinent occurred during phase 1 of the Taconic orogeny, which was initiated between 500 and 493 Ma (Szybinski, 1995; Swinden et al., 1997; van Staal and Barr, 2012). The composite LBOT-

Dashwoods terrane was accreted onto the Humber margin at ca. 480 Ma during phase 2 of the Taconic orogeny, resulting in the closure of the Taconic seaway and producing high-grade metamorphism and polyphase deformation in large parts of the Notre Dame subzone (van Staal, 2007; van Staal et al., 2007; van Staal and Barr, 2012). Collision of the Notre Dame subzone with the Exploits subzone occurred during phase 3 of the Taconic orogeny (455–450 Ma), initiating the collision of composite Laurentia with Ganderia, in which the peak of deformation occurred during the Salinic orogeny (445–423 Ma; Dunning et al., 1991; van Staal et al., 2003; van Staal, 2007; van Staal and Barr, 2012; Zagorevski et al., 2010). Subsequent collision with the Avalon terrane during the Acadian orogeny (421–390 Ma), with the Meguma terrane during the Neoacadian orogeny (ca. 395–340 Ma), and with Gondwana during the Alleghanian orogeny (ca. 340–260 Ma) lead to the formation of the Pangea supercontinent, ending ca. 250 m.y. of convergent tectonism (Hicks et al., 1999; Hatcher, 2002; Keppie et al., 2002; Reynolds et al., 2004; van Staal, 2005; van Staal et al., 2009; van Staal and Barr, 2012).

Geology of the Springdale Peninsula

Following the stratigraphy established by Szybinski (1995), the LBOT is subdivided into three groups that are, from oldest to youngest: 1) the mafic Lushs Bight Group, 2) the dominantly mafic volcanic with minor felsic volcanic Western Arm Group, and 3) the bimodal volcanic Cutwell Group (Fig. 2; Marten, 1971a, 1971b; Kean, 1973; Kean and Strong, 1975; Kean et al., 1995; Szybinski, 1995). The Whalesback deposit occurs in the Lushs Bight Group (Figs. 2, 3). The Lushs Bight Group is 3–4 km thick and is predominantly composed of lower greenschist facies metamorphosed tholeiitic basalt

with local boninite and minor dioritic to gabbroic sheeted intrusions. Rare felsic pyroclastic rocks, jasper and magnetite-bearing chert are also present. The basalts occur as pillowed or massive flows, and can locally be variolitic. In most cases, the variolitic basalts are of boninitic affinity and have quartz-filled amygdules. Areas of VMS mineralization tend to be spatially associated with areas of abundant boninites and felsic pyroclastic rocks (Kean et al., 1995; Szybinski, 1995). The Lushs Bight Group is crosscut by several generations of intrusive rocks, which include gabbro, quartz dacite, plagioclase porphyry, hornblende porphyry, hornblende-plagioclase porphyry, and pyroxene porphyry (Kanehira and Bachinski, 1968; Szybinski, 1995). The Lushs Bight Group is characterized by strong epidote alteration of the basalts, with a general decrease in epidote abundance towards the stratigraphic top of the group, and by locally extensive quartz $\pm$ carbonate veins (Kean et al., 1995).

The Lushs Bight Group was informally subdivided by Papezik and Fleming (1967) and Fleming (1970), on the basis of alteration, into a spilitic and chlorite altered “St. Patrick-type” and an epidote altered “Whalesback-type”. More recently, Szybinski (1995) subdivided the Lushs Bight Group into the basal Indian Head complex dominated by sheeted dikes and the overlying Little Bay Formation dominated by basalt.

Structural geology

Szybinski (1995) recorded five stages of deformation in the Notre Dame Bay region. The first phase of deformation (D_1) is a non-penetrative deformation linked with the mylonitization of pre-existing chlorite-rich syn-volcanic faults that were the feeder conduits to VMS mineralization, creating an S_1 foliation. During D_1 , these chlorite-rich

shear zones accommodated a component of dextral trans-current shearing (Szybinski, 1995) related to dextral oblique convergence between the LBOT and Dashwoods micro-continent (Dewey, 2002), and was synchronous with phase 1 of the Taconic orogeny (van Staal, 2007; van Staal and Barr, 2012). The second phase of deformation (D_2) produced a regional penetrative S_2 foliation and tight to isoclinal upright folds with NE-NNE axial planes throughout the Lushs Bight Group. Szybinski (1995) concluded that D_2 was synchronous with the emplacement of the Colchester and Coopers Cove plutons (465 ± 2 Ma), and coincident with phase 2 of the Taconic orogeny. At Whalesback, D_2 produced a steep S_2 schistosity trending $\sim 060^\circ N$ and dipping toward the southeast (Papezik, 1965; West, 1972). The third phase of deformation (D_3) is also a significant regional event that produced many large NE-trending folds and numerous brittle-ductile NE-striking thrust faults related to the emplacement of an alpine-style nappe in the Notre Dame Bay region (Szybinski, 1995). At Whalesback, D_3 resulted in the creation of an open fold with an axial trace trending $\sim 025^\circ N$ and dipping steeply to the north, deforming the chlorite schists into a major dextral drag fold (West, 1972). The fourth phase of deformation (D_4) is locally evidenced in the Lushs Bight Group as ENE to SE plunging antiforms, folds and thrusts, and is not present near the Whalesback deposit. D_4 reaches its maximum intensity in the vicinity of the Lobster Cove Fault, located approximately 10 km southeast of the study area. The fifth phase of deformation (D_5) represents the third major regional event in the Lushs Bight Group and resulted in the creation of NE-SW trending folds verging to the NW, NNW trending high-angle thrust faults and foreland propagating duplexes of various sizes (Szybinski, 1995). At Whalesback, mafic dykes emplaced during D_1 deformation are affected by NW directed high angle reverse faults (Szybinski,

1995).

Geology of the Whalesback Deposit

The Whalesback deposit is a Cu-rich (Cyprus-type) VMS deposit hosted in mafic volcanic rocks (Kanehira and Bachinski, 1968). It consists of veins, pods, and disseminated sulfides forming 0.3-15m wide mineralized lenses hosted within a 720m long and 120m wide highly chloritized shear zone that strikes 245°N and dips steeply towards the north (Figs. 3, 4). The Whalesback massive sulfide lenses are located in the central and hanging-wall portions of a chlorite shear zone that plunges southwest at about 50° (Fig. 4; Kanehira and Bachinski, 1968). These lenses occur at the site of maximum deformation intensity within the chlorite shear zone(s), suggesting extensive remobilization during post VMS-formation deformation events. With increasing proximity to the main shear zone, pillow basalts from the Lushs Bight Group become increasingly sheared and elongated parallel to the shear planes, primary pyroxenes are replaced by secondary chlorite, epidote alteration decreases in intensity, and quartz aggregates replace albite laths (Kanehira and Bachinski, 1968). Barren, weakly altered, regionally metamorphosed, and deformed gabbro, quartz dacite, plagioclase porphyry, hornblende porphyry, hornblende-plagioclase porphyry, and pyroxene porphyry intrude the chlorite altered shear zone and the ore lenses (Kanehira and Bachinski, 1968).

The sulfide lenses consist of pyrite, chalcopyrite, pyrrhotite and sphalerite with minor mackinawite, pentlandite, magnetite, cubanite, galena, and ilmenite (Kanehira and Bachinski, 1968). The sulfides assemblages at the Whalesback deposit are spatially zoned with pyrrhotite and chalcopyrite being the dominant sulfides in the Eastern lenses,

225 whereas pyrite, sphalerite and chalcopyrite are the most abundant sulfides in the Western
226 lenses (Fig. 4; West, 1972). Silicate alteration minerals associated with mineralization are
227 predominantly chlorite and quartz with minor muscovite, carbonate, titanite, albite, and
228 epidote (Kanehira and Bachinski, 1968). Alteration of the rocks surrounding the
229 Whalesback deposit within the chlorite-rich shear zone is marked by depletion of Na and
230 Ca and enrichment of Fe, S and K (Bachinski, 1977a), and is typical of VMS alteration
231 systems (e.g. Franklin et al., 2005; Hannington, 2014 and references therein).

232 **Mineralization**

233 Detailed drill core logging and conventional optical petrography of mineralized
234 intervals of 13 diamond drill cores acquired from the Western lenses and lower portion of
235 the Eastern lenses, recently drilled by Thundermin Resources Inc., (Fig. 4) allowed the
236 subdivision of the sulfide mineralization into four types: 1) disseminated; 2) veinlet; 3)
237 breccia; and 4) semi-massive to massive. The sulfide minerals consist of chalcopyrite,
238 pyrrhotite, pyrite and sphalerite, and coexist with silicate gangue minerals, which include
239 chlorite (40-100%) with subordinate quartz (<50%), calcite (<40%), muscovite (<30%),
240 epidote (<25%), feldspar (<20%) and titanite (<5%). Trace amounts (<0.01%) of 1-10µm
241 long Ag, Bi and Hg tellurides were identified by scanning electron microscopy within the
242 sulfide matrix.

244 *Textures and Distribution of Major Sulfide Phases*

245 *Western lenses.* Sulfides in the Western lenses consist of pyrite (~85%) with
246 subordinate sphalerite (~10%) ± chalcopyrite (~5%) hosted in highly chloritized basalts

(Fig. 5ab). Pyrite occurs as euhedral to subrounded grains and is spatially associated with quartz and sphalerite. Sphalerite occurs as small (<1 mm long) euhedral crystal aggregates that are interstitial to and infill fractures of pyrite grains, and commonly contains micro-inclusions of chalcopryrite, i.e., chalcopryrite disease (Barton and Bethke, 1987). Greater abundances of sphalerite (up to 20%) are present in the semi-massive and massive pyrite intervals. Chalcopryrite occurs as anhedral grains with no visible individual grain boundaries, forming a matrix that crosscuts and fills fractures in pyrite and sphalerite. The chalcopryrite matrix locally contains 0.1-3 mm long subrounded to rounded pyrite grains.

Eastern lenses. Sulfides in the deeper portion of the Eastern lenses consist of chalcopryrite (~45%) with lesser pyrrhotite (~40%), pyrite (~15%), and trace sphalerite (<1%). Chalcopryrite and pyrrhotite have no observable internal grain boundaries and form a sulfide matrix that varies between pyrrhotite-rich and chalcopryrite-rich (Fig. 5cdef). Chalcopryrite crosscuts and separates pyrrhotite-rich intervals into smaller aggregates, isolating them in its matrix (Figs. 5f, 6d, and 7ad). The chalcopryrite-pyrrhotite matrix contains angular to rounded, 0.1-3 mm long fractured pyrite grains and small aggregates of anhedral sphalerite that contain micro-inclusions of chalcopryrite, i.e., chalcopryrite disease (Fig. 8a; Barton and Bethke, 1987). When sulfides are massive to semi-massive the chalcopryrite-pyrrhotite matrix fills fractures in pyrite and sphalerite (Figs. 7b, 8bc) and is locally intergrown with fibrous chlorite (Fig. 7b).

Types of mineralization

Disseminated-type. This mineralization type is dominated by pyrrhotite and pyrite with subordinate chalcopyrite and sphalerite, hosted by chlorite-rich basalt (Fig. 6a-c). Disseminated sulfides occur as discrete grains in proximity to vein-type mineralization (see below) and manifests itself as 1-3 mm long anhedral and asymmetrical grains that are weakly to strongly elongated parallel to sulfide veins. Disseminated sulfides are associated with >95% silicate matrix and account for ~5% of the total sulfide resource at the Whalesback deposit.

Veinlet-type. This mineralization type consists of chalcopyrite, pyrrhotite and pyrite veinlets that range from 1 to 10 mm in width. The veinlets have an anastomosing texture and commonly surround the angular to rounded fragments of the silicate gangue, usually dominated by quartz, and constitute 10-40% of the host rock (Fig. 6d-f). Pyrrhotite veinlets contain accessory rounded to subrounded pyrite and sphalerite, and are aligned parallel to S_2 . In contrast, chalcopyrite veinlets are generally monomineralic and crosscut the pyrrhotite veinlets at a high angle. This type of mineralization accounts for ~50% of the total sulfide resource at the Whalesback deposit.

Breccia-type. This mineralization type consists of chalcopyrite, with minor pyrrhotite and pyrite. It contains 50-95% angular to subrounded dark chloritized basalt fragments within a 5-50% sulfide matrix (Fig. 6g-i). The breccia-type mineralization accounts for ~25% of the total sulfide resource at the Whalesback deposit.

Semi-massive to massive sulfide. This mineralization type consists of a mixture of chalcopyrite and subordinate pyrite and pyrrhotite in the Eastern lenses and of a mixture of pyrite and subordinate sphalerite and chalcopyrite in the Western lenses (Fig. 7a-f). Small, 1-5 mm long, subrounded to flattened (parallel to S_2) fragments of silicate

minerals, mainly chlorite-altered wall rock, and constitute between 5 and 50% of the rock (Fig. 7ad). In areas with more intense deformation, the flattened fragments are heterogeneously distributed throughout the sulfide matrix, giving this type of mineralization a *durchbewegung* texture. Small 0.2-3 mm wide chalcopyrite veinlets commonly crosscut the semi-massive to massive sulfide assemblage (Fig. 7d). This type of mineralization accounts for ~20% of the total sulfide resource at the Whalesback deposit.

Sulfur Isotopes

One representative sample of massive sulfide from the Western lenses and five representative samples of massive sulfide (n=2), semi-massive sulfide (n=1), breccia (n=1) and veinlets (n=1) from the Eastern lenses (n=5) were chosen for *in situ* sulfur isotope analysis by secondary ion mass spectrometry (SIMS). Sample mounts were prepared from thin section off cuts by embedding ~1 cm x 1 cm pieces in epoxy in 1 inch diameter aluminum retaining rings. Samples were polished using standard lapidary procedures, and then sputter coated with 300 Å of Au to reduce charging under primary ion bombardment. All analyses were performed using a Cameca IMS 4f SIMS instrument at the MAF-IIC Microanalysis Facility at Memorial University.

The $\delta^{34}\text{S}$ data was collected from pyrite, pyrrhotite, and chalcopyrite using the method described in Toman (2013) and Brueckner et al. (2014). In summary, determinations were performed by bombarding the sample with a primary ion microbeam of 600-850 picoamps (pA) of Cs^+ , accelerated through a nominal 10 kiloelectron volt (keV) potential, and focused into a 5–15 μm diameter spot on the sample. To exclude

exotic material in the polished surface from analysis, each spot was first pre-sputtered for 100 s with a 25 μm square raster applied to the beam. Negatively charged sputtered secondary ions were accelerated into the mass spectrometer of the instrument through a nominal potential of 4.5 keV. A combination of mass resolution (Mass Resolving Power approx. 2975) and energy filtering (sample offset -60 eV with a 40 eV energy window) were used to effectively eliminate isobaric interferences from $^{16}\text{O}_2^-$ and $^{33}\text{SH}^-$. Signals for $^{32}\text{S}^-$, $^{34}\text{S}^-$ and a background position at 31.67 Dalton (Da) were obtained by cyclical magnetic peak switching (typically 80 cycles were accumulated for each spot analysis), and detected using an electron multiplier in pulse counting mode. Standard counting times were: 0.5 s at the background position, 2.0 s on $^{32}\text{S}^-$, and 6.0 s on $^{34}\text{S}^-$. Waiting times of 0.25 s were inserted before each peak counting position to allow for magnet settling. All analyses are reported using the standard δ notation in units of per mil (‰) relative to the international Vienna Canyon Diablo Troilite (VCDT) standard.

Internal precision on individual $\delta^{34}\text{S}$ analysis is typically better than $\pm 0.3\text{‰}$ (1σ), whereas the overall reproducibility is typically better than $\pm 0.5\text{‰}$ (1σ). In house standards for pyrite (UL9B), pyrrhotite (PoW1) and chalcopyrite (Norilsk) were used to correct for instrumental mass fractionation for each specific mineral phase.

Results

Sulfide minerals from the Whalesback deposit display a relatively narrow range of sulfur isotope compositions. Measured $\delta^{34}\text{S}$ of 18 pyrite samples range from +2.7 to +6.6 per mil with an average of +5.0 per mil; 7 pyrrhotite samples range from +2.1 per mil to +4.0 per mil with an average of +3.3 per mil; and 17 chalcopyrite samples range

from +1.3 to +4.7 per mil with an average of +2.9 per mil (Figs. 9a-f, 10, 11; Table 1). Within individual samples, the internal variations of $\delta^{34}\text{S}$ in pyrrhotite and chalcopyrite are generally low (<0.6 per mil), with a maximum observed variation of 1.0 per mil (Fig. 9a-f). Pyrite shows the greatest internal variations, and most of the $\delta^{34}\text{S}$ range observed for Whalesback as a whole is present in one sample from the Eastern lenses (e.g., WB-12-106A_561.05m, +4.5 to +6.4 per mil; Fig. 9c; Table 1). However, even for pyrite, the internal variation within individual grain is quite restricted in overall magnitude (3.5 to 5.2 per mil; Fig 9c).

Discussion

Morphology and Metal Zoning of the Whalesback Deposit

The Whalesback deposit consists of a tightly folded and boudinaged deposit dominated by chalcopyrite and pyrrhotite in the Eastern lenses and by pyrite in the Western lenses (Fig. 4; Papezik, 1965; West, 1972; Szybinski, 1995). In most genetic and descriptive models for Cyprus-type VMS deposits, there is a well developed zonation of metals with chalcopyrite and pyrrhotite concentrated at the base and in the core of the deposits, and pyrite and sphalerite concentrated at the top and more distal parts of the deposit (Fig. 12a; e.g. Large, 1977; Franklin, 1981; Lydon, 1984, 1988; Ohmoto, 1996; Franklin et al., 2005; Hannington, 2014). The metal distribution at Whalesback suggests that the pyrite-rich Western lenses represent the top or distal parts of the Whalesback deposit, whereas the chalcopyrite-pyrrhotite-rich Eastern lenses represent the core and base of the deposit, in accordance with a younging direction towards west. Metal zoning projected above the present day surface implies that D₂ deformation produced a closed

359 overturned fold (Fig. 12c). Overturned folds at this location have not been previously
360 documented and have implications for the longer-term exploration of the deposit.
361 Furthermore, it confirms the Whalesback deposit has been highly deformed, folded and
362 boudinaged during D₂ resulting in a closed overturned anticline (Fig. 12bc), and creating
363 the characteristic durchbewegung deformation texture observed at hand specimen scale.
364

365 *Sulfide textural evolution during deformation and remobilization*

366 Durchbewegung textured rocks are generally hosted within durchbewegung
367 structures (Marshall and Gilligan, 1989), which form from the progression of folding and
368 shearing related to the formation of tight to isoclinal folds. In these structures, the silicate
369 layers surrounding the less competent sulfide layers become disrupted, detached and
370 incorporated into the sulfides as they deform ductily. This process results in the creation
371 of detached fold cores and boudins of varying geometry depending on competency
372 contrast between the silicate and sulfide layers (Marshall and Gilligan, 1989). Relatively
373 late piercement (or ductile) veins are commonly associated with the formation of
374 durchbewegung structures. The veins range in size from a few centimeters to a few
375 meters wide and generally form at high angle to the layer interface and fill in extensional
376 planar fractures (Marshall and Gilligan, 1989).

377 At Whalesback, the majority of deformation and remobilization occurred during
378 the D₂ deformation event, as most of the sulfide-rich zones are parallel to the main S₂
379 schistosity (Papezik, 1965; West, 1972; Szybinski, 1995). Remobilization and
380 deformation of sphalerite, pyrrhotite, and chalcopyrite was due to ductile flow, whereas
381 pyrite deformed brittly as demonstrated by its rounded habit and high degree of fracturing

(Figs. 7ab, 8a-d). Sphalerite, pyrrhotite, and chalcopyrite fill the fractures in pyrite, supporting their ductile remobilization. Moreover, pyrrhotite and chalcopyrite remobilization incorporated and transported pyrite grains, rounding and fracturing them (Figs. 8cd, 11def). Based on crosscutting relationships, sphalerite was the first sulfide phase to be remobilized followed by pyrrhotite and chalcopyrite, and the latter were followed by the formation of chalcopyrite veins (Fig. 8a-d). The sequence in which the sulfides crossed the brittle-ductile transition zone is indicative of low to medium pressure (>175 MPa) and low temperature deformation conditions, which were insufficient conditions for pyrite to deform ductily (Fig. 13; Marshall and Gilligan, 1987). These conditions are consistent with the lower greenschist grade metamorphism recorded elsewhere in the Lushs Bight Group (Kean et al., 1995).

The main transport mechanism of ductile sulfides at Whalesback, during peak metamorphism conditions, is interpreted to have been plastic flow, with minor solution transport and precipitation. Cox (1987) showed that the intergrowth of sulfides with fibrous phyllosilicates in low-grade metamorphic environments, like those observed at Whalesback, indicates that some material remobilization occurred by dissolution, solution transport and precipitation. However, since the intergrowth of fibrous chlorite with massive pyrrhotite and chalcopyrite is only observed locally (Figs. 7hi, 8b), it is assumed that plastic flow was likely the dominant transport mechanism.

The occurrence, and likely late remobilization, of chalcopyrite in high angle veins relative to S_2 across the sulfide assemblages, is very common and expected in durchbewegung structures (Marshall and Gilligan, 1989), and suggests that the late chalcopyrite veins are a direct consequence of the structural evolution of the

durchbewegung structure hosting the Whalesback deposit. Close folding and boudinage of Cu-rich ore (Fig. 4) also support a structural evolution within a durchbewegung structure.

Influence of Deformation on $\delta^{34}\text{S}$ Distribution

Only a few studies have investigated the effect of metamorphism on the primary sulfur isotopic composition of sulfide minerals in VMS deposits (e.g. Bachinski, 1977b; Craig and Vokes, 1993; Crowe, 1994; Cook and Hoefs, 1997; Wagner et al., 2004). Most of these studies were undertaken in areas affected by classic Barrovian-style metamorphism, i.e., where the temperature and the pressure gradually increase together, and all concluded, except Crowe (1994), that the original sulfur isotopic composition of individual sulfide phases was preserved during metamorphism. Bachinski (1977b) investigated the sulfur isotopic composition of VMS deposits within the Lushs Bight Group, including the former Whalesback mine, and reached a similar conclusion.

The $\delta^{34}\text{S}$ values recorded from all sulfide phases during this study display a narrow range of +1.3 to +6.6 per mil and are heterogeneously distributed at the scale of the Whalesback deposit (Figs. 9a-f, 10, 11; Table 1). Pyrrhotite is the most consistent sulfide mineral with a range of only 1.9 per mil, followed by chalcopyrite with a range of 3.4 per mil and pyrite with a range of 3.9 per mil. Equilibrium temperature derived from coexisting sulfides phases (chalcopyrite-pyrite, chalcopyrite-pyrrhotite, and pyrrhotite-pyrite) using Kajiwara and Krouse (1971) fractionation factors yield temperatures between 27 and 526°C (Fig. 14a-c) with one additional sample yielding a temperature of 4605°C, above normal hydrothermal fluids and metamorphism temperature range. In

addition, one sample had a $\Delta^{34}\text{S}_{\text{Py-Ccp}}$ of -0.6 that could not be solved using the equilibrium temperature equations of Kajiwara and Krouse (1971), indicating isotopic disequilibrium. Overall, most derived equilibrium temperatures are below the 200°C isotherm, which is lower than the typical Cu-rich VMS deposits formation temperature (250-350°C; Large, 1977; Franklin et al., 1981, 2005; Lydon, 1984, 1988; Hannington, 2014). These low equilibrium temperatures can be due to a partial to complete re-equilibration of the sulfide phases at low temperature during the deformation and associated metamorphism (isotopic equilibrium), the product of primary precipitation of coexisting sulfide phases at different temperature yielding a low “apparent” equilibrium temperature (isotopic disequilibrium), or the result of isotopic heterogeneity of the mineralizing fluids through time induced by sulfur extraction from more than one source (isotopic disequilibrium).

Paragenetic relationships of sulfide phases at Whalesback suggest that the increase in pressure and temperature during D₂ produced a remobilization and deformation sequence wherein sphalerite crossed the brittle-ductile transition first, followed by pyrrhotite and by chalcopyrite. This remobilization may have lead to partial to complete re-equilibration of the sulfides. If complete isotopic equilibration was attained, each sulfide phases within the deformation zone should have a homogenous composition (i.e., all chalcopyrite should have similar $\delta^{34}\text{S}$ values) and the derived equilibrium temperatures should be constant throughout the deposit, which is not the case at the Whalesback deposit (Figs. 10, 14a-c). Partial equilibration at hand sample scale is also unlikely as samples containing more than one sulfide pairs yield a wide range of equilibrium temperatures (Fig. 14a-c). It is therefore suggested that the sulfides are not in

isotopic equilibrium and at least partly retained their original isotopic composition, as previously proposed by Bachinski (1977b). However, given that the isotopic compositions of the ductile phases (pyrrhotite and chalcopyrite) are more homogeneous than the pyrite, it is suggested that some local homogenization may have occurred.

Using the fractionation factors of Ohmoto and Rye (1979), the maximum expected variation in $\delta^{34}\text{S}$ between for a sulfide pair that precipitated from the same fluids, but at different temperatures (250-350°C), is 1.4 per mil for the pyrite-chalcopyrite pair, 0.3 per mil for the pyrrhotite-chalcopyrite pair, and 1.2 per mil the pyrite-pyrrhotite pair. These variations are not sufficient to explain the $\delta^{34}\text{S}$ variation at Whalesback. Therefore, we suggest that variation in primary precipitation temperature of the coexisting sulfide phases is not important at the Whalesback deposit and that the provenance of sulfur is the main controlling factor on the $\delta^{34}\text{S}$ distribution.

Origin of sulfur

The sulfur in VMS deposits may originate from multiple sources, including: 1) sulfur derived from thermochemical sulfate reduction (TSR) of seawater sulfate via the partial reduction of oxidized seawater to isotopically lighter H_2S during seawater alteration of basement rocks. This process commonly results in sulfur in VMS sulfides that is ~17 to 25 per mil lighter than coexisting seawater sulfate (e.g., Sangster, 1968; Ohmoto and Rye, 1979; Ohmoto and Goldhaber, 1997; Huston, 1999; Shanks, 2001; Seal, 2000, 2006); seawater sulfate itself has varied greatly through geological time (+4 to +34.5 per mil; $\delta^{34}\text{S}_{\text{Cambrian}} = +34.5$ per mil; Claypool et al., 1980; Canfield et al., 2004; Kampschulte and Strauss, 2004; Paytan and Grey, 2012); 2) sulfur derived from anoxic

seawater with a sulfide isotopic composition 15 to 60 per mil lighter than sulfate derived from oxidized seawater (-25.5 to +19.5 per mil for Cambrian anoxic seawater; Goldhaber and Kaplan, 1975), and wherein the magnitude of the fractionation is related to sedimentation rates. Low sedimentation rates are associated with slower rate of sulfate reduction and larger fractionation, and high sedimentation rates are associated with faster rates of sulfate reduction and small fractionation. Sediments in anoxic environment are generally associated with elevated concentration of organic matter and sulfide minerals in which pyrite commonly display a framboidal texture (Goldhaber and Kaplan, 1975); 3) sulfide sulfur sourced by direct leaching from the basaltic rocks of the Lushs Bight Group underlying the mineralized stratigraphic horizons. In the case of submarine mafic volcanic hosted Cu-rich deposits, like Whalesback, the $\delta^{34}\text{S}_{\text{Sulfides}}$ of basalt ranges from +0.3 to +0.7 per mil (Torssander, 1992); 4) magmatic sulfur with $\delta^{34}\text{S}_{\text{Reservoir}} \approx +0 \pm 2$ per mil (Hoef, 2009). At temperatures above 300°C the dominant sulfur species is SO_2 , but at lower temperatures the magmatic SO_2 disproportionates rapidly to form H_2S and H_2SO_4 . This latter reaction is accompanied by a large isotopic fractionation that produces negative $\delta^{34}\text{S}_{\text{H}_2\text{S}}$ values often resulting in sulfides that have negative $\delta^{34}\text{S}$ (Rye, 1993; Huston et al., 2011); and 5) bacteriogenic reduction of sulfate in marine or pore waters. Precipitation of sulfide by bacteria is associated with a strong fractionation, ranging from 24 to 71 per mil and averaging 51 ± 10 per mil, depending on the sulfate availability of the system (Ohmoto and Rye, 1979; Canfield and Teske, 1996; Ohmoto and Goldhaber, 1997; Canfield, 2001; Seal, 2006). In the case of the Cambrian Whalesback deposit, $\delta^{34}\text{S}$ values ranging between -36.5 and +10.5 per mil with an average of -16.5 ± 10 per mil are expected for sulfide minerals derived from biogenic reduction of seawater sulfate during

the Cambrian (Kamprschulte and Strauss, 2004).

The absence of negative $\delta^{34}\text{S}$ values, pyrite framboids, and the narrow positive range of $\delta^{34}\text{S}$ obtained from the Whalesback deposit during this study (Table 1) suggest that biogenic sulfur input was insignificant. Furthermore, a significant component of magmatic sulfur is also highly unlikely, given these restricted positive $\delta^{34}\text{S}$ values (i.e., no evidence of sulfur disproportionation; Rye, 1993; Huston et al., 2011), and the absence of other features that are common to VMS deposits formed from magmatic fluid input, i.e. Au-Ag-enrichment, intermediate- to high-sulfidation ore mineral assemblages, and aluminous alteration assemblages (e.g., Sillitoe et al., 1996; Hannington et al., 1999; Dubé et al., 2007). Anoxic sulfur is also an unlikely sulfur source as carbonaceous and sulfur-rich sediments are rare in the vicinity of the Whaleback deposit. Moreover, the basalts of the Lushs Bight Group near the Whalesback deposit contain hematitized pillow lava with common to extensive magnetite-bearing red cherts (Kean et al., 1995), consistent with deposition in oxidized seawater.

To evaluate the relative roles of leaching from igneous host rocks and TSR, modeling was undertaken using the method described in Brueckner et al. (2014) using the equations outlined in Ohmoto and Rye (1979), Ohmoto and Goldhaber (1997), and Seal (2006). Thermochemical sulfate reduction (TSR) of seawater sulfate was modeled on the basis of three assumptions: 1) Cambrian seawater sulfate had a $\delta^{34}\text{S} \approx +34.5$ per mil (Claypool et al., 1980; Canfield et al., 2004; Kamprschulte and Strauss, 2004; Paytan and Grey, 2012); 2) TSR occurred at $\sim 300^\circ\text{C}$ (or higher), based on the observation of chalcopyrite-pyrrhotite assemblages in the sulfides; and 3) the system was open and was continuously replenished as the available heavy sulfate (SO_4) was reduced to lighter H_2S ,

producing H₂S with $\delta^{34}\text{S}_{\text{H}_2\text{S}(300^\circ\text{C})} \approx +12.7$ per mil during TSR (i.e., assuming that $f = 1$, where f is the fraction of SO₄ remaining in solution at site of reduction reaction). The fluid was then cooled from 300°C and various sulfide phases were precipitated, with most minerals precipitated at temperatures > 250-300°C given the chalcopyrite-rich assemblages at the Whalesback deposit. Figure 15 illustrates the sulfide isotopic composition as a function of cooling for the interpreted temperatures of deposition for sulfide solely derived from TSR. The model shows that TSR alone cannot generate the $\delta^{34}\text{S}$ range determined for the sulfide phases at the Whalesback deposit. Even at the modeled temperature of 250°C, TSR-derived sulfides would still have $\delta^{34}\text{S}$ substantially heavier than the range observed at Whalesback. This implies that sulfur from leaching of igneous rocks must be an important contributor to the overall sulfur budget of the deposit.

Binary mixing proportions of sulfur sourced from mafic host rock leached H₂S at 300°C ($\delta^{34}\text{S}_{\text{py}} \sim +1.2$ per mil, $\delta^{34}\text{S}_{\text{po}} = +0.3$ per mil, $\delta^{34}\text{S}_{\text{ccp}} = -0.2$ per mil) and sulfur derived from TSR at 300°C can be calculated using the following equation:

$$\delta^{34}\text{S}_{\text{sample}} = X\delta^{34}\text{S}_{\text{TSR}} + (1-X)\delta^{34}\text{S}_{\text{leached}} \quad (1),$$

where $\delta^{34}\text{S}$ is the measured sulfur isotopic signature of the sample, $\delta^{34}\text{S}_{\text{TSR}} = \delta^{34}\text{S}$ derived from TSR calculation at 300°C for a given mineral phase, $\delta^{34}\text{S}_{\text{leached}} = \delta^{34}\text{S}$ of igneous-derived sulfur in a given mineral phase at 300°C, and X is a proportionality factor that is 1 when 100% of the sulfur is from TSR and 0 when there is 0% TSR (i.e., 100% leaching of igneous sulfur). The results show that sulfur derived from TSR at 300°C accounts for a maximum of ~40% of the sulfur budget of the Whalesback deposit, whereas between 60-90% has been derived from leaching from the surrounding footwall rocks (Fig. 16a-c).

This result is expected given the high temperature nature of the mineralization at the

Whalesback deposit (i.e., Cu-rich assemblages) and the relatively restricted range of measured $\delta^{34}\text{S}$ values, further arguing that leaching from nearby footwall rocks was the most important source for sulfur in the Whalesback deposit. Since the Lushs Bight Group is 3-4 km in thickness (Kean et al., 1995), there is no restriction for the depth at which the sulfur and metals present at Whalesback were extracted. However, given that copper solubility is low below 225°C and greatly increase with increasing temperature (Hannington et al., 2014), it is suggested that the sulfur and the metals were leached from the lower portion of the mafic Lushs Bight Group. This single mafic source for the extraction of the metals is also reflected in the simple mineralogy (chalcopyrite, pyrite, pyrrhotite, and subordinate sphalerite) encountered at the Whalesback deposit.

Conclusions

The Whalesback deposit is a Cu-rich VMS deposit hosted within a closed, boudinaged and overturned fold located within a durchbewegung structure. Intense deformation of the Whalesback deposit occurred at low temperature and medium pressure (>175 MPa) during the accretion of the LBOT-Dashwoods composite terrane onto the Humber margin at ca. 480 Ma (D_2). The increase in pressure and temperature produced a remobilization sequence where sphalerite was the first sulfide phase to cross the brittle-ductile transition, followed by pyrrhotite and subsequently by chalcopyrite. The maximum metamorphic temperature was not high enough for pyrite to deform ductily. Instead, pyrite grains were incorporated and transported during the ductile remobilization of less competent sulfide phases (pyrrhotite and chalcopyrite). This led to the rounding and fracturing of the pyrite grains. Remobilization of sulfides occurred

565 primarily by plastic flow but some evidence of dissolution, solution transport and
566 precipitation was observed locally.

567 Sulfur isotopes retained their original signatures and did not re-equilibrate during
568 the remobilization events. The dominant source of sulfur was determined to be from
569 sulfides leached from surrounding basalts, with a lesser component derived from
570 thermochemical sulfate reduction of seawater sulfate during alteration of the basalts.

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Figure 1. Simplified geological map of the Newfoundland Appalachians with tectonostratigraphic zones (modified from van Staal, 2007, and van Staal and Barr, 2012).

Abbreviations: BBL = Baie Verte Brompton Line; DBL = Dog Bay Line; DF = Dover Fault; GBF = Green Bay Fault; LCF = Lobster Cove Fault; LRF = Lloyds River Fault; RIL = Red Indian Line.

; LR = Long Range; LRF = Lloyds River Fault; PP = Pipestone Pond Complex; RIL = Red Indian Line; SA = St. Anthony; TP = Tally Pond Belt; TU = Tulks Volcanic Belt; VA = Victoria Arc; WB = Wild Bight Group.

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(brittle) and aggregates of pyrrhotite (ductile) displaced by chalcopyrite. Sample WB-12-109_578.6m. Abbreviations are as follow: Ccp = chalcopyrite, Po = pyrrhotite, Py = pyrite, Sp = sphalerite, RL: reflected light and XP: cross-polar.

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Figure 10. Notched whisker plots for sulfur isotopic data for chalcopyrite, pyrrhotite and pyrite. The width of the notches is proportional to the half width (HW = [75th Percentile

– 25th Percentile] $\times 1.57 / [\sqrt{N}]$, where N is the population size). The central portion of the notched is the median and is represented by a line across the whisker box. The whiskers represent the extent of the dataset. Statistically, the chalcopyrite and pyrrhotite represents one population while the pyrite represents a different population.

Figure 11. Spatial distribution of $\delta^{34}\text{S}$ samples within the Whalesback deposit. Also shown is range in the $\delta^{34}\text{S}$ values obtained on pyrite and chalcopyrite mineral separates from the former Whalesback mine (Bachinski, 1977).

Figure 12. Schematic representation of the evolution of the Whalesback deposit (A-C) at the deposit scale and (D-F) at the micro-section scale. (A and D) Formation stage. The Whalesback deposit is deposited near and/or on the seafloor and displays a spatial zonation of metals with chalcopyrite- and pyrrhotite-rich sulfides at the base and in the core of the deposit and pyrite-rich sulfides at the top and more distal parts of the deposit. The chalcopyrite- and pyrrhotite-rich zone contains of euhedral pyrite while the pyrite-rich zone is composed of euhedral pyrite and sphalerite with minor pyrrhotite and chalcopyrite. (B and E) Early D_2 deformation stage. The deposit is compressed along the lateral axis, creating the main S_2 schistosity in the basalt host rock and ductily deforming and remobilizing sphalerite and pyrrhotite, and rounding the pyrite grains. Chalcopyrite-rich lenses were aligned parallel to main S_2 schistosity during that stage. (C and F). Present day disposition of the Whalesback deposit showing a closed and boudinaged overturned fold configuration. The late stage of deformation ductily remobilized chalcopyrite and incorporated fragments of ductile pyrrhotite and brittle pyrite,

continuing the mechanical rounding of the pyrite grains. Late chalcopyrite veins crosscut all the previously formed mineral assemblages.

Figure 13. Metamorphic pressure-temperature diagram with superposed brittle-ductile behavior boundaries. Mechanical behavior of common base-metal sulfides at the brittle-ductile transitions at 5% ductile strain before faulting, and at strain rates in the order of $7.2 \times 10^{-5} \text{ s}^{-1}$ (diagram modified from Marshall and Gilligan, 1987). The general ductile remobilization sequence at Whalesback is depicted by the arrow and starts in the sphalerite ductile domain, followed by sphalerite-pyrrhotite ductile domain and ends in the sphalerite-pyrrhotite-chalcopyrite ductile domain. This sequence is only possible for temperature below 150°C (grey highlighted area). Abbreviations are as follow: Ccp = chalcopyrite, Gn = galena, Po = pyrrhotite, Py = pyrite, Sp = sphalerite.

Figure 14. $\delta^{34}\text{S}$ vs $\delta^{34}\text{S}$ plots of coexisting sulfide phases: (A) pyrite-chalcopyrite, (B) pyrrhotite-chalcopyrite, and (C) pyrite-pyrrhotite. Isotherms were calculated for temperatures between 100 to 750°C using the fraction factors of Kajiwara and Krouse (1971). Also shown are the isotopic equilibrium temperatures (in italic).

Figure 15. Sulfur isotopic composition for sulfide derived solely from thermochemical sulfate reduction of seawater sulfate (TSR) at 250, 300 and 350°C. (A) Chalcopyrite, (B) pyrrhotite, and (C) pyrite. Also shown is the range in sulfur isotopic composition of sulfide phases observed during this study.

1014 Figure 16. Modeled sulfide isotopic composition of (A) pyrite, (B) pyrrhotite and (C)
1015 chalcopyrite deposited at 250, 300 and 350°C in relation to the proportion of TSR and
1016 igneous sulfide required to produce them. The range of TSR proportion required to
1017 produce the observed sulfide isotopic composition at Whalesback is highlighted in the
1018 shaded area. Also shown is the range in sulfur isotopic composition of sulfide phases
1019 observed during this study.
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1023 Table 1. SIMS $\delta^{34}\text{S}$ stable isotopic data for pyrite, pyrrhotite and chalcopyrite phases
1024 from the Whalesback deposit. (See sulfur isotopes section for analytical details).

1025

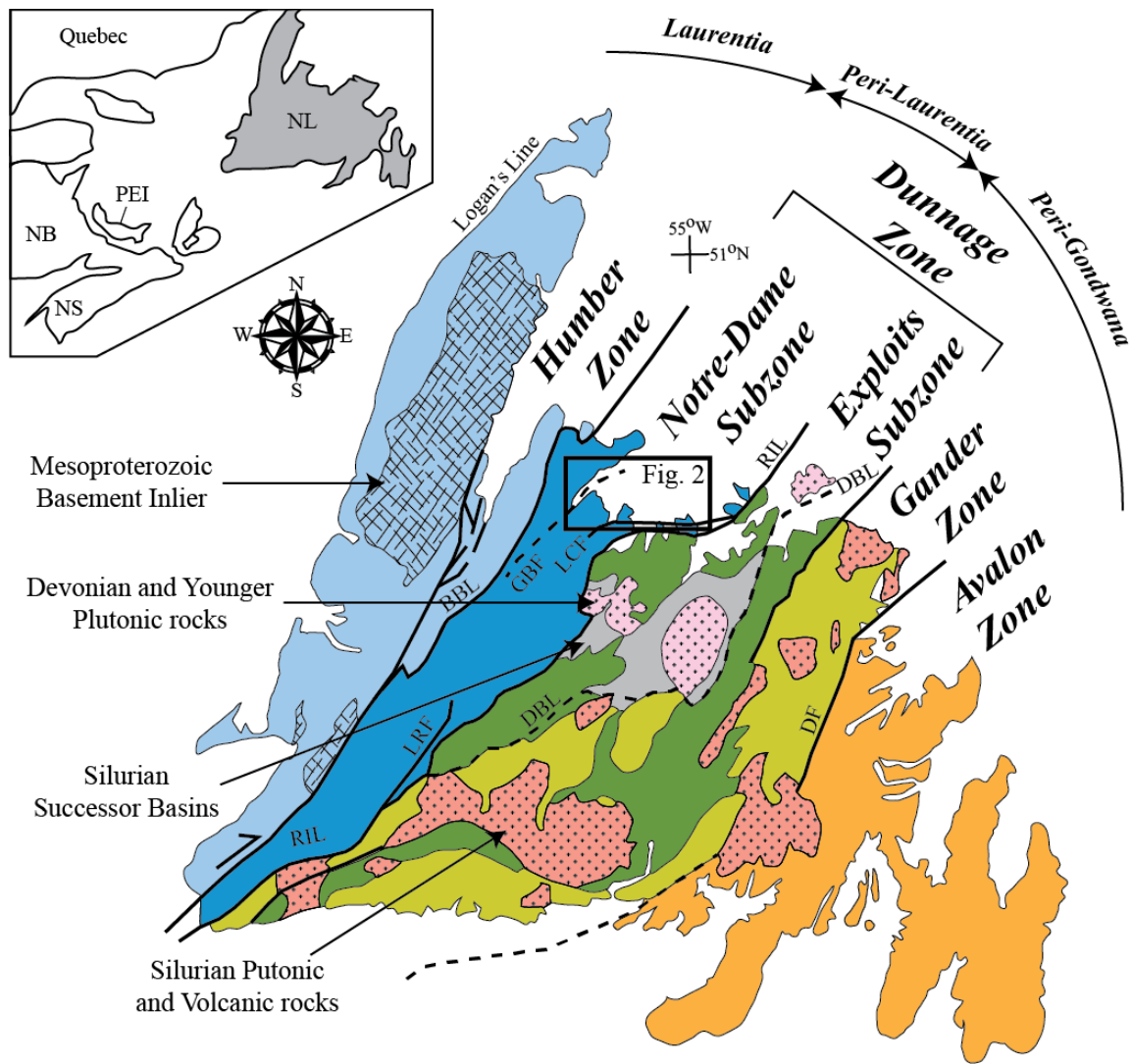


Figure 1

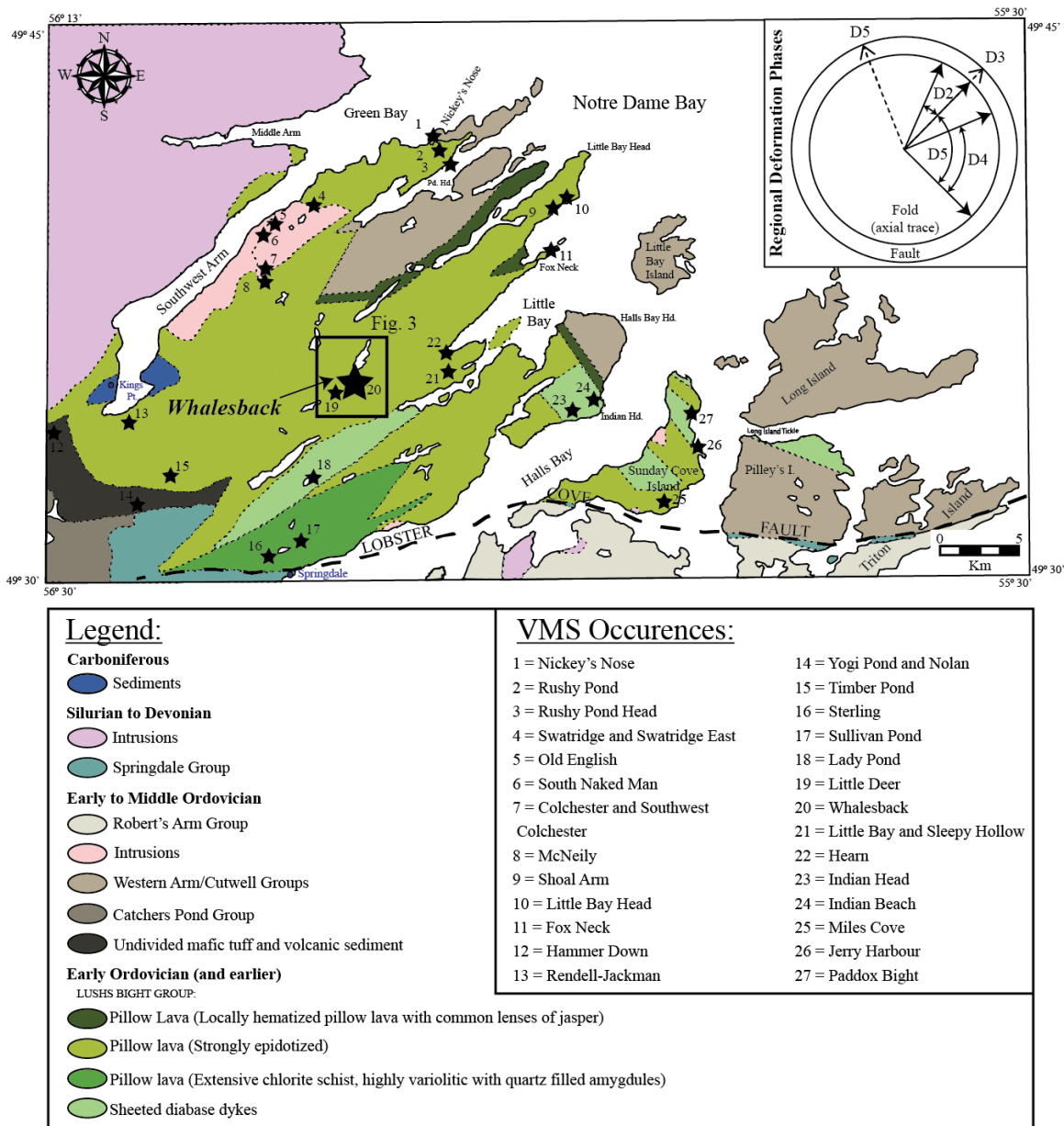


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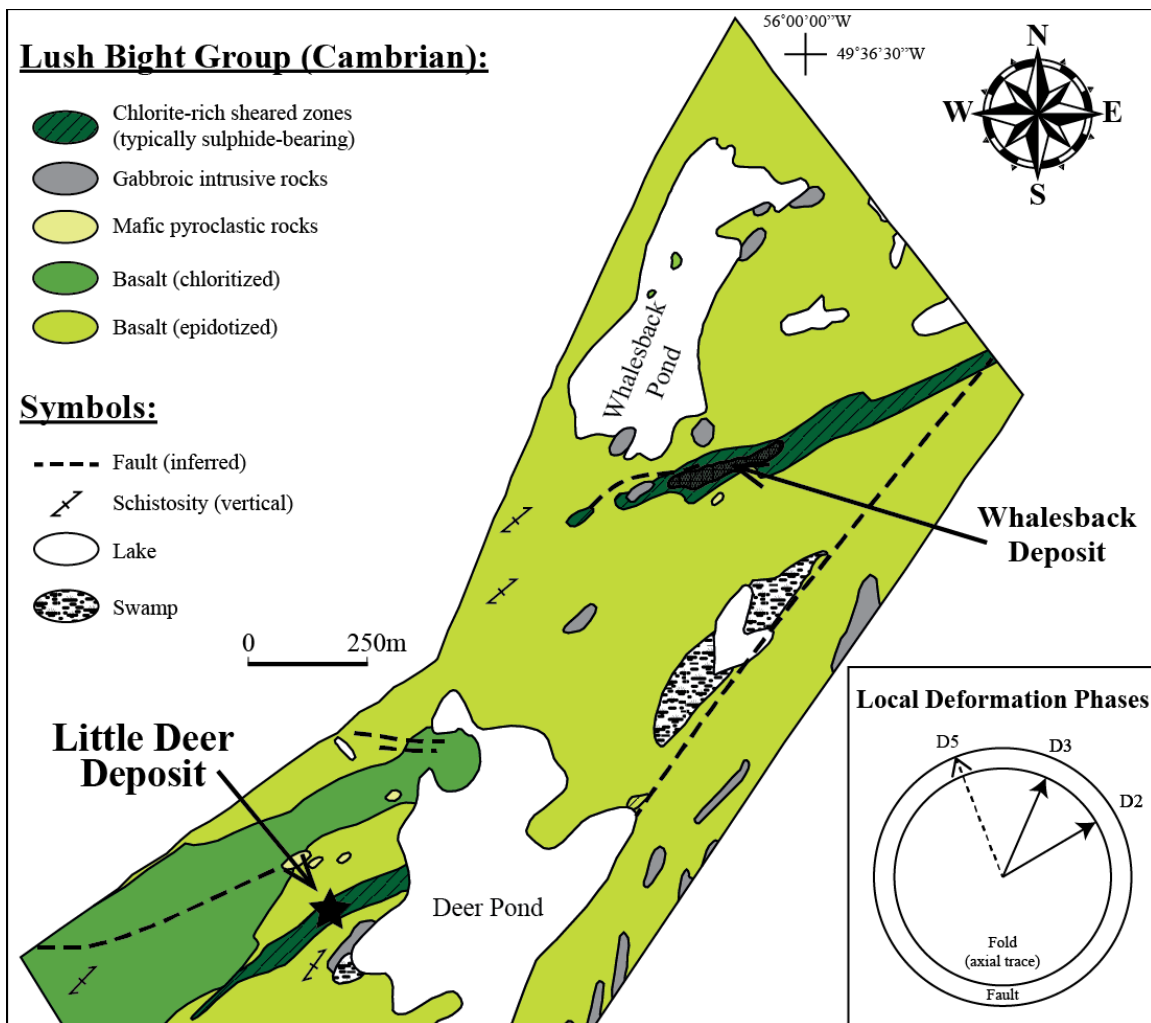


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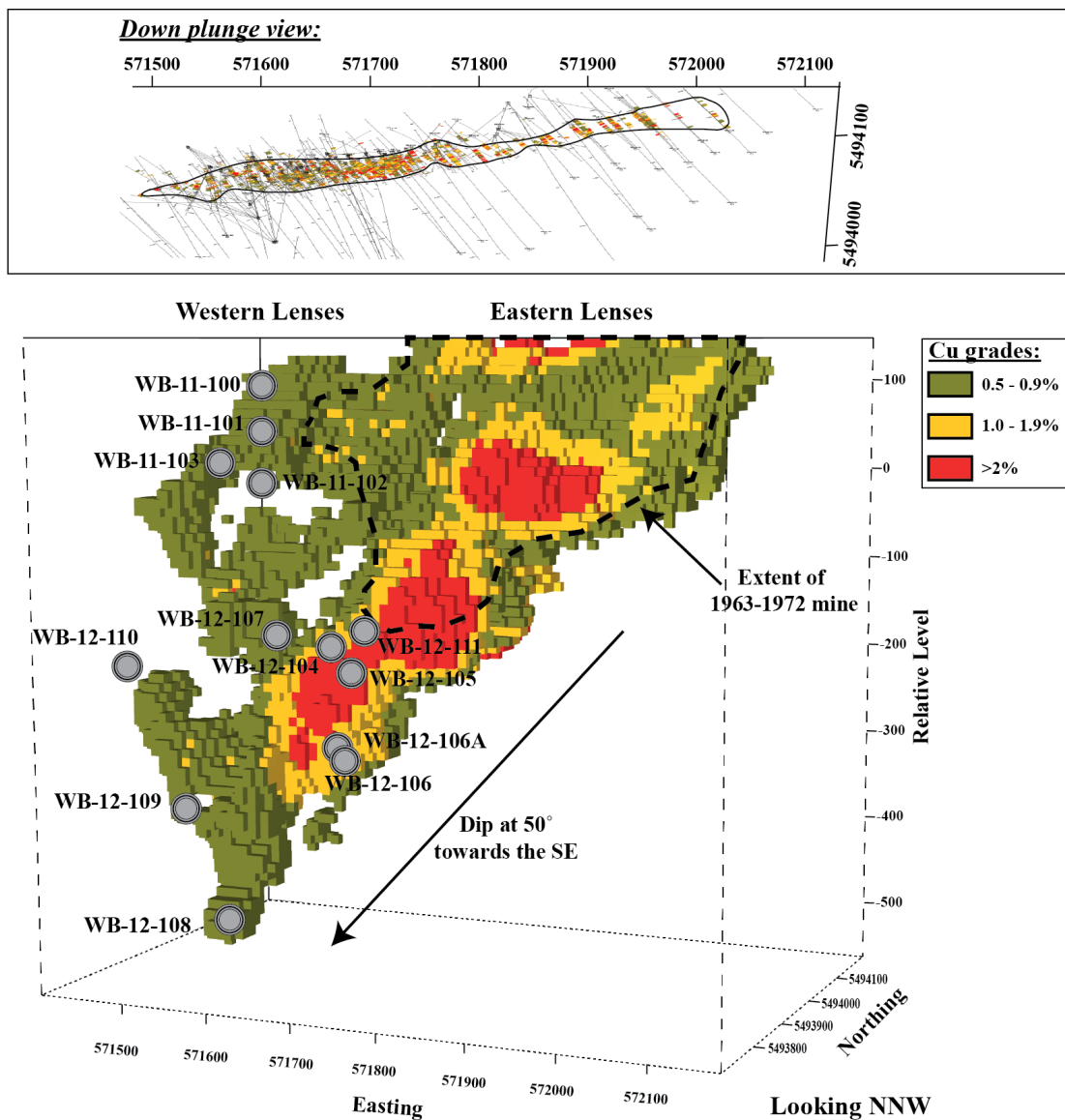
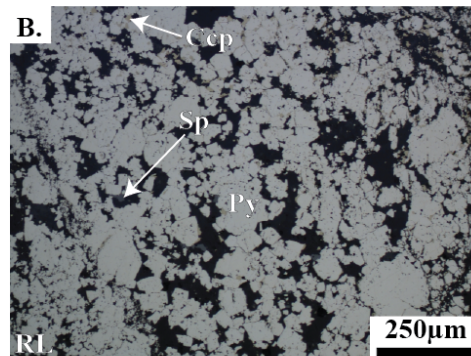
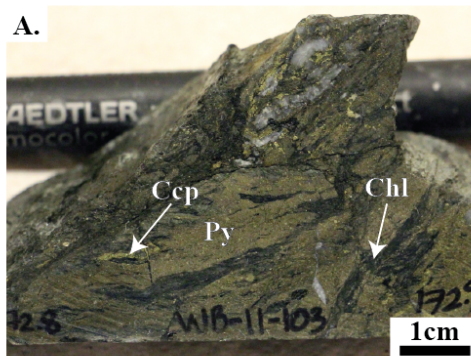


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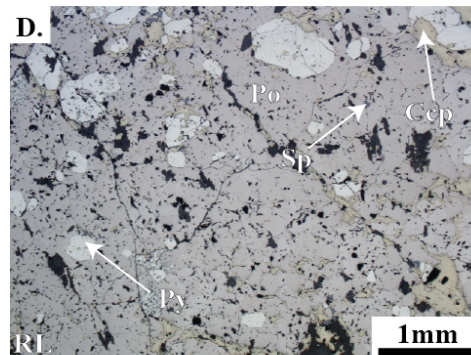
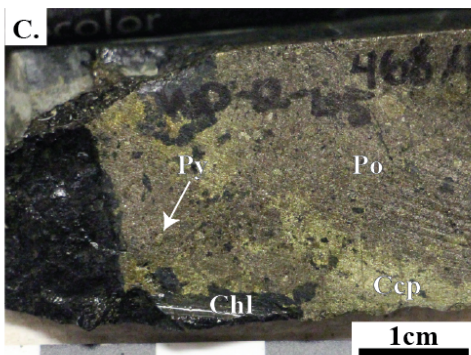
Western lenses

Pyrite-rich

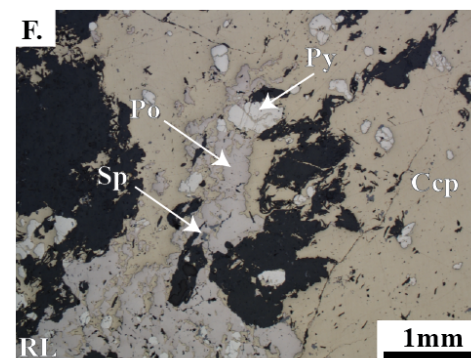
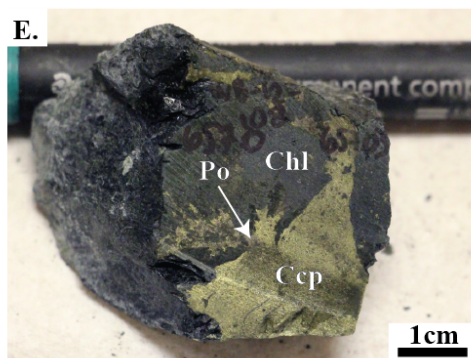


Eastern lenses

Pyrrhotite-rich



Chalcopyrite-rich



1038

1039 Figure 5

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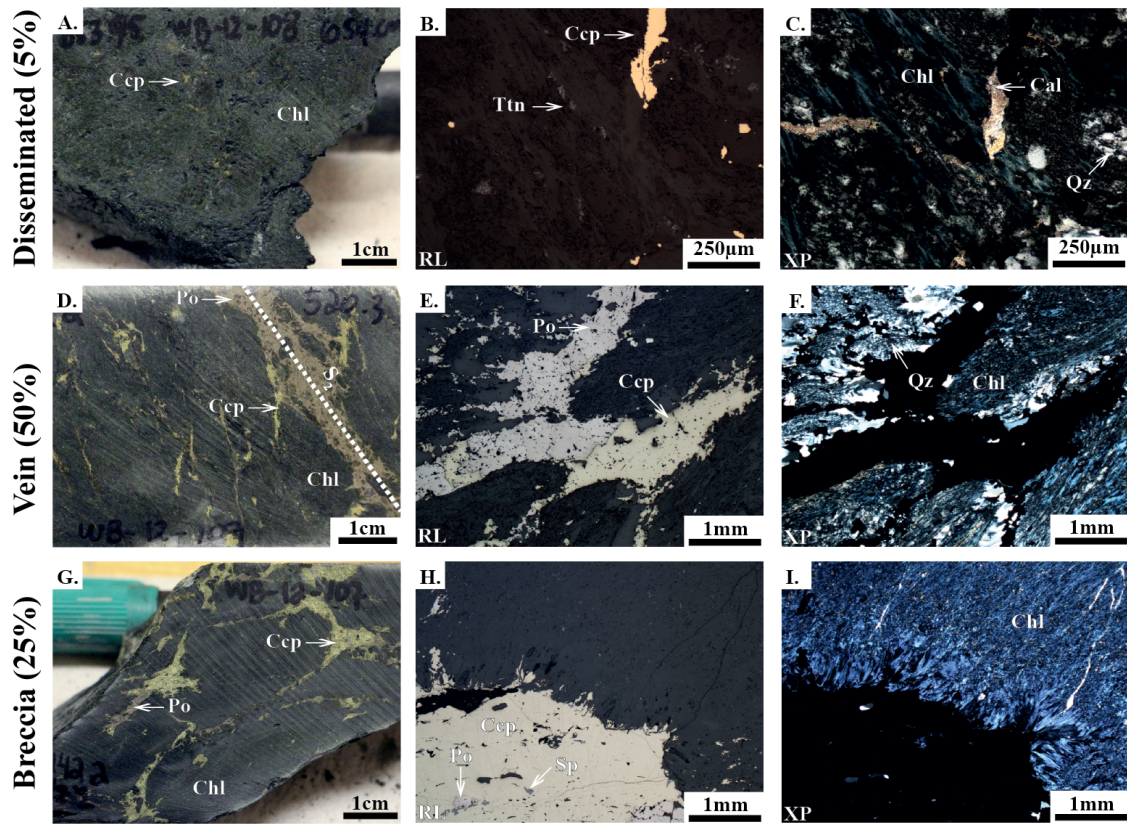


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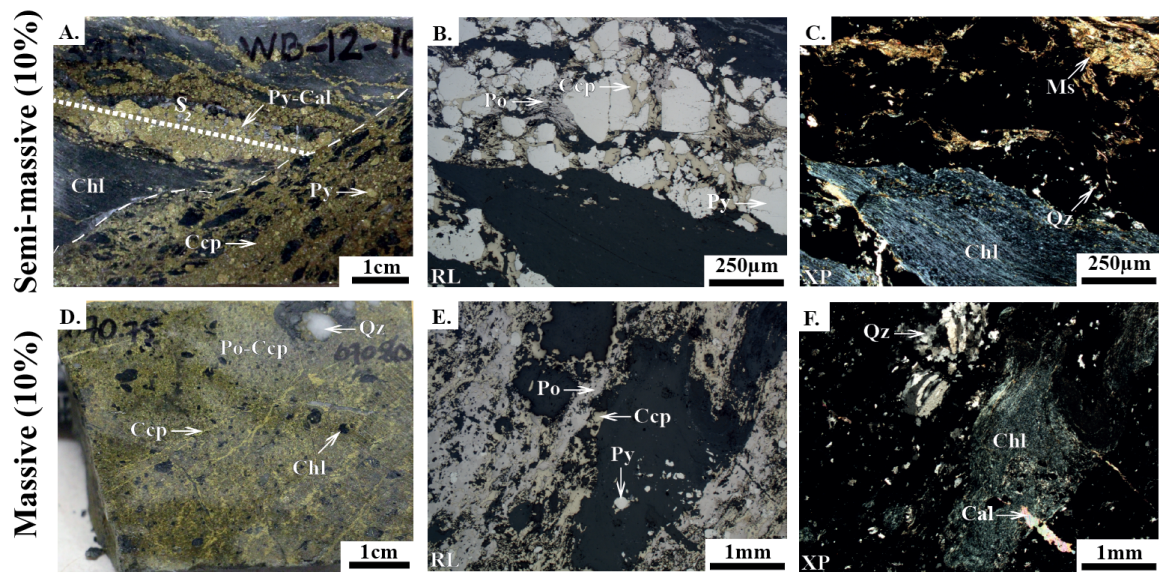


Figure 7

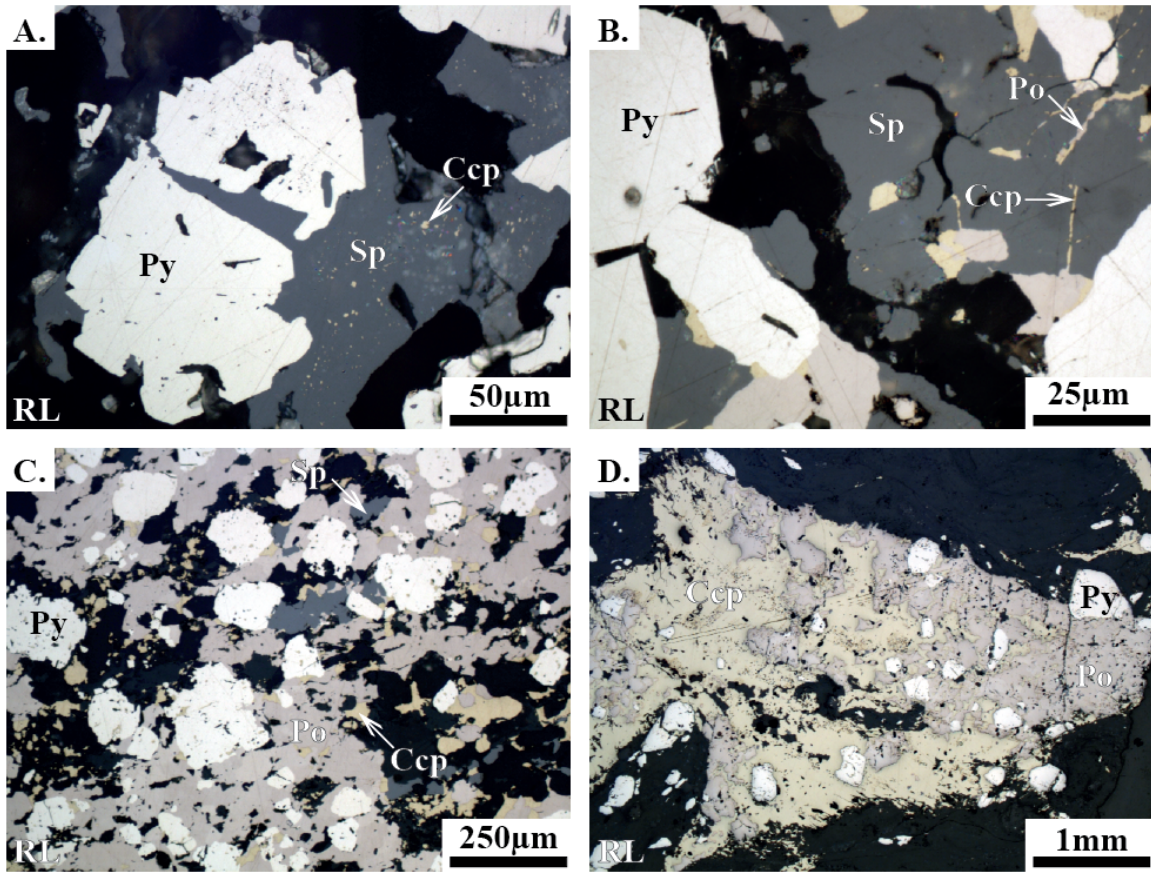
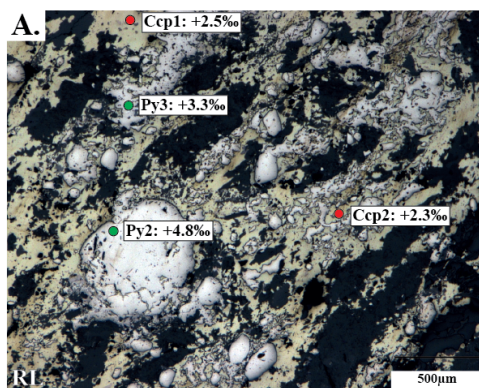
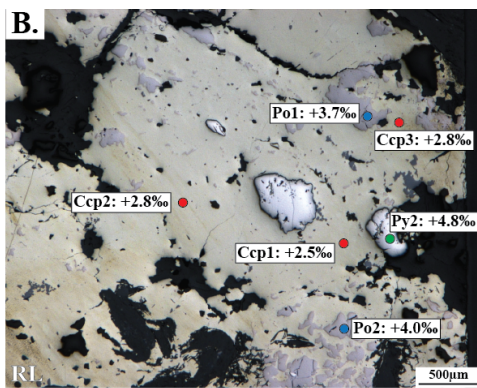


Figure 8

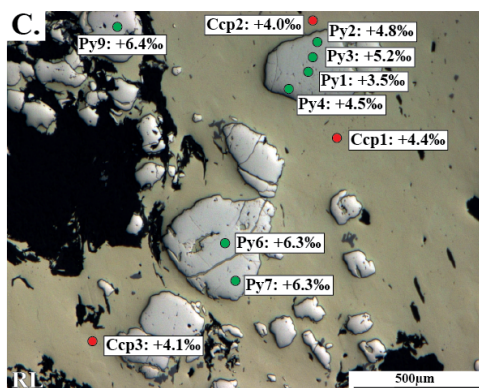
Western lenses
Massive Sulfide



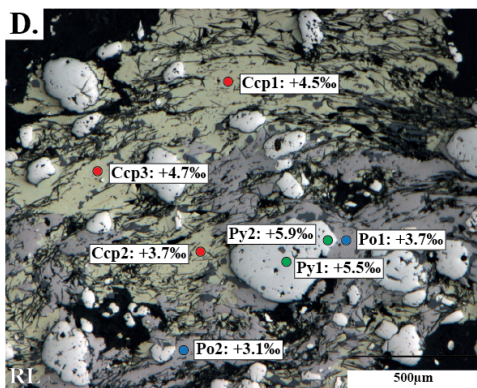
Eastern lenses
Veinlet



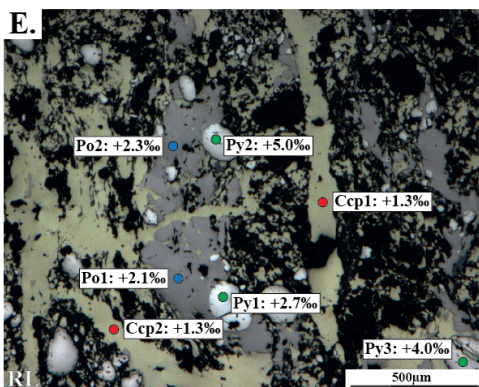
Eastern lenses
Breccia



Eastern lenses
Semi-massive Sulfide



Eastern lenses
Massive Sulfide



Eastern lenses
Massive Sulfide

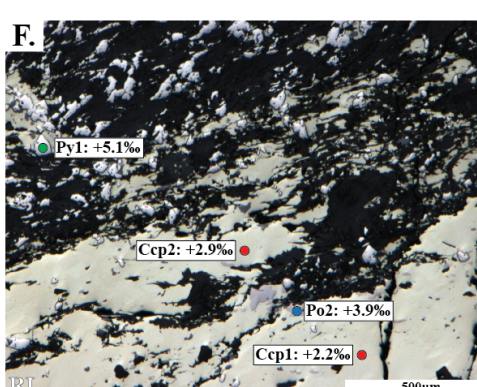
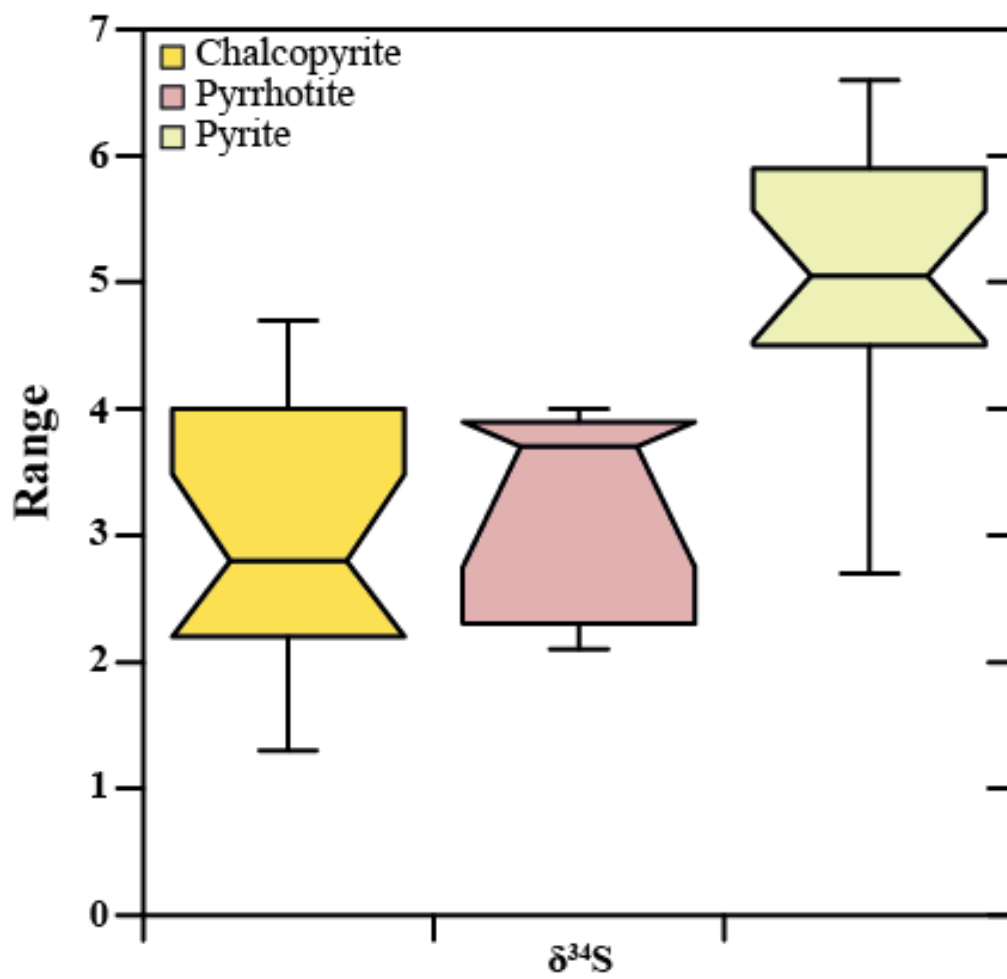


Figure 9



1053

1054 Figure 10

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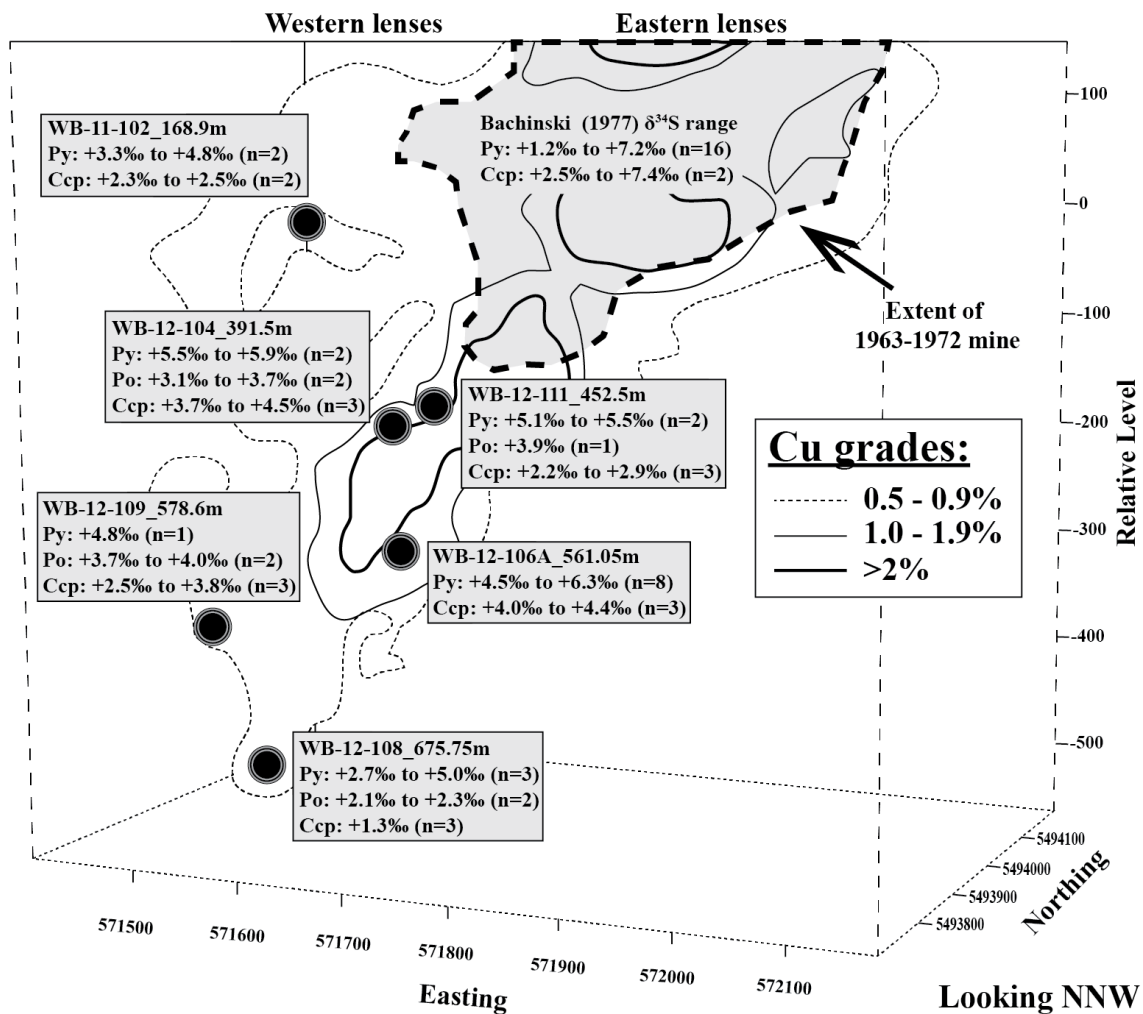
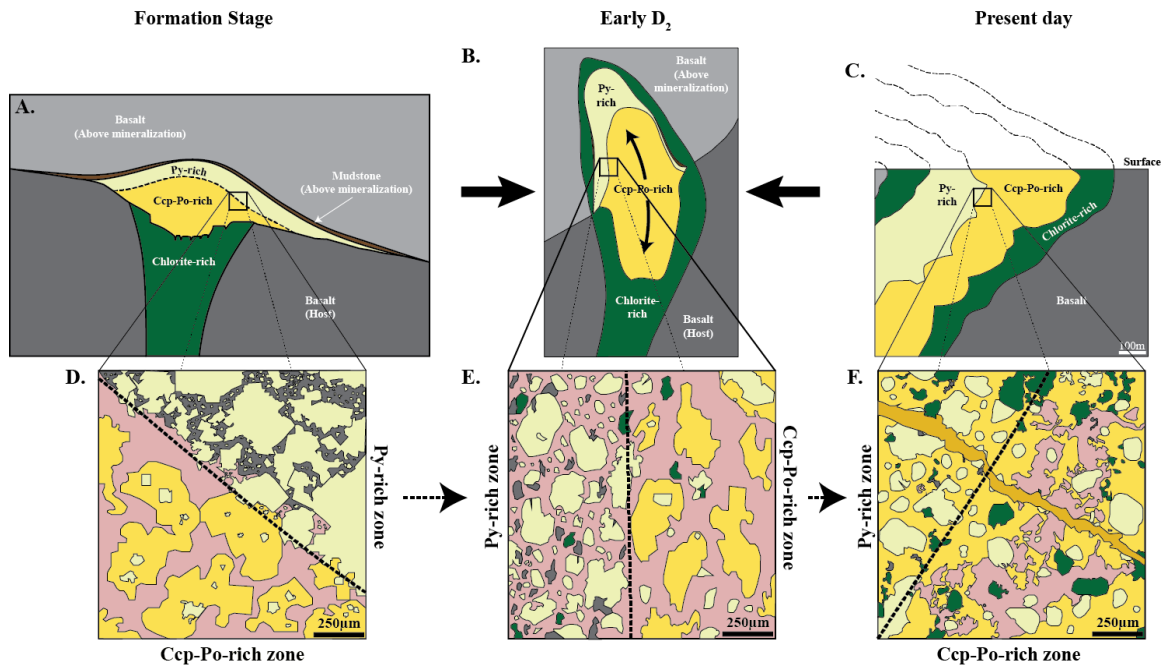
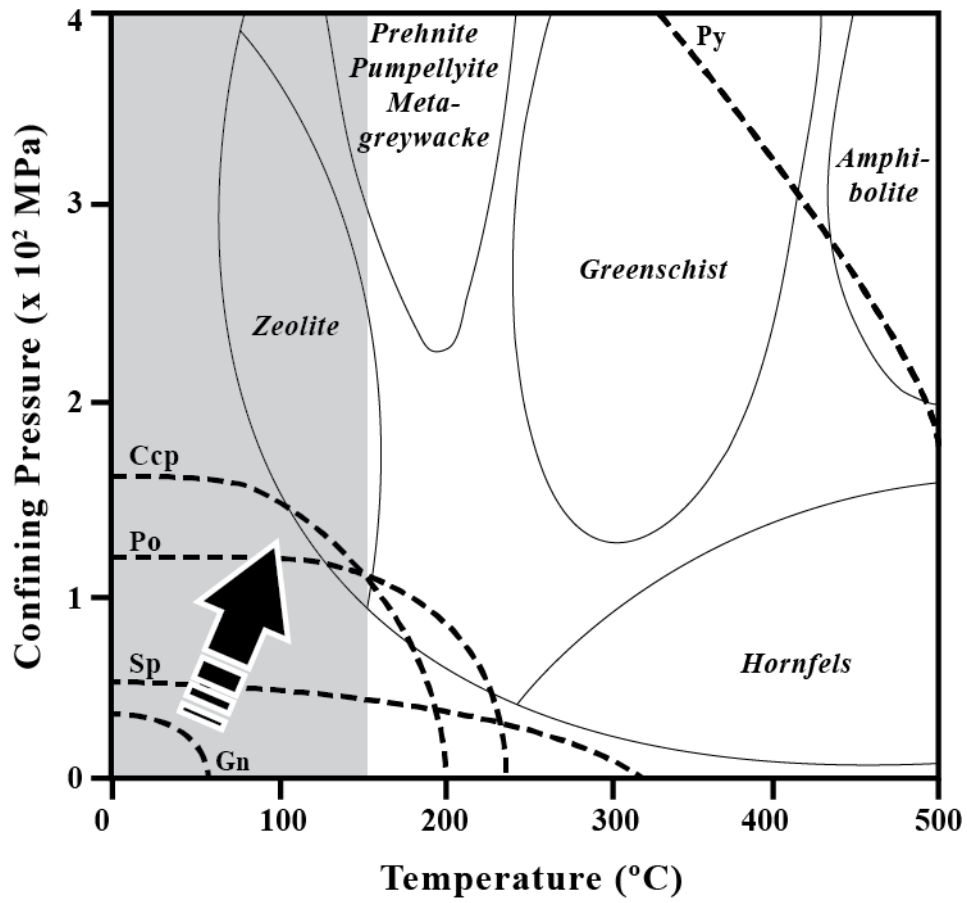


Figure 11



Legend: Pyrite Sphalerite Pyrrhotite Chalcopyrite Chalcopyrite (late; vein-type)

Figure 12



1062

1063 Figure 13

1064

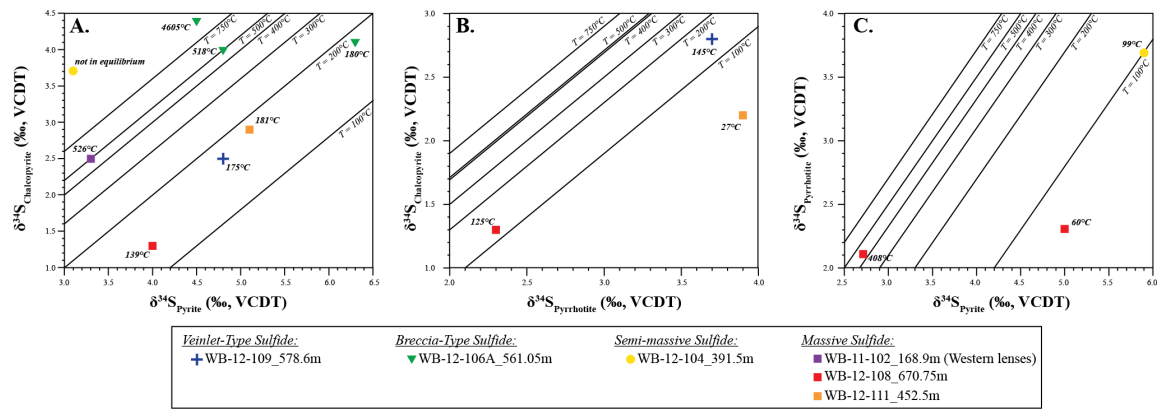


Figure 15

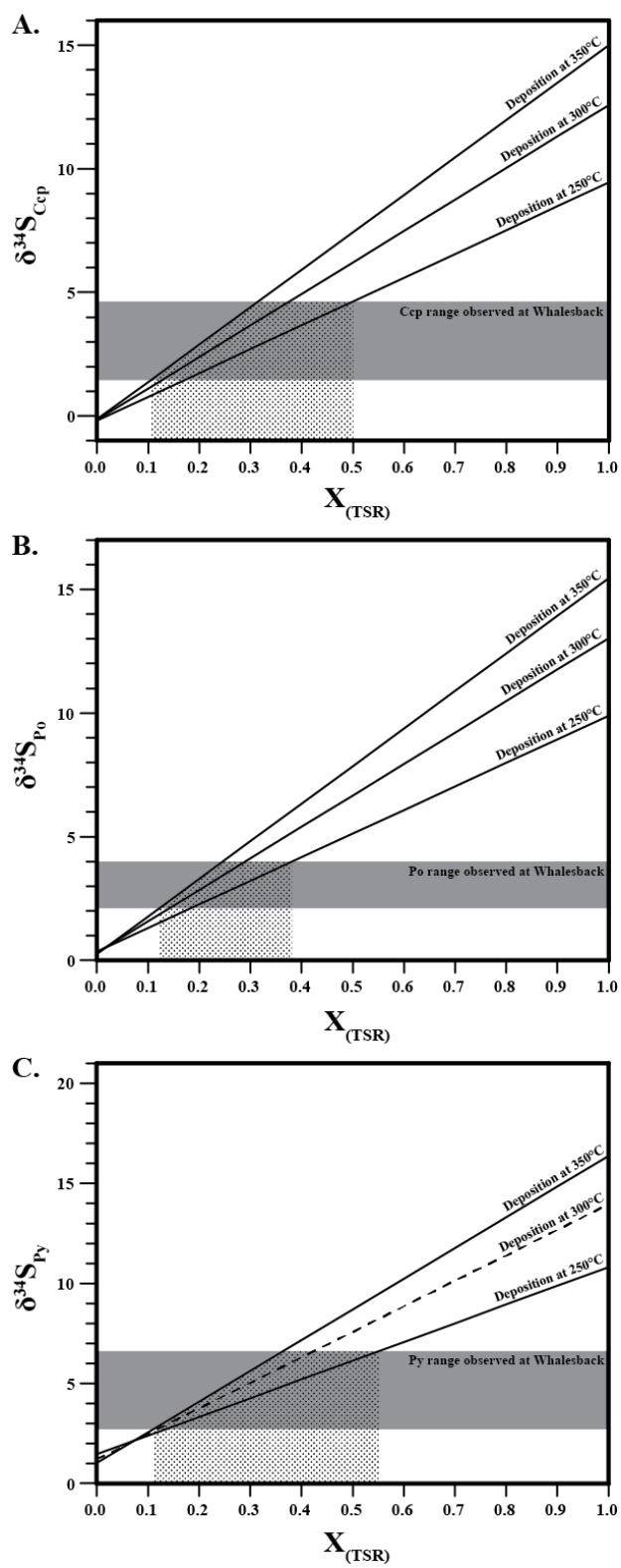


Figure 16

Table 1. SIMS $\delta^{34}\text{S}$ stable isotopic data for pyrite, pyrrhotite and chalcopyrite phases from the Whalesback deposit. (See sulfur isotopes section for analytical details)

Sample Name	Mineralized Zone	Mineralization Style	$\delta^{34}\text{S}$ (‰)	1 σ
<i>Pyrite:</i>				
WB-11-102_163.9m_Py2	Western lenses	Massive	4.8	0.4
WB-11-102_163.9m_Py3	Western lenses	Massive	3.3	0.3
WB-12-104_391.5m_Py1	Eastern lenses	Semi-Massive	5.5	0.2
WB-12-104_391.5m_Py2	Eastern lenses	Semi-Massive	5.9	0.2
WB-12-106A_561.05m_Py1	Eastern lenses	Breccia	3.5	0.2
WB-12-106A_561.05m_Py2	Eastern lenses	Breccia	4.8	0.2
WB-12-106A_561.05m_Py3	Eastern lenses	Breccia	5.2	0.3
WB-12-106A_561.05m_Py4	Eastern lenses	Breccia	4.5	0.2
WB-12-106A_561.05m_Py6	Eastern lenses	Breccia	6.3	0.3
WB-12-106A_561.05m_Py7	Eastern lenses	Breccia	6.3	0.3
WB-12-106A_561.05m_Py9	Eastern lenses	Breccia	6.4	0.2
WB-12-108_670.75m_Py1	Eastern lenses	Massive	2.7	0.2
WB-12-108_670.75m_Py2	Eastern lenses	Massive	5.0	0.3
WB-12-108_670.75m_Py3	Eastern lenses	Massive	4.0	0.2
WB-12-109_578.6m_Py2	Eastern lenses	Veinlets	4.8	0.3
WB-12-111_452.5m_Py1	Eastern lenses	Massive	5.1	0.2
WB-12-111_452.5m_Py2	Eastern lenses	Massive	5.5	0.2
<i>Pyrrhotite:</i>				
WB-12-104_391.5m_Po1	Eastern lenses	Semi-Massive	3.7	0.3
WB-12-104_391.5m_Po2	Eastern lenses	Semi-Massive	3.1	0.4
WB-12-108_670.75m_Po1	Eastern lenses	Massive	2.3	0.2
WB-12-108_670.75m_Po2	Eastern lenses	Massive	2.1	0.2
WB-12-109_578.6m_Po1	Eastern lenses	Veinlets	3.7	0.2
WB-12-109_578.6m_Po2	Eastern lenses	Veinlets	4.0	0.2
WB-12-111_452.5m_Po2	Eastern lenses	Massive	3.9	0.3
<i>Chalcopyrite:</i>				
WB-11-102_163.9m_Ccp1	Western lenses	Massive	2.5	0.2
WB-11-102_163.9m_Ccp2	Western lenses	Massive	2.3	0.4
WB-12-104_391.5m_Ccp1	Eastern lenses	Semi-Massive	4.5	0.7
WB-12-104_391.5m_Ccp2	Eastern lenses	Semi-Massive	4.7	0.4
WB-12-104_391.5m_Ccp3	Eastern lenses	Semi-Massive	3.7	0.3

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Table 1. (continued)

Sample Name	Mineralized Zone	Mineralization Style	$\delta^{34}\text{S}$ (‰)	1 σ
WB-12-106A-561.05m_Ccp1	Eastern lenses	Breccia	4.4	0.3
WB-12-106A-561.05m_Ccp2	Eastern lenses	Breccia	4.0	0.3
WB-12-106A-561.05m_Ccp3	Eastern lenses	Breccia	4.1	0.3
WB-12-108_670.75m_Ccp1	Eastern lenses	Massive	1.3	0.3
WB-12-108_670.75m_Ccp2	Eastern lenses	Massive	1.3	0.3
WB-12-108_670.75m_Ccp3	Eastern lenses	Massive	1.3	0.3
WB-12-109_578.6m_Ccp1	Eastern lenses	Veinlet	2.5	0.4
WB-12-109_578.6m_Ccp2	Eastern lenses	Veinlet	2.8	0.4
WB-12-109_578.6m_Ccp3	Eastern lenses	Veinlet	2.8	0.3
WB-12-111_452.5m_Ccp1	Eastern lenses	Massive	2.2	0.3
WB-12-111_452.5m_Ccp2	Eastern lenses	Massive	2.9	0.2
WB-12-111_452.5m_Ccp3	Eastern lenses	Massive	2.0	0.6

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