Rhenium-osmium systematics and major and trace element chemistry of cobaltite (CoAsS): Evidence for Late Mesoproterozoic sediment-hosted Co-Cu sulfide mineralization with Grenvillian and Cretaceous remobilization in the Idaho Cobalt Belt, Belt-Purcell Basin, USA

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Abstract

We report the first study of the Re-Os systematics of cobaltite (CoAsS) using disseminated and massively mineralized samples from two breccia-type and two stratabound deposits from the Co-Cu-Au Idaho Cobalt Belt, Lehmi Sub-basin to the Belt-Purcell Basin, Idaho, USA. Using a $^{185}Re+^{190}Os$ spike solution, magnetic and non-magnetic fractions of cobaltite mineral separates give reproducible Re-Os analytical data for aliquot sizes of 150 to 200 mg. Cobaltite has highly radiogenic $^{187}Os/^{188}Os$ (17–45) ratios and high $^{187}Re/^{188}Os$ (600–1800) ratios but low Re and total Os contents of ca. 0.4−4 ppb and 14−64 ppt, respectively. With proportions of radiogenic ^{187}Os between 30 and 74%, cobaltite from the Idaho Cobalt Belt is amenable to Re-Os age determination using the isochron regression approach.

The Re-Os data for disseminated cobaltite mineralization in a quartz-tourmaline breccia from the Haynes-Stellite deposit yield a Model 1 isochron age of 1349 ± 76 Ma (2σ, n = 4, mean squared weighted deviation MSWD = 2.1, initial $^{187}Os/^{188}Os$ ratio of 4.7 ± 2.2). This middle Mesoproterozoic age information was preserved despite a possible phase of metamorphism or a pulse of metamorphic-hydrothermal remobilization of pre-existing colbatite mineralization along fold cleavages during the Grenvillian orogeny. This phase of remobilization was tentatively identified through a Model 3 isochron age of 1132 ± 240 Ma (2σ, n = 7, MSWD = 9.3, initial $^{187}Os/^{188}Os$ ratio of 9.0 ± 2.9) in the quartz-tourmaline breccia from the Idaho Zone.

Although all Mesoproterozoic cobaltite mineralization was subject to lower-upper greenschist to lower amphibolites in garnet zone metamorphism from Late Jurassic to Late Cretaceous during the Cordilleran orogeny in the Blackbird district, the fine- to coarse-grained massive cobaltite mineralization from the Chicago Zone is the only studied deposit that has severely disturbed Re-Os systematics with evidence for a linear trend of mixing with (metamorphic?) fluids in the $^{187}Os/^{188}Os$ vs. $1/^{192}Os$ space. Given the position of the Chicago Zone in the Blackbird metamorphic domain characterized by middle to upper greenschist metamorphism, and, the thermal conditions of Cretaceous metamorphism between 400 and 520°C in the overlying garnet-bearing Indian Creek metamorphic domain, a maximum closing temperature of 400°C for the Re-Os isotopic system in cobaltite is proposed. In addition, our study

43   suggests that small, high-grade, structurally-controlled satellite deposits, which are located ca. 15 km to

44   the SE of the Blackbird district (e.g., Black Pine), are product of Cretaceous remobilization of

45   Mesoproterozoic cobaltite deposits in the Blackbird district.

46   The new Re-Os ages and the extremely high initial $^{187}Os/^{188}Os$ ratios are in favor of a magmato-

47   hydrothermal genetic model for a multi-stage REE-Y-Co-Cu-Au mineralization between ca. 1370 Ma and

48   1349 Ma related to the emplacement of the Big Deer Creek granite at ca. 1377 Ma. Following deposition

49   of paragenetically early xenotime, gadolinite and Be-minerals, a mixture of Mesoproterozoic evaporitic

50   brines and magmatic fluids derived metals and reduced sulfur from mafic and oceanic island-arc Archean

51   to Paleoproterozoic rocks from the Laurentian basement of the Great Falls Tectonic Zone. Cobaltite

52   mineralization (and the closure of the Re-Os isotope system in this mineral) occurred upon cooling of

53   these fluids at an inferred temperature of 300°C or below.

54

1. Introduction

57

58   Cobaltite –CoAsS–, which crystallizes in the orthorhombic system, belongs to the group of

59   sulfarsenides along with gersdorffite (NiAsS, cubic) and arsenopyrite (FeAsS, monoclinic). Cobaltite is

60   found in a variety of ore deposits, including mafic to ultramafic-hosted massive sulfide deposits (e.g.,

61   Main Uralian Fault Zone, southern Urals; Nimis et al., 2008), orogenic gold deposits (e.g., Oberon gold

62   prospect, Northern Territory, Australia; Cook et al., 2013; Huston et al., 2007; Tunks and Cooke, 2007;

63   Loulo mining district, Mali; Lawrence et al., 2013; Otago and Alpine Schists, New Zealand; Pitcairn et al.,

64   2006), and sediment-hosted Co-Cu deposits of the Idaho Cobalt Belt, Idaho, USA (cf. Bookstrom et al.,

65   2016). Cobaltite is also known to crystallize with gersdorffite from immiscible sulfarsenide liquid from

66   calc-alkaline magma (Beziat et al., 1996).

67   Other (sulf-)arsenides such as arsenopyrite and löllingite (FeAs$_2$, orthorhombic) have been proven

68   to be hosts for both rhenium (Re) and osmium (Os) with Re concentrations in the order of 10 to 1000s of

69   ppb, and, Os concentrations in the 10-100s ppt range and up to the ppb level. The use of the Re-Os

70   isotopic system in these minerals with isochron age determinations constrained the timing of gold

71   mineralization in hydrothermal gold deposits (Mikulski et al., 2005; Davies et al., 2010; Morelli et al.,

72   2010) and also brought new insights into the metamorphic evolution of the Broken Hill area in Australia

73   (Saintilan et al., *accepted in Canadian Mineralogist*). Hence, these results and the similarity of crystalline

74   structure between cobaltite and löllingite form a logical basis to investigate the Re-Os isotope system in

75   cobaltite and evaluate the potential of this mineral for Re-Os age determination.

76   In this study, cobaltite was sampled from Co-Cu-Au deposits hosted by metasedimentary rocks of

77   the Idaho Cobalt Belt, Idaho, USA (Fig. 1). This metallogenic province was chosen as the best example

78   for which the parameters of basin stratigraphy, structural and tectonic features resulting from basin

79   inversion, compressive tectonics and related metamorphic history are well-constrained by geologic and

80   geochronological evidence (cf., Lund and Tysdal, 2007; Zirakparvar et al., 2007; Lund et al., 2011;

81   Aleinikoff et al., 2012; Bookstrom et al., 2016). We chose to focus on the Haynes-Stellite deposit and the

82   Chicago and Idaho Zones of the Blackbird mine, and the Black Pine deposit to the southeast of the

83   Blackbird area (Fig. 1B). However, no geochronological data exist on the Co- and Cu-rich sulfides in

84   these deposits and both Proterozoic (Aleinikoff et al., 2012, Slack, 2012; Bookstrom et al., 2016) and

85   Cretaceous (Lund et al., 2011) ages for mineralization have been proposed. The unresolved issue of the

86   timing of mineralization feeds the debate on the genetic model for Co-Cu-Au mineralization in the Idaho

87   Cobalt Belt (cf., Nold, 1990; Slack, 2006; Lund et al., 2011; Slack, 2012).

88   Herein, we carry out a two-fold Re-Os investigation of cobaltite mineralization: (1) to assess the

89   degree of preservation of the Re-Os systematics of cobaltite in the four deposits through Cretaceous

90   greenschist to lower amphibolites metamorphic overprint; (2) to determine if cobaltite mineralization is

91   Proterozoic or Cretaceous in age for each of the deposits. Given the framework in which Re-Os isotope

92   data can be assessed and interpreted further by using available geologic, petrographic and stable and

93   radiogenic isotope data (Slack, 2006; Slack, 2012; Johnson et al., 2012; Panneerselvam et al., 2012;

94   Bookstrom et al., 2016), the new Re-Os ages for cobaltite mineralization provide a better understanding of

95   the origin of these deposits and the derived initial Os ratio further constrains the source of Os, and by

96   inference, other metals like Co.

97

98   **2.   Geologic background and chronostratigraphy of the Idaho cobalt belt**

99   The Idaho cobalt belt (ICB; Fig. 1), in the Salmon River Mountains of east-central Idaho, U.S.A.,

100   is a northwest-trending belt of 45 known cobalt (Co) ± copper (Cu) ± gold (Au)-bearing mineral

101   occurrences (Johnson et al., 1998). Mineralization is hosted by metasedimentary rocks of the Lemhi sub-

102   basin in the southwestern part of the Mesoproterozoic Belt-Purcell basin, which is east of and inboard

103   from the rifted continental margin (Fig. 1A). The rocks of the western Belt-Purcell basin underwent

104   regional metamorphism during the Middle to Late Mesoproterozoic (Ectasian and Stenian) and the

105   Jurassic−Tertiary interval, whereas magmatism occurred during Ectasian, Neoproterozoic,

106   Cambrian−Ordovician, and Triassic−Tertiary intervals (Bookstrom et al., 2016 and references therein).

107   Co-Cu-Au occurrences of the ICB are hosted by the coarse siltite and banded siltite units of the

108   Apple Creek Formation and the Gunsight Formation of the Lemhi Group (Fig. 1B; Fig. 2). In the ICB, the

109   coarse siltite unit consists mostly of biotitic siltite whereas the banded siltite unit comprises mostly

110   interlayered siltite and biotite phyllite to schist (after argillite; Bookstrom et al., 2016). Depositional ages

111   of these rocks are bracketed by the youngest sets of detrital zircons in samples from the underlying

112   Hoodoo Formation (ca. 1445 Ma; Link et al., 2007) and from the uppermost part of the banded siltite unit

113   (ca. 1409 Ma; Aleinikoff et al., 2012).

114   The granite pluton of Big Deer Creek, which bounds the northeastern margin of the ICB, belongs

115   to a bimodal suite of A-type megacrystic granite and meta-gabbroid intrusions that was dated at 1377 ± 4

116   Ma (Aleinikoff et al., 2012). Gabbroid to syenitoid plutons (dated at 485−530 Ma; Lund et al., 2011;

117   Gillerman, 2008) and chemically similar mafic dikes represent within-plate magmatism, inboard from the

118   rifted continental margin, associated with early subsidence and growth of the Cordilleran miogeocline in

119   Cambrian−Ordovician time (Evans, 1984; Lund et al., 2010; Bookstrom et al., 2016).

120   Structurally, the metasedimentary host rocks to the ICB, the granite pluton of Big Deer Creek, and

121   the syenitoid Deep Creek plutons are within the Poison Creek thrust plate which is bounded by the Poison

122   Creek and Iron Lake thrust faults of Cretaceous age (Fig. 1B, Fig. 3). Similarly, the Blackbird Co-Cu-Au

123   deposit lies within a structural block between the Little Deer and White Ledge faults of Cretaceous age

124   (Fig. 1B). Nold (1990) recognized that ore minerals and metasedimentary host rocks of the ICB are

125   increasingly deformed, metamorphosed, and recrystallized from southeast to northwest (and from the fold-

126   thrust belt to the hinterland of the East Kootenay orogen, as shown in Fig. 1A). Phyllitic biotite-chlorite-

127   bearing rocks of greenschist facies at the southeast end of the ICB pass northwestward to schistose garnet-

128   biotite-bearing rocks of lower-amphibolite facies in the northwest part of the Blackbird mine area, and to

129   gneissic sillimanite-garnet-biotite-quartz rocks of upper-amphibolite facies at the northwest end of the

130   ICB (Bookstrom et al., 2016). Inasmuch as the lower boundary of the garnet zone is gradational, it appears

131   that the thermal gradient increases upward, possibly in the footwall of a hotter thrust plate that overrode

132   the northwestern part of the ICB before being removed during uplift and erosion (Fig. 3). As a result of

133   this tectono-thermal setting, three metamorphic and/or structural domains in this order upwards are

134   sandwiched between the Poison Creek and Indian Creek faults: (1) the lowermost domain called the

135 "Haynes-Stellite domain" features lower to middle greenschist facies rocks, (2) then the "Blackbird

136 domain" is characterized by middle to upper greenschist facies rocks and contains the Idaho and Chicago

137 zones within the Blackbird mine, (3) and finally, the Indian Creek metamorphic domain characterized by

138 upper greenschist to lower amphibolites rocks, and rocks above garnet isograd (Fig. 3; cf. Lund et al.,

139 2011).

140

3. Mineralization, available geochronological constraints and sampling

142 *Blackbird mine area*

143 The Blackbird mine is central to the ICB and contains at least 8 strata-bound and semi-concordant

144 ore zones (e.g., Chicago and Idaho zones), several strata-bound prospects (e.g., Merle), and many

145 prospects comprising discordant veins (e.g., Sunshine), mineralized breccias, and small replacement-style

146 occurrences. A minimum estimate of total resources of the Blackbird district was reported by Slack (2013)

147 as 16.8 Mt of ore, averaging 0.73% Co, 1.37% Cu, and 1 ppm Au. This estimate does not include

148 downward projections of ore zones that are open down-dip.

149 In the Blackbird mine area, stratabound cobaltite-biotite ore zones ± tourmaline are preferentially

150 hosted in biotite-rich layers in banded siltite, which is folded into large- and small-scale chevron-style

151 folds that plunge moderately northward. The Idaho Zone is located in the tight hinge zone of a large-scale

152 syncline whereas the Chicago Zone lies in a thickened limb. Unaltered banded siltite consists of

153 interlayered siltite and argillite, but in cobaltite-biotite ore, argillite is "altered" to biotitite, containing

154 >75% of greenish-black biotite that is unusually rich in ferrous iron and chlorine, and is interpreted as

155 hydrothermal biotite (Bookstrom et al., 2016). Two samples for Re-Os analysis were taken from (1) the

156 tabular and vein-like body of the Chicago zone (CH) where alternating bands of massive and coarse-

157 grained and fine-grained cobaltite are surrounded by greenish-black biotite, and, (2) the quartz-

158 tourmaline-cobaltite breccias in the Idaho zone (ID; Bookstrom et al., 2016; Fig. 1B, Fig. 4).

159   Oscillatory-zoned xenotime grains intergrown with biotite in cobaltite-biotite ore, which formed

160   at 1370 ± 4 Ma as a late stage of hydrothermal activity related to intrusion of the Big Deer Creek granite,

161   are found as inclusions in cobaltite at the Merle prospect (in situ U-Pb SHRIMP ages, Aleinikoff et al.,

162   2012). These authors proposed that the first generation of cobaltite is thus younger than ca. 1370 Ma.

163   Slack (2012) proposed that the 1370 Ma-old hydrothermal system responsible for xenotime growth may

164   have also triggered precipitation of cobaltite (with xenotime inclusions), chalcopyrite and gold at a later

165   stage. Additional Proterozoic xenotime precipitation occurred at 1315 to 1270 Ma and ca. 1050 Ma.

166   However, no introduction of metals is reported during these time intervals in the Belt-Purcell basin

167   (Aleinikoff et al., 2012).

168   Mafic dikes of probable Cambrian−Ordovician age cut older cobaltite-biotite lodes but are in turn

169   cut by younger quartz-siderite-Fe-Cu-sulfide veins and breccias ± minor cobaltian arsenopyrite. These Fe-

170   Cu-sulfide-bearing veins and breccias are not folded and lack metamorphic fabrics (Bookstrom et al.,

171   2016). Lund et al. (2011) determined a $^{40}Ar/^{39}Ar$ age of 83 ± 1 Ma for muscovite selvage around a pyrite-

172   chalcopyrite-quartz vein at the Sunshine prospect. This Upper Cretaceous age is in accord with late Early

173   Cretaceous (110 ± 3 Ma) to early Late Cretaceous (92 ± 5 Ma) in situ U-Pb SHRIMP ages of monazite

174   enclosing cobaltite in places at the Merle prospect (Alenikoff et al., 2012). This Cretaceous hydrothermal

175   event also triggered the growth of xenotime grains or rims in conjunction with cobaltite. Garnets from the

176   northwestern part of the ICB yield Lu-Hf late Upper Jurassic to Upper Cretaceous ages of ca. 151 to 94

177   Ma (Zirakparvar et al., 2007). Garnet porphyroblasts in ore zones commonly contain cobaltite inclusions,

178   which is consistent with a pre-Cretaceous age for cobaltite-biotite ore.

179

180   *Haynes-Stellite mine*

181   The Haynes-Stellite mine (HS in Fig. 1B) produced a small tonnage of cobaltite-rich ore from a pipe-like

182   ore body in a NW-striking and steeply-dipping tabular quartz-tourmaline breccia, which cuts in quartzite

183   with thin interbeds of biotite phyllite, near the base of the Gunsight Formation (which overlies the banded-

184   siltite unit). The dated sample (Fig. 4) contains aligned lenticular clasts of quartz and quartz-tourmaline

185   breccia in a matrix of tourmaline with cobaltite stringers, interpreted to indicate streamlined flow foliation

186   (Bookstrom et al., 2016).

187

188   *Black Pine mine and prospects*

189   The Black Pine mine and neighboring prospects are about 14 km southeast of the Blackbird mine (Fig.

190   1B), where they are hosted in thin-bedded banded siltite, consisting of interlayered siltite and locally

191   chloritized biotite phyllite (sample in Fig. 4). The lower level of the Black Pine mine followed thin

192   concordant and discordant veins of the strata-bound copper zone, and the upper level followed the strata-

193   bound cobalt zone. These mineralized zones and their host strata generally strike northwest and dip about

194   60º northeast. Chalcopyrite, pyrite, and arsenopyrite are the main ore minerals, but quartz, siderite,

195   cobaltian arsenopyrite, cobaltite, and magnetite siderite are also present in some veins and mineralized

196   breccias (Nold, 1990; Bookstrom, 2013).

197

198   **4.   Material and methods**

199   **4.1   Sample preparation**

200   A total of four samples of cobaltite mineralization from the Idaho Cobalt belt (Fig. 4) were

201   processed prior to Re-Os isotope geochemistry: 1) tourmalinized breccia of quartz, biotite, K-feldspar,

202   xenotime cemented by disseminated brownish pink cobaltite at the Haynes-Stellite deposit (HS, Fig. 4a);

203   2) layers of massive coarse-grained to fine-grained brownish pink cobaltite ± xenotime at the Chicago

204   Zone (CH, Fig. 4b), Blackbird mine; 3) layers of disseminated brownish pink cobaltite in a biotite, quartz,

205   K-feldspar rock at the Idaho Zone (ID, Fig. 4c), Blackbird mine; and, 4) coarse-grained disseminated

206   shiny grey cobaltite in a quartz-chlorite rock, Black Pine deposit (BP, Fig. 4d).

207   All samples were cut into slabs that were thoroughly cleaned using silicon carbide grit and paper to

208   remove any metal traces left by hammering or sawing (Fig. 5). These samples were crushed using agate

209   mortar and pestle and sieved through disposable home-made nylon sieves to produce 70−200 and +70

210 mesh size fractions. Following heavy liquid separation using Sodium Polytungstate (SPT, specific gravity

211 of 2.86), a Frantz Isodynamic Separator was used to produce magnetic (M) and non-magnetic (NM)

212 cobaltite fractions from the 70−200 mesh fractions by applying successive 0.3, 0.6 and 0.9 amp currents

213 for all samples (except the Black Pine sample for which 1.8 and 2.1 amp currents had to be applied) with

214 15° side slope and 10° forward slope. The composition of each mineral separate was verified by X-ray

215 diffraction (XRD) analyses. Aliquots of each magnetic fraction were embedded in epoxy and the mounts

216 were studied by scanning electron microscopy (SEM) using a Zeiss Sigma 300 Field Emission SEM (VP-

217 FESEM) operated in backscattered electron mode (SEM-BSE, beam conditions of 15kV). In addition,

218 a Bruker energy dispersive X-ray spectroscopy (EDS) system with dual silicon drift detectors each with an

219 area of 60 mm^2 and a resolution of 123 eV was used for single spot analysis to further control the

220 mineralogy of each mineral separate. After magnetic separation, the final mineral separates contained

221 ~50−100% cobaltite, 0−50 % non-sulfide gangue minerals (e.g., xenotime, quartz, biotite) in all samples,

222 and up to 1% native gold and <0.5% Bi-telluride in the case of the Chicago Zone sample.

223

224 **4.2   Cobaltite petrography and major and trace element chemistry of cobaltite**

225   Detailed petrography of each cobaltite fraction mounted in epoxy was complemented by major

226 and trace element chemical analysis. Compositional analyses were performed on a Cameca SX-100

227 electron microprobe equipped with 5 tunable wavelength dispersive spectrometers at the Department of

228 Earth & Atmospheric Sciences, University of Alberta, Edmonton, Canada. Operating conditions were 40°

229 take-off angle and a beam with energy of 20 keV, a current set at 20 nA, and a diameter of 2 microns.

230 Elements were acquired using analyzing crystals LLIF for As Kα, Co Kα, Cu Kα, Fe Kα, Zn Kα, Ni Kα,

231 PET for Te Lβ, Bi Mα, Te Lβ (2), Bi Mα (2), S Kα, and LTAP for Se Lβ. The standards were Ni metal for

232 Ni Kα, Cu metal for Cu Kα, Co metal for Co Kα, gallium arsenide GaAs for As Kα, bismuth telluride

233 Bi_2Te_3 for Bi Mα, Te Lβ, Bi Mα, Te Lβ, gallium selenide GaSe for Se Lβ, iron disulfide FeS_2 for Fe Kα,

234 and sphalerite ZnS for S Kα, Zn Kα. The counting time was 30 seconds for each element with an off peak

235 counting time of 30 seconds. Off peak corrections were linear for all elements. Unknown and standard

236   intensities were corrected for deadtime whereas standard intensities were corrected for standard drift over

237   time. Interference corrections were applied to S for interference by Co, and Te for interference by Cu (cf.

238   Donovan et al., 1993). Detection limits (at the 99% confidence level) are 0.02 wt% for Co, Cu, S, and Zn,

239   0.01 wt% for Fe, 0.03 wt% for Ni, 0.05 wt% for As, 0.07 wt% for Te, 0.09 wt% for Bi and 0.12 wt% for

240   Se. The matrix correction method was ZAF or Phi-Rho-Z Calculations and the mass absorption

241   coefficients dataset was LINEMU. The ZAF or Phi-Rho-Z algorithm utilized was Conventional

242   Philibert/Duncumb-Reed (cf. Armstrong, 1988).

243

244   ## 4.3   Re-Os analytical procedure

245   For each analysis, between 50 and 350 mg of cobaltite mineral separates was weighed and

246   transferred into a thick-walled borosilicate Carius tube. Each sample was dissolved in inverse aqua regia

247   (~2mL of 10 N HCl and ~6 mL 16 N HNO_3) with a known amount of $^{185}Re+^{190}Os$ spike solution at 210ºC

248   for 24 hours (Fig. 5). The full Re-Os laboratory protocol used in the present work is described in full in

249   Hnatyshin et al. (2016). Re and Os isotopic compositions were determined by negative thermal ionization

250   mass spectrometry (NTIMS) using a ThermoScientific Triton mass spectrometer at the Canadian Centre

251   for Isotopic Microanalysis, University of Alberta, Edmonton, Canada. Rhenium was measured as ReO_4^- in

252   static mode on Faraday collectors, whereas Os was measured as OsO_3^- in peak-hopping mode on SEM

253   with a constant flow of oxygen (Creaser et al., 1991; Völkening et al., 1991; Hnatyshin et al., 2016).

254   Measurement quality was monitored by repeated measurements of in-house Re ($^{185}Re/^{187}Re = 0.59774 \pm$

255   0.00065, n = 23) and Os ("AB-2 standard", $^{187}Os/^{188}Os = 0.10682 \pm 0.00009$, n = 100) standard solutions.

256   Total procedural blanks for each set of samples are reported in Table 1.

257   All Re-Os ages are reported as Model 1 or Model 3 isochrons through regression in $^{187}Os/^{188}Os$ vs.

258    $^{187}Re/^{188}Os$ space of the Re-Os data which are reported at the 2σ level (95% level of confidence). The

259   *Isoplot* v 4.15 program (Ludwig, 2011) was used for isochron regression by considering the ^{187}Re decay

260   constant of $1.666e^{-11}$ year^{-1} (Smoliar et al., 1996).

261

5.   Results

5.1   Petrography of the cobaltite mineral fractions

SEM-BSE images of the different cobaltite mineral fractions studied are shown in Fig. 6. Mineral separates (M0.6, M0.9 and NM0.9) from the Chicago Zone are characterized by densely packed cobaltite grains (up to 15 µm in width) with about 10 to 20% gangue minerals (mainly biotite) in the M0.6 and M0.9 fractions (Figs. 6a and b). The NM0.9 fraction corresponds to 95 % cobaltite with up to 5% gold whereas Bi-tellurides make up to 1−2% of the M0.6 and M0.9 separates (Figs. 6a, b and c). The unique M0.9 mineral fraction from Haynes-Stellite corresponds to cobaltite (70−80%, between 10 and 50 µm in width), biotite, K-feldspar, quartz and xenotime (Figs. 6d and e). Xenotime postdates precipitation of cobaltite. Cobaltite crystals have sharp edge when in contact with gangue minerals but xenotime with which it is in contact along irregular, dented edges. The M0.3 and M0.6 mineral fractions from the Idaho Zone comprise ca. 5 to 10 % euhedral to subhedral cobaltite (between 5 to 35 µm in width) embedded in gangue minerals, being mainly quartz, sulfur-bearing Na-silicate (tentatively suggested to being scapolite with a -SO_4 component) and up to 3 to 4 % xenotime (Figs. 6g and h).

276

5.2   Major and trace element chemistry of cobaltite mineral fractions

A total of 159 analyses were performed on the 70−200 mesh size mineral fractions of cobaltite from the Chicago and Idaho Zones, the Haynes-Stellite deposit and the Black Pine prospect (Electronic Supplementary Material 1). Besides the major elements (Co, As, and S), Fe and Ni were detected in all studied fractions whereas Zn was sporadically detected (values < 0.03 wt%) and Cu was only detected in the fractions from the Haynes-Stellite deposit (< 0.55 wt%, Table 2). Selenium was not detected at all. Tellurium values are below 0.10 wt% whereas Bi values reach 0.2 wt% at maximum. Based on the mineralogy of the cobaltite mineral fraction, it is suggested that these elements are contributed by micro-inclusions of Bi-telluride, particularly in the case of the Chicago Zone (Fig. 6). Modal mineral formulae

286   were calculated for each fraction. The various fractions of each mineral deposit have similar compositions

287   despite variable magnetic susceptibilities (Table 3). When considering the relative positions of the mineral

288   deposit in the NW−SE striking Idaho Cobalt Belt and the metamorphic grade of each mineral deposit

289   (Figs. 1B and 3), one notices that cobalite in the north-westernmost Chicago Zone (middle to upper

290   greenschist facies) has the lowest Co and highest Ni contents ($S_{0.97-0.98}Fe_{0.06-0.07}Co_{0.88-0.89}Ni_{0.04}As_{1.03}$; Fig.

291   7). Cobalt and Ni contents in cobaltite in the Idaho Zone, located in the same metamorphic domain as the

292   Chicago Zone, has slightly higher Co and lower Ni contents ($S_{0.99}Fe_{0.05-0.06}Co_{0.92-0.93}Ni_{0.02}As_{1.01}$). Cobaltite

293   in the lowermost Haynes-Stellite structural block with lower to middle greenschist facies grade, has

294   mineral composition in the range of that found in the Idaho Zone ($S_{0.98}Fe_{0.05}Co_{0.92}Ni_{0.02}As_{1.02}$). Cobalt

295   EMP mapping reveals that cobaltite in the Haynes-Stellite mineral fraction is oscillatory zoned with

296   respect to Co (Fig. 6f). Outside the Blackbird domain, towards the southwest, cobaltite at Black Pine is

297   devoid of Ni and has the highest Co contents encountered in the current work

298   ($S_{0.98}Fe_{0.05-0.06}Co_{0.94-0.95}Ni_{0.00}As_{1.02}$).

299

300   **5.3   Re-Os results for cobaltite mineral fractions**

301   Results for 4 trial analyses of cobaltite mineral separates from Black Pine did not yield reliable Re-Os

302   isotope data because of low Re concentrations (< 0.3 ppb; Table 3). By contrast, the results of 23 Re-Os

303   isotope analyses of cobaltite mineral separates from the Chicago Zone (n = 7), Idaho Zone (n = 10) and

304   Haynes-Stellite (n = 6) are presented in full in Table 3. Osmium concentrations range from 12 to 64 ppt

305   whereas Re concentrations range from 0.4 to 3.4 ppb. Uncertainties (±2σ) are 0.89% or better for Re

306   abundances whereas they amount to 9.7% or better for Os, except seven outliers with uncertainties above

307   12%. These outliers are readily explained by higher percentages of Re and Os blanks, with values above

308   2.8% and 0.75%, respectively. Ratios of $^{187}Re/^{188}Os$ vary from 512 to 1818. Ratios of $^{187}Os/^{188}Os$ range

309   from 17.4 to 44.2. These ratios are highly radiogenic and the mineral fractions of these deposits qualify as

310   Low-Level Highly Radiogenic (LLHR) samples (cf. Stein et al., 2000). There is no trend of magnetic

311   fractions having more radiogenic $^{187}Os/^{188}Os$ ratios than the respective non-magnetic fractions (e.g.,

312   NM0.9 vs. M0.6 & M0.9 for the Chicago Zone) or "less" magnetic fractions (e.g., M0.6 vs. M0.3 in the

313   Idaho Zone).

314   Data points for the Chicago Zone scatter greatly, not defining any cluster nor linear trend in a

315    $^{187}Re/^{188}Os$ vs. $^{187}Os/^{188}Os$ plot (Fig. 8). By contrast, data points for the Idaho Zone and the Haynes-

316   Stellite deposit plot in particular fields of this diagram. Most data points for the Idaho Zone plot along or

317   close to a 1100 Ma reference isochron (initial $^{187}Os/^{188}Os$ ratio of 9) with two outliers closer to the 1200

318   Ma reference isochron (same initial $^{187}Os/^{188}Os$ ratio). By contrast, data points for the Haynes-Stellite

319   deposit plot close to a 1300 Ma reference isochron (initial $^{187}Os/^{188}Os$ ratio of 5) with two outliers plotting

320   close to a 1500 Ma reference isochron (same initial $^{187}Os/^{188}Os$ ratio). The ten data points for the Idaho

321   Zone define a Model 3 isochron with an age of 1105 ± 240 Ma (2σ; mean squared weighted deviates,

322   MSWD = 7.5, initial $^{187}Os/^{188}Os$ ratio [Os]i = 9.5 ± 2.9; Fig. 9). However, three data points display

323   extremely large error ellipses. These uncertainties are explained by high Re blank contribution for these

324   analyses (Table 3). Regression of the seven other data points yields a Model 3 isochron with an age of

325   1132 ± 240 Ma (2σ; MSWD = 9.3, initial $^{187}Os/^{188}Os$ ratio [Os]i = 9.0 ± 2.9). The four data points of

326   Haynes-Stellite cobaltite that plot close to the 1300 Ma reference isochron all have aliquot weights above

327   200 mg. These data points define a Model 1 isochron with an age of 1349 ± 76 Ma (2σ; MSWD = 2.1,

328   initial $^{187}Os/^{188}Os$ ratio [Os]i = 4.7 ± 2.2; Fig. 10). The two outliers mentioned above have aliquot weights

329   of 50 and 145 mg and present large uncertainties and extended error ellipses due to high Re blank

330   contribution (even high Os contribution in the case of the aliquot with weight of 50 mg; Table 3).

331

332   # 6.   Discussion

333   ## 6.1   Cobaltite petrography, chemistry, metamorphism and validation of Re-Os isochron

334   ## ages

335   In the Chicago Zone, the alternating bands of close-packed fine-grained and coarse-grained massive

336   cobaltite (Figs. 4b, 6) were interpreted as a product of recrystallization of pre-existing cobaltite

337   mineralization in response to Cretaceous metamorphism (Bookstrom et al., 2016), producing cross-aligned

338   cobaltite fabrics shown in Fig. 4b. Our results support this interpretation. This cobaltite metamorphosed to

339   middle to upper greenschist facies has the lowest Co and the highest Ni content of the study (Fig. 3, Table

340   2). We suggest that Co was lost from cobaltite and retained in the rock in other minerals during prograde

341   metamorphism. This mechanism would explain the relative Co-enrichment of biotite hornfels in the

342   Blackbird area (i.e., 19 ppm Co; Connor, 1991; Bookstrom et al., 2016), thereby implying that these rocks

343   are not the source of metals but became rather enriched in metals as a result of metal redistribution during

344   metamorphism. This preferred interpretation is supported by a similar pattern observed in pyrite and other

345   sulfide minerals in unmetamorphosed to amphibolites facies-metamorphosed greywackes in the Otago and

346   Alpine Schists of New Zealand (Pitcairn et al., 2010). These authors observed that prograde

347   metamorphism and breakdown of pyrite to pyrrhotite was accompanied by a redistribution of Co and Ni

348   that remained in the rock and were incorporated in other sulfide minerals than pyrite. Whereas Co was

349   probably lost to biotite hornfels during prograde metamorphism in the Blackbird area, Ni was enriched in

350   cobaltite in the Chicago Zone. Although it would be tempting to link this Ni-enrichment to the presence of

351   pyrrhotite that was observed in other samples of the Chicago Zone (cf. Bookstrom et al., 2016), no

352   pyrrhotite inclusions were observed in cobaltite in the studied sample. The significant scatter of the Re-Os

353   data for the Chicago Zone in the $^{187}Re/^{188}Os$-$^{187}Os/^{188}Os$ space would logically be expected for cobaltite

354   that recrystallized during metamorphism and underwent disturbance of the Re-Os chronometer.

355   Disturbance of the chronometer during metamorphism would be expected to result in mixed signatures of

356   the metamorphic fluid and the primary Re-Os budget in cobaltite. A diagram of $^{187}Os/^{188}Os$ vs. $1/^{192}Os$

357   using "common" ^{192}Os rather than total Os (including radiogenic ingrowth of ^{187}Os) enables a fair

358   assessment of mixing (van Acken et al., 2014). Some degree of mixing may be observed in the pseudo-

359   linear trend of the Chicago Zone data points in such a diagram (labeled "CH-x" in Fig. 11, green ellipse).

360   The absence of isochronous behavior of the data and evidence of mixing with metamorphic fluid suggest

361   significant disturbance of the system that cannot even be exploited to determine the timing of this

362   disturbance (Figs. 8 and 11).

363   Fine-grained cobaltite from the Haynes-Stellite deposit is oscillatory-zoned with respect to Co with

364   Co contents varying over 5wt. % (Table 2, Figs. 6f and 7). Within the lower to middle greenschist

365   Haynes-Stellite structural block (Fig. 3), cobaltite shows limited recrystallization as illustrated by

366   coexisting very fine-grained grains (<10 µm) in between a few larger cobaltite crystals (ca. 20 to 50 µm;

367   Fig. 6). An isochronous behavior is only observed for 4 Re-Os data points of fine-grained cobaltite that

368   correspond to aliquots with weight above 200 mg (Fig. 8). The other two data points corresponding to

369   aliquots of ca. 50 and 150 mg deviate towards older ages in the range of 1500 Ma (Fig. 8, Table 3). We

370   suggest that sampling cobaltite from Haynes-Stellite to a minimum aliquot size of 200 mg overcomes the

371   possible, yet minor, disturbance of the Re-Os budget in cobaltite by this limited recrystallization during

372   lower to middle greenschist metamorphism. Indeed, the isochronous behavior of the 4 data points for

373   aliquots of at least 200 mg combined with the absence of linear correlation in $^{187}Os/^{188}Os$-$1/^{192}Os$ space

374   (Figs. 8 and 11) indicate that Re-Os systematics were preserved in cobaltite at Haynes-Stellite, and, the

375   late Mesoproterozoic (ca. 1349 Ma) age may be further interpreted in the next section.

376   To end with, fine-grained cobaltite from the Idaho Zone has an extremely uniform chemical

377   composition whatever the mineral fraction (i.e., M 0.3 and M 0.6; Fig. 7 and Table 2). In both fractions, in

378   particular the M 0.6 fraction, cobaltite shows evidence of recrystallization of formerly very fine-grained

379   cobaltite grains (<10 µm) that coalesced together to produce coarser-grained cobaltite (ca. 20 µm)

380   embedded in silicates ± xenotime (Fig. 6h). Given the position of the Idaho Zone within the same middle

381   to upper greenschist metamorphic facies structural block as the Chicago Zone (Fig. 3), it is likely that the

382   Re-Os budget of cobaltite was either reset or completely disturbed without any age information. Yet, the

383   isochronous behaviour of the seven data points of the Idaho Zone with low-uncertainty ellipses and the

384   lack of linear correlation in the $^{187}Os/^{188}Os$-$1/^{192}Os$ space (Figs. 7 and 11) imply that the 1132 ± 240 Ma

385   age is meaningful and is interpreted geologically in the next section.

386

387   **6.2   From Middle Mesoproterozoic cobaltite mineralization to Grenvillian and Cordilleran**

388   **metamorphic overprint**

389   A summary of events from the formation of the Belt Basin in Early Mesoproterozoic to the

390   compressive tectonics of the Cordilleran fold-and-thrust belt is proposed in Fig. 12. Although the cobaltite

391   Re-Os age of 1349 ± 76 Ma for the Haynes-Stellite deposit is associated with a relatively large uncertainty

392   (~ ± 5.6%, 2σ), due primarily to the limited range of $^{187}Re/^{188}Os$ and $^{187}Os/^{188}Os$ ratios (Fig. 8, Table 3), it

393   is consistent with existing constraints for a middle Mesoproterozoic sulfide mineralization in the Idaho

394   Cobalt Belt. Overall, this age agrees with a tectonic and plutonic engine for mineralization related to the

395   East Kootenay orogeny (ca. 1379 to 1325 Ma). Our Re-Os age is younger than the youngest SHRIMP U-

396   Pb detrital zircon age (1409 ± 10 Ma) constraining the maximum age of the host sedimentary unit at

397   Haynes-Stellite (Gunsight Formation, Fig. 2; Aleinikoff et al., 2012; Bookstrom et al., 2016). In addition,

398   this age draws a connection with the emplacement of bimodal gabbro-granite suite of plutons between ca.

399   1383 and 1359 Ma (Anderson and Davis, 1995; Evans et al., 2000; Aleinkoff et al., 2012). In particular, a

400   timing for cobaltite mineralization at ca. 1349 Ma supports the proposal of Slack (2012) that sulfide

401   mineralization may have occurred in connection with the emplacement of the Big Deer Creek granite

402   dated at 1377 ± 4 Ma (Aleinikoff et al., 2012). This granite is presently found in the Haynes-Stellite

403   structural block (Fig. 3) and originally intruded into the host rocks to the Haynes-Stellite deposit.

404   Aleinikoff et al. (2012) and Slack (2012) convincingly connected the precipitation of xenotime in the

405   Apple Creek Formation with the emplacement of the Bid Deer Creek granite. Of particular interest is the

406   oldest generation of xenotime dated at 1370 ± 4 Ma that locally forms inclusions in cobaltite (e.g., Merle

407   deposit, Fig. 3; Aleinikoff et al., 2012), suggesting that cobaltite is younger than ca. 1370 Ma at the Merle

408   deposit. Our ca. 1349 Ma Re-Os age, although not produced from cobaltite with xenotime inclusions dated

409   at ca. 1370 Ma as at Merle, indicates timing of Co-sulfide mineralization shortly after a phase of REE-Y

410   mineralization in connection with the emplacement of the Big Deer Creek granite at ca. 1370 Ma, as

411   originally proposed by Slack (2012) in his effort to explain the close textural association of cobaltite and

412   xenotime in many samples from the district. It is also agrees with the bracketing of formation of the

413   quartz-tourmaline-cobaltite breccia between 1370 and 1320 Ma proposed by Bookstrom et al. (2016) in

414   their synthesis of the geologic history of the Blackbird mine area based on observed relative age

415   relationships and available isotope age determinations on gangue minerals. In the light of this independent

416   data and our Re-Os age of ca. 1349 Ma, we propose that the Haynes-Stellite cobaltite mineralization was

417   formed in middle Mesoproterozoic time. The Re-Os isotopic system closed in this mineral at that time

418   without being disturbed or reset by Grenvillian metamorphism (ca. 1190 to 1006 Ma) nor by tectonism,

419   metamorphism, and plutonism associated with Cordilleran orogenesis in Early to Late Cretaceous time

420   (ca. 151 Ma to 83 Ma; Bookstrom et al., 2016 and references therein) in this sample.

421   The imprecise age for mineralization in the Idaho Zone (1132 ± 240 Ma, 21%, 2σ) overlaps within

422   uncertainty with the age of cobaltite mineralization at Haynes-Stellite. However, three arguments favor the

423   interpretation that this age records a younger event for the Idaho Zone that is distinct from that of the

424   Haynes-Stellite deposit: (1) cobaltite in the Idaho Zone has at least 3 times less Re than cobaltite at

425   Haynes-Stellite (Table 3), (2) the data points for each of these deposits define separate clusters in the

426   $^{187}Re/^{188}Os$-$^{187}Os/^{188}Os$ space (Fig. 8), and, (3) the initial $^{187}Os/^{188}Os$ ratios for cobaltite in these deposits

427   are distinct within error (Figs. 9 and 10). In addition, the strongest evidence for a younger event in the

428   Idaho Zone than at Haynes-Stellite is that of Bookstrom et al. (2016) who proposed that unlike the quartz-

429   tourmaline-colbatite breccia at Haynes-Stellite, the quartz-tourmaline-breccia of the Idaho Zone parallels

430   axial planar S_2 cleavage of F_2 folds, and, is coeval with a phase of cobaltite-xenotime precipitation

431   bracketed at 1058 to 990 Ma. Our Re-Os age, albeit imprecise, overlaps with the timing proposed by

432   Bookstrom et al. (2016) and corresponds to the time span of the Grenvillian orogeny. Although the

433   Grenvillian orogen in present-day Sweden and eastern Canada is not known to be productive in terms of

434   mineral deposits (Stephens M.B., pers. comm., 2015), evidence for regional Grenville-age metamorphism

435   in western Laurentia continues to accumulate (cf. Zirakparvar et al., 2010; Vervoort et al., 2015).

436   Arsenopyrite from the main ore stage silver mineralization in the Galena Mine in the neighboring Coeur

437   d'Alene district, Idaho yielded a Re-Os age of ca. 1220 Ma, i.e., a Grenvillian age (Arkadakskiy et al.,

438   2009). Other evidence of hydrothermal mineral precipitation of Grenvillian age was brought by Aleinikoff

439   et al. (2015) with xenotime precipitation at sub-greenschist temperatures in the upper section of the Belt

440   Supergroup at about 1160-1050 Ma. Given the relationship between the quartz-tourmaline-breccia in the

441   Idaho Zone and axial planar cleavage and fold, our tentative interpretation would suggest that a proto-ore

442   which formed in the middle Mesoproterozoic (as in the case of the Haynes-Stellite quartz-tourmaline-

443   cobaltite breccias) was affected by dynamothermal events (above 400°C as implied by the presence of

444   garnet) during the Grenvillian orogeny. The Re-Os isotopic system in cobaltite in the Idaho Zone was

445   reset above 400°C to yield a Grenvillian, yet, highly imprecise age. The uncertainty is primarily due to the

446   relatively low Re and radiogenic ^{187}Os contents. If the Idaho Zone originally formed at the same time as

447   Haynes-Stellite, then Re (and Os) would have been lost during Grenvillian resetting thereby explaining the

448   current Re and Os contents in cobaltite in the Idaho Zone with a much younger late Mesoproterozoic age.

449   To end with, the Grenvillian age recorded by the Re-Os isotopic system in the Idaho Zone was preserved

450   despite incorporation of this deposit in the Blackbird domain (Fig. 3) where metamorphic grade reached

451   middle to upper greenschist facies in Cretaceous time.

452   The Chicago Zone failed to be dated by the isochron method and mixing between the Re-Os budget

453   of cobaltite and later hydrothermal fluid is suggested. Yet, the highly radiogenic ^{187}Re/^{188}Os and

454   ^{187}Os/^{188}Os ratios in the range of those encountered at Haynes-Stellite and in the Idaho Zone favor a

455   middle Mesoproterozoic proto-ore in which the Re-Os isotopic system was severely disturbed post-

456   mineralization. Given the evidence of recrystallization of most cobaltite during Cretaceous metamorphism

457   (Aleinikoff et al., 2012 in Slack, 2012), it is likely that the scatter in the Re-Os data (Fig. 8) observed in

458   recrystallized fine- to coarse-grained cobaltite (Fig. 4 and Fig. 6) is the result of Cretaceous

459   dynamothermal events and compressive tectonics (i.e., shear zone hosting the Chicago Zone). According

460   to Eiseman (1988), Cretaceous metamorphism reached 400 to 520°C. Given the disturbance of the Re-Os

461   isotopic system in cobaltite in the Chicago Zone, this temperature range is therefore concluded as being

462   above the closing temperature for Re-Os in cobaltite.

463   The origin of coarse-grained cobaltite associated with cobaltian arsenopyrite in chloritized biotite

464   phyllite remains unknown. A highly tentative hypothesis would consider that the small high-grade Black

465   Pine deposit (1.0 Mt at 4.5 wt%, 0.08 wt% Co, and 1.03 g/t Au; Johnson et al., 1998) is a by-product of

466   metamorphism and re-deposition of metals originally contained in Mesoproterozoic cobaltite ores of the

467 Blackbird area. Indeed, the mineral chemistry of cobaltite at Black Pine is the richest in cobalt in the

468 current study and is devoid of Ni. Considering our interpretation for the Chicago Zone that cobalt was

469 released during metamorphism while Ni was proportionally enriched, cobalt would have been re-deposited

470 elsewhere, i.e., possibly towards the distal Black Pine site, where temperature and pressure levels were

471 lower as indicated by lower metamorphic grade. In addition to enriching biotite hornfels in cobalt, it is

472 possible that (Cretaceous?) metamorphism caused the formation of small high-grade deposit in NW-

473 striking northwest and steeply NE-dipping strata like Black Pine.

474

475 **6.3   Preferred genetic model for Mesoproterozoic cobaltite mineralization in the light of**

476 **the new Re-Os isotope geochemistry data**

477   Considering the new Re-Os ages which indicate primary cobaltite mineralization in middle

478 Mesoproterozoic during the East Kootenay orogeny with reworking during the Grenvillian orogeny and/or

479 the Cordilleran orogeny (Fig. 12), we propose to refine the genetic model for cobaltite mineralization in

480 the Idaho Cobalt Belt. By contrast to a trend in modern geology with an urge to classify ore deposits in

481 "boxes" with acronyms that get often misused, we choose to adopt a purely descriptive approach by

482 synthesizing existing literature data and new Re-Os data within the geologic framework of the engine that

483 was the East Kootenay orogeny, followed by regional metamorphism in Grenvillian time, and by

484 Cordilleran metamorphism, deformation, and plutonism in Late Jurassic to Late Cretaceous time. Any

485 attempt to classify these deposits according to deposit-types (e.g., IOCG, Iron Oxide Copper-Gold) should

486 "see through" metamorphism, and, take into account the established reaction of sulfide (e.g., pyrite)

487 destabilization and H_2S devolatization during prograde metamorphism, i.e., pyrite $\rightarrow$ pyrrhotite + H_2S $\rightarrow$

488 magnetite + H_2S.

489   As originally proposed by Slack (2012), cobaltite mineralization in the Idaho Cobalt Belt is indeed

490 an expression of multistage hydrothermal mineralization related to a magmato-hydrothermal system.

491 Although the middle Mesoproterozoic ca. 1349 Ma age for cobaltite mineralization applies to the Haynes-

492 Stellite tourmalinized breccias, mineralogical and geochemical ties between mineralization of these

493   breccias and the strata-bound sulfide deposits of the Blackbird district have been put forward by Slack

494   (2012) and include: (1) the presence of abundant xenotime with cobaltite locally in both deposit types

495   (Slack, 2012; our Fig. 6), (2) additional links in the similarity of boron isotope values ($\delta^{11}B = -6.9$ to

496   +3.2‰) determined for the breccia-hosted tourmalines and those within the strata-bound sulfide deposits

497   such as within the Idaho Zone (Trumbull et al., 2011; Slack, 2012). Hence, we propose that Co-Cu-Au

498   sulfide deposits were produced as part of an epigenetic Co-Cu-Au-Bi-Y-REE mineralizing system in

499   connection with the emplacement of a ca. 1383 to 1359 Ma suite of bimodal gabbro-granite intrusions, in

500   particular the Big Deer Creek granite dated at 1377 ± 4 Ma (Slack, 2012; Aleinikoff et al., 2012). The

501   mineralizing system would have deposited first Y, REEs and Be-bearing minerals (xenotime, gadolinite-

502   (Y)) within these deposits at 1370 ± 4 Ma (Slack, 2012; Aleinikoff et al., 2012). The second phase of

503   mineralization consisted in the precipitation of cobaltite, and possibly, by extension, deposition of the Cu-

504   sulfides. Given the fact that cobaltite mineralization occurred prior to Cretaceous metamorphism, a valid

505   genetic model must see through the present metamorphosed rocks and metamorphic minerals to

506   understand what were the conditions at the time of mineralization by documenting the origin and

507   composition of the three key ingredients for mineralization: (1) hydrothermal fluids, (2) source of sulfur,

508   and (3) source of metals.

509   Landis and Hofstra (2012) proposed that the hydrothermal fluid trapped in gangue quartz associated

510   with cobaltite mineralization was a mixture of evaporated seawater and subordinate magmatic fluid.

511   Trumbull et al. (2011) proposed that hydrothermal fluids at a formational temperature of 300°C and an

512   boron isotopic of -7 to 3 ‰ originated from marine carbonates and/or evaporitic marine borates;

513   chemogenic sediments that are currently found in form of scapolite-tourmaline beds in the Yellowjacket

514   and Apple Creek Formations (Fig. 1A; Tysdal and Desborough, 1997; Tysdal et al., 2003). The presence

515   of evaporitic marine borates is compatible with the equatorial paleolatitudes that existed in the

516   Mesoproterozoic (Landis and Hofstra, 2012). We suggest that intrusion of the Big Deer Creek granite at

517   ca. 1377 Ma put these evaporitic fluids in motion and these mixed with magmatic-related fluids as

518 indicated by mantle-derived ^{3}He and fluorine-derived ^{22}Ne compositions of volatiles trapped in the same

519 gangue quartz (Landis and Hofstra, 2012).

520   The initial ^{187}Os/^{188}Os ratio determined through isochron regression reflects the origin of bulk Os

521 incorporated in a mineral at the time of precipitation prior to ingrowth of radiogenic ^{187}Os below the

522 blocking temperature of the Re-Os system in the given mineral (e.g., Walker et al., 1991). By extension, it

523 may be used to trace the source of metals and cobalt in the present case study. The extremely high initial

524 ^{187}Os/^{188}Os ratio for the Haynes-Stellite deposit (4.7 ± 2.2) points to Archean and Paleoproterozoic source

525 rocks from which Os (and metals) would have been derived. The Idaho Cobalt Belt in the Lehmi sub-

526 basin is underlain by a NE−SW striking corridor of Laurentian basement rocks of the Great Falls tectonic

527 zone which is sandwiched between the 3.2−2.8 Ga Medicine Hat block in the north and the >2.5 Ga

528 Wyoming Craton in the south (Fig. 1B; Vervoort et al., 2015; Wang, 2015; Kilian et al., 2016). The Great

529 Falls tectonic zone comprises Paleoproterozoic oceanic island-arc terrane which is in turn underlain by a

530 very thick mafic body interpreted from seismic data (Foster et al., 2006). These Paleoproterozoic ±

531 Archean rocks with overall mafic affinity appear as sound candidates able to supply metals like Co for

532 mineralization. Arsenopyrite in the Galena mine has similarly high initial ^{187}Os/^{188}Os ratio of ca. 6.

533 Although arsenopyrite accompanying silver mineralization in the Galena mine is younger (ca. 1220 Ma;

534 Arkadakskiy et al., 2009) than cobaltite mineralization at Haynes-Stellite, the similar range of initial ratio

535 suggest derivation of metals from also these Paleoproterozoic to Archean rocks for mineralization in the

536 Coeur d'Alene district, or else possibly, from older (ca. 1400 to 1500 Ma) ore deposits equivalent of the

537 Sullivan deposit of British Columbia (Fleck et al., 2002) in which metals might have previously been

538 mobilized from these Paleoproterozoic to Archean rocks.

539   Our interpretation is compatible with the uniform sulfur isotope values of cobaltite throughout the

540 Blackbird district (8.0 ± 0.4‰) that was interpreted as sulfur being sourced from deep, crustal, highly

541 metamorphosed ± magmatic rocks (Johnson et al., 2012). Sulfur isotopic homogenization would have

542 occurred during transport of possibly aqueous H_2S in the mixed hydrothermal fluids described above

543 (Johnson et al., 2012; this study). Thus, considering that this mixture of magmatic and evaporated-

544   seawater-derived hydrothermal fluids transported both metals and reduced sulfur, the only viable

545   precipitation mechanism for sulfide mineralization is cooling, as originally proposed by Johnson et al.

546   (2012). The biotitite zones (up to 1.87 wt% Cl in biotite) spatially and paragenetically associated with

547   sulfide mineralization would thus represent metamorphosed alteration zones in which Fe- and Cl-rich

548   hydrothermal fluids reacted with aluminous clastic sediments to form abundant Cl-rich Fe biotite (Slack,

549   2012). Both hydrothermal alteration envelopes and sulfides were deformed during Cretaceous deformation

550   that produced biotitites.

551   Following primary cobaltite mineralization in middle Mesoproterozoic at ca. 1349 Ma, cobaltite

552   mineralization was locally remobilized along S_2 cleavages during the Grenvillian orogeny at ca. 1132 $\pm$

553   240 Ma, as in the case of the quartz-tourmaline-cobaltite breccia in the Idaho Zone (Fig. 12).

554   Remobilization of pre-existing middle Mesoproterozoic cobaltite mineralization with high initial

555    $^{187}Os/^{188}Os$ ratio of ca. 5 is a preferred interpretation in order to explain cobaltite mineralization in the

556   Idaho Zone with an initial ratio at ca. 9 and much lower Re contents than in older colbatite at Haynes-

557   Stellite (Fig. 9).

558   The final stage of cobaltite mineralization consists of a Cretaceous overprint (Fig. 12), recorded in

559   the massive cobaltite from the Chicago Zone. Although most cobaltite in the Blackbird district was

560   affected by lower to upper greenschist metamorphism, recrystallization and local remobilization

561   (Bookstrom et al., 2007), Re-Os isotope systematics were preserved in those cobaltite deposits that were

562   not affected by temperatures above 400°C. Cretaceous overprint would have reset the Pb isotopic

563   composition of sulfides in the Blackbird district at ca. 100 Ma (data in Panneerselvam et al., 2012). The

564   Pb isotopic composition of chalcopyrite at the Black Pine prospect (Panneerselvam et al., 2012) and our

565   interpretation of the Re-Os budget and mineral chemistry in this small high-grade deposit favor the

566   original idea of Bookstrom et al. (2007, 2016) that satellite deposits like Black Pine and Iron Creek (Fig.

567   1A) represent Cretaceous remobilization in structures of pre-existing cobaltite mineralization. Most of

568   these prospects, besides Black Pine where cobaltite is identified, do not contain cobaltite but cobaltian

569   arsenopyrite and cobaltian pyrite in addition to pyrite and chalcopyrite. This interpretation is in line with

570　the presence of late quartz-pyrite-chalcopyrite veins in the Sunshine deposit with muscovite selvage that

571　was dated as having formed at ca. 83 Ma.

572

5　Conclusions

574　　　In the present study, we have constrained for the first time the Re-Os systematics in cobaltite and

575　produced Re-Os ages for cobaltite mineralization in two of the four studied deposits hosted by

576　metasedimentary rocks of the Idaho Cobalt Belt (ICB), Idaho, USA (i.e., Haynes-Stellite deposit, Chicago

577　and Idaho Zones of the Blackbird mine, Black Pine prospect). The results of our work are fourfold:

578　　1.　Variably magnetic fractions of cobaltite in the ICB have low Re and total Os contents (ca. 0.4−4

579　　　　ppb and 14−64 ppt, respectively) but elevated $^{187}Re/^{188}Os$ (600−1800) and highly radiogenic

580　　　　$^{187}Os/^{188}Os$ (17−45) ratios. The Black Pine prospect has too low Re concentrations (<0.3 ppb) to

581　　　　produce any reliable Re-Os data. By contrast, Re-Os data of cobaltite from the quartz-tourmaline-

582　　　　cobaltite breccia Haynes-Stellite deposit are accurate and reproducible providing a minimum

583　　　　aliquot size of 200 mg whereas data for cobaltite from the quartz-tourmaline-cobaltite breccia of

584　　　　the Idaho Zone require a minimum aliquot size of 150 mg.

585　　2.　The Re-Os systematics in cobaltite at Haynes-Stellite and at the Idaho Zone are preserved and Re-

586　　　　Os data are regressed using the isochron approach. By contrast, the Re-Os systematics of cobaltite

587　　　　in the Chicago Zone are disturbed and a linear array of mixing with post-depositional

588　　　　hydrothermal fluids is identified in the $^{187}Os/^{188}Os$ vs. $1/^{192}Os$ space. Cobaltite was primarily

589　　　　deposited in the ICB at Haynes Stellite at 1349 ± 76 Ma and was preceded by REE-Y-mineral

590　　　　precipitation at ca. 1370 Ma in connection with the emplacement of the Big Deer Creek granite at

591　　　　ca. 1377 Ma as part of a suite of bimodal magmatic activity between ca. 1383 and 1359 Ma

592　　　　(Aleinikoff et al., 2012). We propose that our age of 1132 ± 240 Ma for cobaltite in the quartz-

593　　　　tourmaline breccia in the Idaho Zone and consideration of other geologic and structural

594   constraints (Bookstrom et al., 2007, 2016) illustrate a remobilization phase of pre-existing

595   cobaltite mineralization along cleavages and folds during the 1190−1006 Ma Grenvillian orogeny.

596   3.   Cobaltite formed at ca. 1349 Ma upon cooling at an inferred temperature of ≤ 300ºC of a mixture

597   of Mesoproterozoic evaporitic brines and magmatic fluids that carried both metals and reduced

598   sulfur. The extremely high initial Os ratio derived from regression of the Re-Os data at Haynes-

599   Stellite (4.7 ± 2.2) is used a tracing tool to identify the source of metals. The most likely scenario

600   considers that metals were derived from lower crustal mafic source rocks of probable Archean

601   age, overlain by oceanic island-arc rocks of Paleoproterozoic age, which are tectonically

602   interlayered with Archean rocks of the Wyoming Province continental nucleus, in the collisional

603   Great Falls tectonic zone, which projects beneath the Idaho Cobalt Belt. This scenario is

604   compatible with the sulfur isotopic composition of cobaltite (+8.0 ± 0.4‰) that is best explained

605   by reduced sulfur being sourced from deep, crustal and highly metamorphosed rocks (Johnson et

606   al., 2012).

607   4.   We propose that the Re-Os isotope system in cobaltite in the Chicago Zone, located in a middle

608   greenschist facies metamorphic zone, may have been subjected to temperatures of 400 to 520ºC

609   proposed for Cretaceous garnet-bearing rocks of a lower amphibolites facies metamorphic domain

610   above the domain hosting the Chicago Zone. A maximum closing temperature of 400ºC is

611   inferred for Re-Os in cobaltite. By contrast, cobaltite at Haynes-Stellite and the Idaho Zone, which

612   underwent middle to lower greenschist Cretaceous metamorphism, preserved Mesoproterozoic

613   age information delivered by the Re-Os isotope system.

614

6 Acknowledgments

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804

805   **Figure captions**

806   **Fig. 1** Geologic and tectonic maps (modified from Bookstrom and others, 2016), showing the Idaho cobalt belt

807   (ICB), its geologic surroundings in the Salmon River Mountains, and its regional setting relative toselected tectonic

808   features. A. Geologic map of the Salmon River Mountains (modified from Evans and Green, 2003) showing the

809   geologic setting of selected Co-Cu-bearing mines, ore zones, and prospects of the ICB, after Johnson and others

810   (1998). Traces of the iron-oxide zone (OZ) and biotitic Blackbird zone (BZ) after Nash and Connor (1993). Samples

811   of ore minerals from the Blackbird Chicago (CH) and Idaho (ID) ore zones, the Haynes-Stellite mine (HS), and the

812   Black Pine prospect (BP) were collected for Re-Os analysis. Names of selected plutons and faults are spelled out, but

813   names of the following faults are abbreviated as Big Deer fault (BDf), Little Deer fault (LDf), Salmon Canyon fault

814   (SCf), and White Ledge fault (WLf). B. Tectonic index map (inset) showing the Idaho cobalt belt (ICB) in relation to

815   the Mesoproterozoic Belt-Purcell basin (Y) and the Lemhi sub-basin (after O'Neill et al., 2007, Box et al., 2012, and

816   Burmester et al., 2013). Basement domains that surround or project beneath different parts of the Belt-Purcell basin

817   include the Archean Wyoming Province (A), Medicine Hat block (A), and Paleopoterozoic-Neoarchean Clearwater

818   block (Clwtr blk, XA) after Vervoort et al. (2015), Wang (2015) and Kilian et al. (2016); Paleoproterozoic Selway

819   terrane (X) after Foster et al. (2006). Major faults of the Great Falls tectonic zone [GFTZ (X) and thrust belt (XA]

820   are from O'Neill (1993) and Sims et al. (2004). The Perry line (PL) is a Belt-basin-growth fault (after Winston,

821   1986). The eastern margin of the East Kootenay Orogen (Y) is drawn to include sites where igneous or metamorphic

822   minerals yield isotopic age determinations between about 1370 and 1320 Ma. Other features shown include the Great

823   Divide megashear (GDM, after O'Neill et al., 2007), the rifted continental margin (RCM, after O'Neill et al., 2007),

824   and accreted oceanic terranes (PzMz = Paleozoic to Mesozoic). Much of the western Belt-Purcell Basin is

825   overprinted by the metamorphic-plutonic hinterland of the Cordilleran orogen, and its eastern part is overprinted by

826   the Cordilleran fold-and-thrust belt.

827   **Fig. 2** Simplified columnar stratigraphic section for Mesoproterozoic strata in the Blackbird district (modified after

828   Bookstrom et al., 2016). The stratigraphic positions of the cobaltite mineralized bodies sampled for Re-Os work are

829   shown as red ellipses. Blue ellipses correspond to cobaltite mineralized bodies in which U-Pb SHRIMP ages were

830   produced for xenotime and/or monazite texturally associated with cobaltite (cf. Aleinikoff et al., 2012).

831   **Fig. 3** Highly idealized block diagram representing thrust plates and metamorphic domains in and around the

832   Blackbird district in the central part of the Idaho Cobalt Belt.Metamorphic grade is illustrated by colored infill.

833   Approximate locations of the three cobaltite mineralized bodies (CH: Chicago Zone, ID: Idaho Zone, HS: Haynes-

834   Stellite) sampled for Re-Os geochronology are shown.

835   **Fig. 4** Macrophotographs of the samples processed for Re-Os analyses: a. Haynes-Stellite (HS), b. Chicaco Zone

836   (CH), c. Idaho Zone (ID), d. Black Pine (BP).

837   **Fig. 5** Complete workflow applied to the samples presented in Figure 4.

838   **Fig. 6** Back-scattered images using a scanning electron microscope of the selected mineral fractions liberated from

839   each sample following the workflow presented in Figure 5 and mounted in epoxy. Figure 6f is a cobalt distribution

840   map produced by scanning using an electron microprobe. Abbreviations: CH: Chicago Zone, HS: Haynes-Stellite,

841   ID: Idaho Zone, BP: Black Pine.

842   **Fig. 7** Graphical representation of the variability of cobalt (Co), iron (Fe), and nickel (Ni) contents determined by

843   electron microprobe analysis in the various mineral fractions of the samples considered for Re-Os analyses.

844   Abbreviations: CH: Chicago Zone, HS: Haynes-Stellite, ID: Idaho Zone, BP: Black Pine.

845   **Fig. 8** Conventional Re-Os isochron plot of the analyses of the various magnetic fractions from the Haynes-Stellite

846   deposit (HS) and the Chicago and Idaho Zones (CH and ID, respectively) in the Blackbird mine. Reference isochrons

847   with given initial $^{187}Os/^{188}Os$ ratio are shown for reference. Abbreviations: M x.x : Magnetic fraction at current x.x

848   amps ; NM : x.x : non-magnetic fraction at current x.x amps.

849   **Fig. 9** Diagram of isochron age determination (case 1 with n = 7, case 2 with n = 10) for the Idaho Zone, Blackbird

850   mine, in the $^{187}Os/^{188}Os$ vs. $^{187}Re/^{188}Os$ space. The empty ellipses represent the 3 data points with high uncertainty.

851   **Fig. 10** Diagram of isochron age determination (n = 4) for the Haynes-Stellite deposit in the $^{187}Os/^{188}Os$ vs.

852     $^{187}Re/^{188}Os$ space. The empty ellipses represent the 2 data points with high uncertainty with aliquot size below the

853   minimum aliquot size of 200 mg.

854   **Fig. 11** $^{187}Os/^{188}Os$ vs. $1/^{192}Os$ plot to identify potential mixing relationships.

855   **Fig. 12** Timeline of events for the Belt-Purcell and Lehmi sub-basin. Data after 1: Bookstrom et al., 2016; 2: this

856   study; 3: Lund et al., 2011; 4: Aleinikoff et al., 2012; 5: Jiang et al, 2001; 6: Aleinikoff et al., 2015; 7: Evans et al.,

857   2000; 8: Gasching et al., 2010; 9: Anderson and Davies, 1995; 10: Mulder et al., 2016. Abbreviations: BDCG: Big

858   Deer Creek granite, Fm: Formation.

859

Table labels

861

Table 1 Blank data for Re-Os analysis. Series « a » and « b » indicate what blank values is appended for each

analysis in Table 2.

Table 2 Statistical summary of electron microprobe analyses and model mineral formula of each variably magnetic

cobaltite fraction in the Chicago Zone and the Idaho Zone in the Blackbird mine, the Haynes-Stellite deposit, and the

Black Pine prospect. ICB : Idaho Cobalt Belt. Abbrevations: M x.x : Magnetic fraction at current x.x amps ; NM :

x.x : non-magnetic fraction at current x.x amps.

Table 3 Re-Os analytical results for the Chicago Zone and the Idaho Zone in the Blackbird mine, the Haynes-Stellite

deposit, and the Black Pine prospect. Due to low Re levels and high uncertainties in the analyzed Os levels in

cobaltite at the Black Pine prospect, only these values are reported because other items in the Re-Os data set have no

meaning. M x.x: Magnetic fraction at current x.x amps ; NM : x.x : non-magnetic fraction at current x.x amps, side

slope of 15º and forward slope of 10º. Model ages were calculated for aliquots of the Chicago Zone and the Idaho

Zone using an initial $^{187}Os/^{188}Os$ ratio of 9 ± 3 whereas model ages were calculated for aliquots of the Haynes-Stellite

deposit using an initial $^{187}Os/^{188}Os$ ratio of 5 ± 2.

875

Electronic Supplementary Material

877

ESM 1. Full data set of electron microprobe analyses of each variably magnetic cobaltite fraction in the Chicago

Zone and the Idaho Zone in the Blackbird mine, the Haynes-Stellite deposit, and the Black Pine prospect. Detection

limits are: 0.02 wt% for Zn, 0.01 wt% for Fe, 0.03 wt% for Ni, 0.02 wt% for Cu, 0.02 wt% for S, 0.12 wt% for Se,

0.07 wt% for Te, and 0.09 wt% for Bi.

A 114°30'W

Tci, Ylgn, Yig, Yim, Salmon River Intrusive Complex, Brushy Gulch fault, Yla, SCf

Clearwater orogenic zone

Ylgn, Yig, Yig, Yim, Yig, Ki, +Yig, Salmon River, Shoup, Yig

Salmon Canyon CuCo

SR, SCf, Ylg

Yla?, SR, middle fork, Ylgn, Yla?, Yla?, Iron Lake fault, Tci

45°15'N, Clear Cr., Panther Cr., +Yig, granite pluton of Big Deer Creek, Poison Creek fault, Ylg

Ys, Arnett Creek pluton, F4, O xi, Beartrack Au, F4, Ylg, +Yig, Creek shear zone, F4, Ys, F4, QTu, Salmon

Big Deer Cr., BDf, Yyh, Ylsq, Ylgb, 40, LDf, Little Deer Cr., Ylg, Deep Ck pluton, O i, Tcv

Meadow, CH, Blackbi, ID, HS, Ylg, 5, F4, F2, QTu

Blackbird CoCuAu, Blackbird west fork, BZ, F2, Deep Cr, Ylg

Ylac, BDf, **Black Pine CuCo**, Ylab, BZ, Ylg, QTu, Jcv

Iron Lake fault, Ylab, Iron Cr., north fork, Ys, 45°N

Ylac, OZ, QTu, **Iron Creek CuCo**, Tcv

Yellowjacket Cu-Au, Yellowjacket pluton, O i, QTu, Iron Cr., Yyh, Tcv, SOs, Tci

MAP SYMBOLS

- ⬛ Idaho cobalt belt (ICB)
- ✕ Mine
- • Prospect
- ✱ Tourmaline breccia
- ✳ Na-scapolite
- BZ Biotite-rich zone
- OZ Fe-oxide zone
- Fold axis, showing plunge and fold-set number
- High-angle fault. Arrows show lateral offset; bar and ball on down-thrown side
- Thrust fault
- Thrust (inferred)
- Shear zone
- ⬛ Town

B

BRITISH COLUMBIA, ALBERTA, CANADA, USA, 49°N

Medicine Hat Block (A), Clrwtr Block (XA), East Kootenay orogen (Y), Belt-Purcell basin (Y)

RCM, WASHINGTON, MONTANA, IDAHO, Helena sub-basin, GFTZ, Selway terrane (X), thrust belt (XA)

Accreted terranes (Pz-Mz), Salmon, PL, Wyoming Province (A)

ICB, OREGON, Lemhi sub-basin (Y), 111°W, GDM, WYOMING

100 km

EXPLANATION OF MAP UNITS

QUATERNARY-TERTIARY

- QTu Sediments ± Volcanics, Undivided
- Tci Challis intrusions
- Tcv Challis Volcanics

CRETACEOUS

- Ki Granitoid intrusions

LOWER PALEOZOIC

- SOs Sedimentary strata (Silurian-Ordovician)
- O i Syenitoid intrusions (Ordovician-Cambrian, 497 ± 6 Ma)

MESOPROTEROZOIC

- Yig Granitic intrusions (1.37 ± 0.01 Ga)
- Yim Mafic intrusions (1.37 ± 0.01 Ga)
- Ys Swauger Fm.

Lemhi Group

- Ylgn Lemhi gneiss (+Grt+Sil)
- Ylg Lemhi schist and quartzite
- Ylg Gunsight Fm. (Ylg=quartzite facies) Ylgb=biotitic facies
- Yla Apple Cr. Fm. (of Yellow Lake)
- Ylab Banded siltite unit
- Ylac Coarse siltite unit
- Yyh Yellowjacket and Hoodoo Fms.

0 5 10 KM
0 6 MI
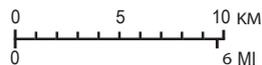
Legend

Black Pine, Idaho, Chicago, Haynes-Stellite — Cobaltite mineralization studied in the present Re-Os work

Sunshine, Merle — Co-Cu mineralization for which geochronology data are available for gangue minerals (xenotime and monazite, *Aleinikoff et al., 2012*)

Gunsight Formation

Haynes-Stellite (HS)

1409 ± 10 Ma

Sunshine
Merle
Idaho (ID), Chicago (CH)
Black Pine (BP)

Apple Creek Formation

10 km

1445 ± 12 Ma

Hoodoo Quartzite

1454 ± 9 Ma

Yellowjacket Formation

Big Deer Creek granite

N

Creek plate

⁺Yig⁺

Ylg

Poison Creek fault

Poison

Creek

Little Deer Creek fault

Ylg

Oei

Deep Creek pluton

Iron Lake fault

HS

Poison Creek fault

ID

CH

Yla

Ylac

Iron Lake fault

Yyh

Bid Deer fault

Blackbird Mtn ramp

Iron Lake fault

Ylab

Indian Creek metamorphic domain
(upper greenschist to lower amphibolite facies, domain containing rocks above garnet isograd)

Haynes-Stellite structural block and metamorphic domain
(lower to middle greenschist facies)

Blackbird metamorphic domain
(middle to upper greenschist facies)

(a) "HS" — qtz-tur; fine-grained disseminated cobaltite

(b) "CH" — coarse-grained massive cobaltite; fine-grained massive cobaltite

(c) "ID" — qtz-tur; quartz vein; fine-grained disseminated cobaltite

(d) "BP" — Quartz-chlorite; Coarse-grained cobaltite

```
┌─────────────────────────────────────┐
│      Sample cleaned with silicon    │
│        carbide grit and paper       │
└─────────────────────────────────────┘
                  │
                  ▼
┌─────────────────────────────────────┐
│   Crushing using agate mortar and pestle │
│   Sieving through disposable sieves │
└─────────────────────────────────────┘
                  │
                  ▼
┌─────────────────────────────────────┐
│       70–200 mesh size fraction     │
└─────────────────────────────────────┘
                  │
                  ▼
┌─────────────────────────────────────────────┐
│  Heavy liquid separation using Sodium Polytungstate │
│   specific gravity adjusted to 2.86–2.90    │
└─────────────────────────────────────────────┘
                  │
                  ▼
┌─────────────────────────────────────┐
│       Frantz isodynamic separation  │
│   side slope at 15°, forward slope at 10° │
└─────────────────────────────────────┘
```

| 70–200 M 0.3 | 70–200 M 0.6 | 70–200 M 0.9 | 70–200 NM 0.9 |

Isotope dilution using mixed ^{185}Re+^{190}Os spike
Carius tubes digestion using inverse Aqua Regia
(cf. Shirey and Walker, 1995; Hnatyshin et al., 2016)

Steps of Re-Os isotope geochemistry
(cf. Hnatyshin et al., 2016)

N-TIMS analysis of Re and Os
(Creaser et al., 1991; Völkening et al., 1991)

Regression of Re-Os isotope data
using Isoplot v.4.1 (Ludwig, 2011)
and the ^{187}Re decay constant of Smoliar et al. (1996)

a cobaltite · Bi-telluride · 20 μm · 25 kV · **CH-M0.6**

b cobaltite · biotite · Bi-telluride · 10 μm · 25 kV · **CH-M0.9**

c gold · cobaltite · 10 μm · 25 kV · **CH-NM0.9**

d K-fds · qtz · biotite · cobaltite · 10 μm · 25 kV · **HS-M0.9**

e xenotime · cobaltite · 10 μm · 25 kV · **HS-M0.9**

f oscillatory zoned cobaltite · 10 μm · **HS-M0.9**

g xenotime · qtz · cobaltite · 20 μm · 25 kV · **ID-M0.3**

h Na-sil. · qtz · biotite · xenotime · cobaltite · 10 μm · 25 kV · **ID-M0.6**

Figure legend:

▲ HS-x M0.9 (Haynes-Stellite)

● ID-x M0.6 (Idaho Zone, Blackbird mine)

● ID-x M0.3 (Idaho Zone, Blackbird mine)

■ CH-x NM0.9 (Chicago Zone, Blackbird mine)

■ CH-x M0.9 (Chicago Zone, Blackbird mine)

■ CH-x M0.6 (Chicago Zone, Blackbird mine)

Axis labels: $^{187}Os/^{188}Os$ (y-axis), $^{187}Re/^{188}Os$ (x-axis)

Isochron labels: 1500 Ma, Osi = 5; 1400 Ma, Osi = 5; 1300 Ma, Osi = 5; 1200 Ma, Osi = 9; 1200 Ma, Osi = 5; 1100 Ma, Osi = 5; 1100 Ma, Osi = 9; 1000 Ma, Osi = 9; 900 Ma, Osi = 9

Figure shows an isochron plot of $^{187}Os/^{188}Os$ versus $^{187}Re/^{188}Os$ with red 2σ data point error ellipses.

data point error ellipses are 2σ

Age = 1132 ± 240 Ma
Initial $^{187}Os/^{188}Os$ = 9.0 ± 2.9
MSWD = 9.3, n = 7

Age = 1105 ± 240 Ma
Initial $^{187}Os/^{188}Os$ = 9.5 ± 2.9
MSWD = 7.5, n = 10

Age = 1349 ± 76 Ma
Initial ^{187}Os/^{188}Os = 4.7 ± 2.2
MSWD = 2.1, n = 4

data point error ellipses are 2σ

| Tectonic scenario (Belt-Purcell Basin)[1,10] | *Belt basin rift* | *Belt basin sag, Lemhi sub-basin rift-sag* | *Convergence along southern margin of Laurentia* | *Continent-continent collision (?)* | *Cordilleran metamorphic and plutonic activity to fold-and-thrust belt* |

Cobaltite and copper mineralization
- FIne-grained cobaltite (Haynes-Stellite)[2] — 1349 ± 76 Ma
- FIne-grained cobaltite (Idaho Zone)[2] — 1132 ± 240 Ma
- cobaltite remobilization at Black Pine (?)
- Age of sericite around late quartz-pyrite-chalcopyrite vein (Sunshine)[3] — 83 ± 1 Ma

Hydrothermal xenotime and / or monazite growth[4]
- oldest age for xenotime (Merle prospect) — 1370 ± 4 Ma
- other xenotime growth — ca. 1300 to 1270 Ma
- Merle prospect xenotime — 1058 ± 11 Ma
- monazite — 110 ± 3 Ma
- monazite — 92 ± 5 Ma

Remarkable hydrothermal event in the Belt-Purcell Basin
- Sullivan Pb-Zn-Ag deposit[5] — ca. 1470 Ma
- hydrothermal xenotime (Belt Supergroup)[6] — 1164 ± 51 Ma
- 1041 ± 42 Ma

Metamorphism / orogenies[1]
- East Kootenay orogeny — ca. 1379 to 1325 Ma
- Grenvillian orogeny — ca. 1190 to 1006 Ma
- Onset of Cretaceous metamorphism — ca. 151 to 113 Ma

Sedimentation[4,7]
- max Yellowjacket Fm (base not exposed) — 1454 ± 9 Ma
- max age, bottom of Apple Creek Fm (host horizon) — 1445 ± 12 Ma
- max age, top of Apple Creek Fm (host horizon) — 1409 ± 10 Ma

Volcanism/ plutonism[4,7,8,9]
- Moyie sills — 1468 ± 2 Ma
- Purcell lava — 1443 ± 7 Ma
- 1383 ± 4 Ma
- BDCG — 1377 ± 4 Ma
- 1359 ± 7 Ma
- Idaho Batholith — ca. 95 to 54 Ma

Age (Ma): 1500 1480 1460 1440 1420 1400 1380 1360 1340 1200 1180 1160 1140 1120 1100 1080 1060 1040 120 100 80 60

Blank series	Blank names	Os blank (pg)	1σ	Blank ^{187}Os / ^{188}Os	1σ	Re blank (pg)	1σ
Series "a", n = 3	BLK-JDH-08, BLK-NS-15Feb16, BLK-NS-23Feb16	1.34	0.03	0.27101	0.0023	4.4	0.98
Series "b", n = 3	BLK-NS-Aug16, BLK-NSEP-1, BLK-NSEP-2	0.08	0.05	0.79623	0.0620	2.3	0.47

Mineral fraction		Zn (wt %)	Fe (wt %)	Co (wt %)	Ni (wt %)	Cu (wt %)	S (wt %)	As (wt %)	Se (wt %)	Te (wt %)	Bi (wt %)	Modal mineral formula	Metamorphic grade in the structural block	Position in the ICB
CH M 0.6 (n = 20)	Min	0.00	1.16	29.49	0.17	0.00	18.50	46.02	0.00	0.00	0.00	$S_{0.97}Fe_{0.06}Co_{0.89}Ni_{0.04}As_{1.03}$	middle to upper greenschist	N
	Max	0.03	3.21	34.06	2.55	0.00	19.32	46.92	0.00	0.09	0.05			W
	average	0.01	2.15	31.72	1.38	0.00	18.81	46.57	0.00	0.02	0.00			
	Standard deviation	0.01	0.51	1.18	0.63	0.00	0.21	0.27	0.00	0.03	0.01			
CH M 0.9 (n = 20)	Min	0.00	1.11	26.91	0.03	0.00	18.78	45.93	0.00	0.00	0.00	$S_{0.98}Fe_{0.07}Co_{0.88}Ni_{0.04}As_{1.03}$	middle to upper greenschist	
	Max	0.03	4.06	34.05	3.70	0.00	19.20	47.03	0.00	0.11	0.20			
	average	0.00	2.20	31.29	1.41	0.00	18.92	46.37	0.00	0.02	0.01			
	Standard deviation	0.01	0.81	1.84	0.96	0.00	0.12	0.30	0.00	0.03	0.04			
CH NM 0.9 (n = 22)	Min	0.00	1.33	29.52	0.47	0.00	18.72	45.35	0.00	0.00	0.00	$S_{0.99}Fe_{0.06}Co_{0.90}Ni_{0.03}As_{1.02}$	middle to upper greenschist	
	Max	0.02	2.58	33.21	2.06	0.00	19.95	46.44	0.36	0.00	0.79			
	average	0.00	1.98	31.91	1.20	0.00	19.10	45.88	0.05	0.00	0.08			
	Standard deviation	0.01	0.52	6.57	0.46	0.00	3.90	9.36	0.08	0.00	0.20			
ID M 0.3 (n = 20)	Min	0.00	1.15	28.93	0.00	0.00	18.83	44.40	0.00	0.00	0.00	$S_{0.99}Fe_{0.06}Co_{0.92}Ni_{0.02}As_{1.01}$	middle to upper greenschist	
	Max	0.01	4.39	34.83	1.65	0.00	19.53	46.67	0.00	0.11	0.03			
	average	0.00	1.96	32.65	0.59	0.00	19.14	45.61	0.00	0.03	0.00			
	Standard deviation	0.00	0.73	1.38	0.51	0.00	0.20	0.55	0.00	0.03	0.01			
ID M 0.6 (n = 20)	Min	0.00	1.01	28.85	0.00	0.00	18.69	44.82	0.00	0.00	0.00	$S_{0.99}Fe_{0.05}Co_{0.93}Ni_{0.02}As_{1.02}$	middle to upper greenschist	
	Max	0.02	3.39	34.99	2.53	0.00	19.59	46.40	0.00	0.07	0.02			
	average	0.01	2.15	31.72	1.38	0.00	18.81	46.57	0.00	0.02	0.00			
	Standard deviation	0.00	0.62	1.48	0.68	0.00	0.22	0.39	0.00	0.03	0.01			
HS M 0.9 (n = 19)	Min	0.00	0.94	29.62	0.00	0.00	18.77	46.27	0.00	0.00	0.00	$S_{0.98}Fe_{0.05}Co_{0.92}Ni_{0.02}As_{1.02}$	lower to middle greenschist	
	Max	0.01	3.19	35.08	2.28	0.00	19.65	46.87	0.00	0.12	0.00			
	average	0.00	1.75	32.90	0.78	0.00	19.11	46.61	0.00	0.03	0.00			
	Standard deviation	0.00	0.56	1.36	0.65	0.00	0.22	0.16	0.00	0.03	0.00			
BP NM 0.8 (n = 18*)	Min	0.00	1.17	30.00	0.00	0.00	18.70	45.69	0.00	0.00	0.00	$S_{0.98}Fe_{0.06}Co_{0.94}Ni_{0.00}As_{1.02}$	lower to middle greenschist	
	Max	0.03	5.57	34.87	0.77	0.01	19.43	46.89	0.00	0.09	0.03			
	average	0.01	1.92	33.63	0.09	0.00	19.10	46.31	0.00	0.02	0.00			
	Standard deviation	0.01	0.99	1.00	0.20	0.00	0.19	0.28	0.00	0.03	0.01			
BP M 2.1 (n = 20)	Min	0.00	0.91	33.28	0.00	0.00	18.69	45.57	0.00	0.00	0.00	$S_{0.98}Fe_{0.05}Co_{0.95}Ni_{0.00}As_{1.02}$	lower to middle greenschist	S
	Max	0.02	2.36	35.11	0.00	0.00	19.41	46.82	0.00	0.08	0.02			E
	average	0.01	1.61	34.19	0.00	0.00	19.04	46.22	0.00	0.02	0.00			
	Standard deviation	0.01	0.40	0.47	0.00	0.00	0.19	0.32	0.00	0.02	0.00			

Sample	Granulometry (mesh size)	Magnetic fraction (15°/10°)	Mineralogy	Sample weight (mg)	Blank series in Table 1	Re (ppb)	±2σ	Os (ppt)	±2σ	$^{187}Re/^{188}Os$	±2σ	$^{187}Os/^{188}Os$	±2σ	rho	^{192}Os (ppt)	^{187}Re (ppb)	±2σ	$^{187}Os*$ (ppt)	±2s	$\%^{187}Os*$	%Re blank	$\%^{187}Os$ blank	$\%^{188}Os$ blank	Model age[1] (Myr)	±2σ
BP-01	+70	M0.8	cobaltite ± chlorite ± quartz	109.70	a	0.295	0.004	3.5	11.3	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
BP-02	+70	NM0.8	cobaltite ± chlorite ± quartz	71.40	a	0.135	0.005	0.8	4.0	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
BP-03	70-200	M1.9	cobaltite ± chlorite ± quartz	36.00	a	<0.100	0.010	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
BP-04	70-200	M2.1	cobaltite ± chlorite ± quartz	96.30	a	0.117	0.004	0.2	1.9	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
CH-01	70-200	NM0.9	cobaltite ± xenotime ± quartz	77.40	a	1.127	0.008	18.6	9.5	1571	696	33.63	14.90	0.999	1.4	0.708	0.005	10.6	3	57	5.05	3.96	83	934	274
CH-02	70-200	M0.9	cobaltite ± xenotime ± quartz	53.90	a	1.574	0.011	16.0	7.2	1545	692	17.40	7.80	0.999	2.0	0.990	0.007	4.7	4	30	5.19	7.74	83	326	280
CH-03	70-200	M0.6	cobaltite ± xenotime ± quartz ± biotite	93.58	a	1.626	0.009	18.1	5.8	1818	532	24.74	7.25	0.999	1.8	1.022	0.006	8.3	3	46	2.89	3.57	77	517	216
CH-04	70-200	M0.6	cobaltite ± xenotime ± quartz ± biotite	351.90	a	1.500	0.008	17.3	1.5	1729	139	24.15	1.94	0.997	1.7	0.943	0.005	7.7	2	45	0.83	1.00	47	524	209
CH-05	70-200	M0.9	cobaltite ± xenotime ± quartz	427.09	a	1.336	0.007	16.1	1.3	1736	130	25.85	1.93	0.996	1.5	0.839	0.004	7.7	2	47	0.77	0.87	45	580	208
CH-06	70-200	NM0.9	cobaltite ± xenotime ± quartz	545.87	a	0.925	0.005	18.7	1.8	1304	101	34.33	3.08	0.859	1.4	0.581	0.003	10.8	2	58	0.87	0.56	41	1155	285
CH-07	70-200	M0.9	cobaltite ± xenotime ± quartz	349.85	a	1.319	0.007	25.4	1.9	817	56	17.52	1.40	0.853	3.2	0.829	0.004	7.6	4	30	0.95	0.75	33	623	444
HS-01	70-200	M0.9	cobaltite ± tourmaline ± xenotime ± biotite ± K-feldspar ± quartz	51.57	a	2.186	0.013	48.4	15.8	1273	352	37.26	10.31	0.999	3.4	1.374	0.008	29.4	5	61	3.90	2.24	76	1503	198
HS-02	70-200	M0.9	cobaltite ± tourmaline ± xenotime ± biotite ± K-feldspar ± quartz	311.90	a	3.131	0.016	63.5	2.7	1309	47	34.69	1.32	0.920	4.8	1.968	0.010	37.1	6	58	0.45	0.29	27	1346	183
HS-03	70-200	M0.9	cobaltite ± tourmaline ± xenotime ± biotite ± K-feldspar ± quartz	205.30	b	3.258	0.014	63.1	2.4	1438	38	36.82	1.39	0.688	4.5	2.048	0.009	38.2	6	60	0.34	0.07	3	1313	170
HS-04	70-200	M0.9	cobaltite ± tourmaline ± xenotime ± biotite ± K-feldspar ± quartz	266.37	b	3.398	0.014	64.0	1.1	1680	25	42.80	0.63	0.932	4.0	2.136	0.009	41.7	5	65	0.25	0.05	3	1336	142
HS-05	70-200	M0.9	cobaltite ± tourmaline ± xenotime ± biotite ± K-feldspar ± quartz	145.75	b	2.106	0.009	49.1	1.6	1114	31	33.8	1.0	0.971	3.8	1.323	0.006	28.3	5	58	0.75	0.13	5	1533	213
HS-06	70-200	M0.9	cobaltite ± tourmaline ± xenotime ± biotite ± K-feldspar ± quartz	268.79	b	3.202	0.013	60.2	1.2	1732	28	44.24	0.71	0.941	3.7	2.012	0.008	39.8	5	66	0.27	0.06	3	1345	137
ID-01	70-200	M0.3	cobaltite ± biotite ± xenotime	158.90	a	0.791	0.005	20.8	3.4	771	116	24.77	3.72	0.999	2.0	0.497	0.003	9.5	3	46	3.50	1.83	63	1216	474
ID-02	70-200	M0.6	cobaltite ± biotite ± quartz ± xenotime	238.60	a	0.650	0.004	15.7	2.3	847	113	24.9	3.3	0.999	1.5	0.497	0.002	7.2	2	46	2.84	1.62	60	1117	430
ID-03	70-200	M0.3	cobaltite ± biotite ± xenotime	100.86	b	0.589	0.004	19.3	1.3	512	34	19.22	1.28	0.989	2.3	0.370	0.003	6.7	3	34	3.87	0.55	12	1186	700
ID-04	70-200	M0.3	cobaltite ± biotite ± xenotime	332.57	b	0.729	0.003	21.4	0.6	622	16	21.45	0.69	0.759	2.3	0.458	0.002	8.4	3	40	0.95	0.15	4	1190	575
ID-05	70-200	M0.6	cobaltite ± biotite ± quartz ± xenotime	107.94	b	0.425	0.004	12.0	1.5	697	81	23.87	2.78	0.995	1.2	0.267	0.002	5.3	2	44	5.01	0.77	19	1267	519
ID-06	70-200	M0.6	cobaltite ± biotite ± quartz ± xenotime	267.67	b	0.578	0.003	15.0	0.6	749	28	23.33	0.87	0.985	1.5	0.364	0.002	6.5	2	43	1.49	0.25	7	1138	477
ID-07	70-200	M0.3	cobaltite ± biotite ± xenotime	260.97	b	1.154	0.005	26.9	0.7	929	22	26.89	0.65	0.969	2.5	0.725	0.003	13.2	3	49	0.76	0.14	4	1145	384
ID-08	70-200	M0.6	cobaltite ± biotite ± quartz ± xenotime	150.03	b	0.511	0.003	14.4	1.0	637	41	21.00	1.35	0.986	1.6	0.321	0.002	5.5	2	39	3.00	0.48	11	1121	563
ID-09	70-200	M0.3	cobaltite ± biotite ± xenotime	288.17	b	0.794	0.003	22.0	0.5	630	14	20.26	0.44	0.954	2.5	0.499	0.002	8.1	3	37	1.00	0.17	4	1063	567
ID-10	70-200	M0.3	cobaltite ± biotite ± xenotime	182.00	b	0.840	0.004	22.4	0.8	695	24	21.91	0.77	0.986	2.4	0.528	0.002	16.6	0	74	1.50	0.25	7	1105	514

Sample	Zn wt%	Fe wt%	Co wt%	Ni wt%	Cu wt%	S wt%
01-CZ-M06_001	0.00	2.36	31.44	1.53	0.00	18.87
01-CZ-M06_002	0.01	1.39	33.37	0.46	0.00	19.00
01-CZ-M06_003	0.00	1.95	31.63	1.16	0.00	18.86
01-CZ-M06_004	0.00	2.54	30.49	1.99	0.00	18.60
01-CZ-M06_005	0.00	2.07	31.76	1.45	0.00	18.67
01-CZ-M06_006	0.00	2.14	31.80	1.28	0.00	18.85
01-CZ-M06_007	0.00	2.11	31.86	1.38	0.00	19.04
01-CZ-M06_008	0.01	2.02	32.25	1.38	0.00	19.00
01-CZ-M06_009	0.00	1.16	34.06	0.17	0.00	19.32
01-CZ-M06_010	0.01	1.99	31.99	1.21	0.00	18.53
01-CZ-M06_011	0.00	2.01	32.10	1.20	0.00	18.62
01-CZ-M06_012	0.01	2.69	30.01	2.45	0.00	18.59
01-CZ-M06_013	0.00	1.38	33.10	0.61	0.00	19.02
01-CZ-M06_014	0.03	3.07	29.49	2.55	0.00	18.50
01-CZ-M06_015	0.02	2.44	31.13	1.58	0.00	18.80
01-CZ-M06_016	0.00	2.02	32.63	1.00	0.00	18.79
01-CZ-M06_017	0.00	3.21	29.58	2.38	0.00	18.60
01-CZ-M06_018	0.01	2.19	31.51	1.48	0.00	18.67
01-CZ-M06_019	0.00	2.39	31.67	1.51	0.00	18.89
01-CZ-M06_020	0.01	1.82	32.52	0.83	0.00	18.98
01-CZ-M09_001	0.00	3.63	27.97	3.09	0.00	18.84
01-CZ-M09_002	0.01	3.17	28.97	2.54	0.00	19.02
01-CZ-M09_003	0.00	2.74	30.51	2.16	0.00	18.78
01-CZ-M09_004	0.00	3.33	28.92	2.41	0.00	18.85
01-CZ-M09_005	0.03	1.98	31.44	1.36	0.00	18.79
01-CZ-M09_006	0.01	2.26	31.10	1.46	0.00	18.91
01-CZ-M09_007	0.01	2.23	31.46	1.28	0.00	19.03
01-CZ-M09_008	0.00	1.92	31.54	1.14	0.00	18.80
01-CZ-M09_009R	0.01	2.24	31.02	1.63	0.00	18.97
01-CZ-M09_010	0.00	2.06	32.01	0.82	0.00	19.08
01-CZ-M09_011	0.01	2.06	31.97	1.28	0.00	18.92
01-CZ-M09_012	0.00	1.11	34.05	0.03	0.00	19.20
01-CZ-M09_013	0.00	1.91	32.15	1.10	0.00	18.78
01-CZ-M09_014	0.01	1.57	32.60	0.75	0.00	19.01
01-CZ-M09_015R	0.00	1.24	32.92	0.29	0.00	19.04
01-CZ-M09_016	0.00	1.51	32.49	0.64	0.00	18.80
01-CZ-M09_017	0.01	2.02	31.73	1.36	0.00	18.83
01-CZ-M09_018	0.00	1.91	32.20	1.04	0.00	18.92
01-CZ-M09_019	0.00	1.11	33.81	0.05	0.00	19.07
01-CZ-M09_020	0.00	4.06	26.91	3.70	0.00	18.80
01-CZ-NM09_001	0.00	1.73	32.17	1.00	0.00	18.97
01-CZ-NM09_002	0.01	1.33	33.21	0.47	0.00	19.82
01-CZ-NM09_003	0.01	1.58	33.01	0.71	0.00	19.15
01-CZ-NM09_004R	0.01	2.07	31.43	1.32	0.00	19.12
01-CZ-NM09_005	0.00	1.56	32.49	0.59	0.00	19.18
01-CZ-NM09_006R	0.01	2.58	29.52	2.06	0.00	18.82

01-CZ-NM09_007	0.00	1.78	32.54	1.11	0.00	19.22
01-CZ-NM09_008R	0.01	1.74	32.28	1.04	0.00	18.98
01-CZ-NM09_009	0.00	2.50	30.51	1.64	0.00	19.04
01-CZ-NM09_010	0.02	1.94	31.63	1.17	0.00	18.91
01-CZ-NM09_011	0.00	1.96	32.14	1.14	0.00	19.95
01-CZ-NM09_012Rb	0.00	1.55	32.97	0.65	0.00	18.98
01-CZ-NM09_013R	0.00	2.56	30.28	1.96	0.00	18.90
01-CZ-NM09_014R	0.00	1.93	32.47	1.26	0.00	19.26
01-CZ-NM09_015R	0.00	2.08	31.53	1.29	0.00	18.83
01-CZ-NM09_016	0.00	2.10	31.71	1.34	0.00	18.99
01-CZ-NM09_017	0.00	1.97	32.57	1.22	0.00	19.03
01-CZ-NM09_018	0.00	2.39	31.19	1.73	0.00	19.01
01-CZ-NM09_019	0.01	2.03	32.21	1.08	0.00	19.09
01-CZ-NM09_020	0.00	1.75	33.16	0.87	0.00	19.18
01-CZ-NM09_021	0.02	2.06	31.40	1.31	0.00	18.72
01-CZ-NM09_022	0.00	2.37	31.51	1.40	0.00	19.00
01-IZ-M0.3_001	0.00	1.95	32.22	0.89	0.00	18.83
01-IZ-M0.3_002	0.00	1.70	33.12	0.34	0.00	19.18
01-IZ-M0.3_003	0.00	1.67	33.41	0.36	0.00	19.18
01-IZ-M0.3_004	0.00	1.49	33.64	0.00	0.00	19.40
01-IZ-M0.3_005	0.00	1.52	33.84	0.29	0.00	19.20
01-IZ-M0.3_006	0.00	1.15	34.83	0.00	0.00	19.53
01-IZ-M0.3_007	0.01	1.57	33.54	0.27	0.00	19.08
01-IZ-M0.3_008	0.00	2.06	32.20	0.77	0.00	19.12
01-IZ-M0.3_009	0.00	1.90	32.80	0.34	0.00	19.23
01-IZ-M0.3_010	0.00	4.39	28.93	1.65	0.00	19.09
01-IZ-M0.3_011	0.01	1.78	32.72	0.89	0.00	18.84
01-IZ-M0.3_012	0.00	2.42	31.77	1.26	0.00	19.15
01-IZ-M0.3_013	0.00	1.41	33.99	0.06	0.00	19.46
01-IZ-M0.3_014	0.00	1.66	33.19	0.47	0.00	19.00
01-IZ-M0.3_015	0.01	2.00	32.17	0.73	0.00	19.06
01-IZ-M0.3_016	0.00	3.11	30.16	1.64	0.00	18.84
01-IZ-M0.3_017	0.00	1.19	34.03	0.00	0.00	19.40
01-IZ-M0.3_018	0.00	2.41	31.30	1.11	0.00	19.14
01-IZ-M0.3_019	0.00	1.79	32.68	0.44	0.00	19.19
01-IZ-M0.3_020	0.00	2.06	32.53	0.38	0.00	18.97
01-IZ-M0.6_001	0.00	1.35	33.95	0.21	0.00	19.14
01-IZ-M0.6_002	0.01	1.43	33.81	0.22	0.00	19.19
01-IZ-M0.6_003R	0.00	1.95	32.10	1.06	0.00	18.95
01-IZ-M0.6_004	0.00	1.24	34.12	0.19	0.00	19.48
01-IZ-M0.6_005	0.00	1.03	34.99	0.00	0.00	19.59
01-IZ-M0.6_006	0.00	1.73	33.30	0.09	0.00	19.36
01-IZ-M0.6_007	0.00	1.39	33.36	0.36	0.00	19.08
01-IZ-M0.6_008	0.00	1.50	33.15	0.28	0.00	19.06
01-IZ-M0.6_009	0.00	1.42	33.91	0.00	0.00	19.23
01-IZ-M0.6_010	0.01	3.04	29.64	2.15	0.00	19.00
01-IZ-M0.6_011	0.00	1.61	33.36	0.28	0.00	18.92

01-IZ-M0.6_012	0.00	2.30	31.84	0.88	0.00	19.03
01-IZ-M0.6_013	0.00	1.54	33.21	0.43	0.00	19.04
01-IZ-M0.6_014	0.00	1.46	33.11	0.57	0.00	19.00
01-IZ-M0.6_015	0.00	1.01	34.21	0.00	0.00	19.28
01-IZ-M0.6_016	0.02	2.37	32.25	0.58	0.00	18.97
01-IZ-M0.6_017	0.01	1.39	33.32	0.32	0.00	19.06
01-IZ-M0.6_018	0.00	1.86	31.86	0.87	0.00	18.69
01-IZ-M0.6_019	0.00	3.39	28.85	2.53	0.00	18.83
01-IZ-M0.6_020	0.00	1.46	32.79	0.48	0.00	18.90
HS-01-M09_001	0.00	0.94	35.08	0.00	0.00	19.23
HS-01-M09_003	0.01	1.45	33.53	0.43	0.00	19.30
HS-01-M09_004	0.01	1.72	33.09	0.71	0.00	19.31
HS-01-M09_005	0.00	1.49	33.97	0.24	0.00	19.65
HS-01-M09_006	0.00	1.01	34.65	0.00	0.00	19.31
HS-01-M09_007	0.00	1.97	32.28	1.30	0.00	19.33
HS-01-M09_008	0.00	1.44	34.09	0.24	0.00	19.12
HS-01-M09_009	0.00	1.81	33.07	0.57	0.00	19.13
HS-01-M09_010	0.00	2.49	31.09	1.77	0.00	19.00
HS-01-M09_011	0.00	3.19	29.62	2.28	0.00	18.79
HS-01-M09_012	0.00	2.61	30.63	1.89	0.00	18.77
HS-01-M09_013	0.01	1.08	34.36	0.02	0.00	19.34
HS-01-M09_014	0.01	1.62	33.02	0.60	0.00	19.03
HS-01-M09_015	0.00	2.07	32.10	1.14	0.00	19.06
HS-01-M09_016	0.00	1.34	33.59	0.43	0.00	19.04
HS-01-M09_017	0.00	2.16	31.88	1.26	0.00	18.90
HS-01-M09_018	0.00	1.44	33.10	0.59	0.00	18.95
HS-01-M09_019	0.00	1.51	33.42	0.44	0.00	18.86
HS-01-M09_020	0.00	1.86	32.43	0.93	0.00	18.96
NS-02-BP M2.1_001	0.00	1.82	33.89	0.00	0.00	18.86
NS-02-BP M2.1_002	0.00	1.22	34.88	0.00	0.00	19.26
NS-02-BP M2.1_003	0.01	1.53	34.33	0.00	0.00	19.25
NS-02-BP M2.1_004	0.00	1.57	34.34	0.00	0.00	19.06
NS-02-BP M2.1_005	0.01	1.78	34.24	0.00	0.00	18.89
NS-02-BP M2.1_006	0.00	1.10	34.93	0.00	0.00	19.34
NS-02-BP M2.1_007	0.01	1.64	34.32	0.00	0.00	19.04
NS-02-BP M2.1_008	0.01	1.82	34.21	0.00	0.00	18.99
NS-02-BP M2.1_009	0.00	1.64	34.03	0.00	0.00	19.11
NS-02-BP M2.1_010	0.01	2.20	33.67	0.00	0.00	19.22
NS-02-BP M2.1_011	0.01	1.03	34.60	0.00	0.00	19.05
NS-02-BP M2.1_012	0.00	0.91	35.11	0.00	0.00	19.41
NS-02-BP M2.1_013	0.00	2.33	33.47	0.00	0.00	18.92
NS-02-BP M2.1_014	0.01	1.55	34.21	0.00	0.00	19.01
NS-02-BP M2.1_015	0.01	1.82	33.74	0.00	0.00	19.11
NS-02-BP M2.1_016	0.01	1.41	34.17	0.00	0.00	19.00
NS-02-BP M2.1_017	0.02	1.48	34.07	0.00	0.00	18.69
NS-02-BP M2.1_018	0.01	2.36	33.28	0.00	0.00	18.90
NS-02-BP M2.1_019	0.00	1.21	34.42	0.00	0.00	18.92

NS-02-BP M2.1_020	0.00	1.79	33.89	0.00	0.00	18.79
NS-02-BP NM0.8_003	0.02	5.57	30.00	0.00	0.00	18.84
NS-02-BP NM0.8_004	0.03	1.72	33.66	0.03	0.00	19.11
NS-02-BP NM0.8_005	0.01	2.01	33.51	0.00	0.00	19.12
NS-02-BP NM0.8_006	0.00	2.17	33.65	0.00	0.00	18.99
NS-02-BP NM0.8_007	0.00	2.04	33.66	0.00	0.00	19.06
NS-02-BP NM0.8_008	0.00	1.98	33.82	0.00	0.00	19.11
NS-02-BP NM0.8_009	0.00	1.42	33.20	0.77	0.00	19.29
NS-02-BP NM0.8_010	0.00	2.00	33.55	0.00	0.00	19.08
NS-02-BP NM0.8_011	0.00	2.48	33.11	0.00	0.00	19.09
NS-02-BP NM0.8_012	0.01	1.55	34.20	0.00	0.00	19.04
NS-02-BP NM0.8_013	0.02	1.21	34.31	0.00	0.00	18.90
NS-02-BP NM0.8_014	0.00	1.24	33.97	0.30	0.00	19.27
NS-02-BP NM0.8_015	0.00	1.21	33.90	0.35	0.00	19.27
NS-02-BP NM0.8_016	0.01	2.02	33.81	0.00	0.00	18.89
NS-02-BP NM0.8_017	0.00	1.17	34.87	0.00	0.01	18.70
NS-02-BP NM0.8_018	0.00	1.47	33.78	0.23	0.00	19.34
NS-02-BP NM0.8_019	0.00	2.03	33.76	0.00	0.00	19.25
NS-02-BP NM0.8_020	0.01	1.34	34.58	0.00	0.00	19.43

As wt%	Se wt%	Te wt%	Bi wt%	Total
46.54	0.00	0.00	0.00	100.75
46.31	0.00	0.06	0.00	100.60
46.83	0.00	0.03	0.00	100.47
46.73	0.00	0.02	0.00	100.37
46.89	0.00	0.00	0.05	100.89
46.53	0.00	0.01	0.00	100.62
46.63	0.00	0.02	0.00	101.03
46.58	0.00	0.00	0.00	101.24
46.12	0.00	0.09	0.00	100.92
46.79	0.00	0.00	0.00	100.53
46.53	0.00	0.03	0.00	100.48
46.28	0.00	0.00	0.00	100.02
46.22	0.00	0.00	0.00	100.31
46.72	0.00	0.01	0.00	100.37
46.84	0.00	0.00	0.03	100.83
46.47	0.00	0.05	0.00	100.95
46.91	0.00	0.04	0.00	100.72
46.92	0.00	0.00	0.00	100.77
46.64	0.00	0.00	0.00	101.10
46.02	0.00	0.05	0.00	100.23
46.29	0.00	0.03	0.00	99.84
46.45	0.00	0.00	0.00	100.15
46.91	0.00	0.00	0.00	101.10
47.03	0.00	0.03	0.00	100.58
46.62	0.00	0.00	0.20	100.41
46.65	0.00	0.00	0.00	100.39
46.49	0.00	0.00	0.00	100.49
45.93	0.00	0.02	0.00	99.35
46.04	0.00	0.00	0.00	99.91
46.01	0.00	0.00	0.00	99.98
46.53	0.00	0.00	0.00	100.78
46.31	0.00	0.00	0.00	100.70
46.50	0.00	0.11	0.00	100.56
46.44	0.00	0.02	0.00	100.39
45.93	0.00	0.04	0.00	99.46
46.15	0.00	0.00	0.00	99.59
46.31	0.00	0.02	0.00	100.27
46.39	0.00	0.03	0.00	100.48
46.14	0.00	0.02	0.00	100.20
46.33	0.00	0.05	0.00	99.85
45.63	0.00	0.36	0.79	100.66
45.50	0.00	0.14	0.56	101.04
45.99	0.00	0.02	0.00	100.47
45.66	0.00	0.01	0.00	99.62
45.89	0.00	0.03	0.00	99.75
45.35	0.00	0.04	0.00	98.38

46.09	0.00	0.05	0.00	100.80
45.87	0.00	0.03	0.00	99.94
45.91	0.00	0.00	0.02	99.63
45.44	0.00	0.00	0.01	99.11
45.78	0.00	0.04	0.04	101.06
46.11	0.00	0.01	0.00	100.28
46.32	0.00	0.00	0.00	100.04
46.05	0.00	0.00	0.00	100.97
46.05	0.00	0.00	0.00	99.77
46.16	0.00	0.00	0.00	100.29
46.44	0.00	0.02	0.00	101.23
46.41	0.00	0.06	0.00	100.78
46.00	0.00	0.04	0.01	100.46
45.58	0.00	0.01	0.00	100.54
45.71	0.00	0.11	0.25	99.58
45.50	0.00	0.05	0.00	99.82
46.13	0.00	0.02	0.00	100.04
45.33	0.00	0.00	0.00	99.67
45.55	0.00	0.11	0.00	100.28
45.12	0.00	0.07	0.03	99.75
46.27	0.00	0.00	0.00	101.13
45.15	0.00	0.00	0.00	100.66
45.62	0.00	0.03	0.00	100.13
45.61	0.00	0.00	0.00	99.76
45.67	0.00	0.00	0.00	99.94
46.29	0.00	0.01	0.00	100.35
46.67	0.00	0.05	0.00	100.97
46.11	0.00	0.01	0.00	100.72
45.97	0.00	0.03	0.00	100.92
45.54	0.00	0.03	0.00	99.89
45.63	0.00	0.05	0.00	99.65
44.74	0.00	0.03	0.00	98.52
44.40	0.00	0.05	0.00	99.08
45.49	0.00	0.03	0.00	99.48
45.05	0.00	0.05	0.00	99.19
45.83	0.00	0.00	0.00	99.77
46.19	0.00	0.00	0.00	100.83
46.32	0.00	0.01	0.00	100.98
46.40	0.00	0.04	0.00	100.50
46.06	0.00	0.07	0.00	101.16
45.56	0.00	0.07	0.00	101.23
44.82	0.00	0.00	0.00	99.30
45.96	0.00	0.03	0.00	100.18
45.41	0.00	0.06	0.00	99.46
45.49	0.00	0.00	0.00	100.05
46.15	0.00	0.07	0.00	100.06
45.70	0.00	0.00	0.00	99.87

45.54	0.00	0.05	0.00	99.64
45.77	0.00	0.06	0.00	100.04
46.02	0.00	0.01	0.01	100.17
45.26	0.00	0.06	0.00	99.82
45.95	0.00	0.00	0.00	100.14
45.95	0.00	0.07	0.00	100.12
45.50	0.00	0.03	0.00	98.82
45.69	0.00	0.00	0.02	99.32
45.57	0.00	0.00	0.00	99.20
46.27	0.00	0.09	0.00	101.61
46.76	0.00	0.06	0.00	101.55
46.87	0.00	0.02	0.00	101.74
46.45	0.00	0.02	0.00	101.82
46.44	0.00	0.00	0.00	101.41
46.68	0.00	0.00	0.00	101.57
46.61	0.00	0.04	0.00	101.54
46.64	0.00	0.00	0.00	101.22
46.87	0.00	0.01	0.00	101.23
46.72	0.00	0.04	0.00	100.64
46.77	0.00	0.08	0.00	100.73
46.35	0.00	0.12	0.00	101.27
46.43	0.00	0.00	0.00	100.71
46.49	0.00	0.04	0.00	100.91
46.57	0.00	0.00	0.00	100.98
46.67	0.00	0.04	0.00	100.92
46.57	0.00	0.03	0.00	100.68
46.60	0.00	0.00	0.00	100.83
46.75	0.00	0.05	0.00	101.00
46.60	0.00	0.01	0.00	101.18
46.19	0.00	0.06	0.00	101.61
46.37	0.00	0.00	0.00	101.49
46.52	0.00	0.03	0.00	101.50
46.64	0.00	0.00	0.00	101.55
46.31	0.00	0.04	0.02	101.75
46.82	0.00	0.00	0.00	101.84
46.58	0.00	0.00	0.00	101.62
46.38	0.00	0.03	0.00	101.20
46.31	0.00	0.02	0.00	101.42
46.08	0.00	0.00	0.00	100.77
45.57	0.00	0.01	0.00	101.01
46.12	0.00	0.08	0.00	100.92
46.07	0.00	0.01	0.00	100.86
45.71	0.00	0.00	0.00	100.40
46.03	0.00	0.04	0.00	100.66
45.97	0.00	0.04	0.00	100.26
46.07	0.00	0.00	0.00	100.63
45.90	0.00	0.01	0.00	100.47

46.17	0.00	0.00	0.00	100.65
45.69	0.00	0.00	0.00	100.11
46.36	0.00	0.09	0.00	101.01
46.60	0.00	0.00	0.00	101.25
46.57	0.00	0.01	0.00	101.39
46.54	0.00	0.00	0.00	101.30
46.26	0.00	0.00	0.00	101.18
46.04	0.00	0.00	0.00	100.71
46.49	0.00	0.02	0.00	101.13
46.36	0.00	0.06	0.00	101.11
46.30	0.00	0.01	0.03	101.14
46.14	0.00	0.00	0.00	100.57
45.94	0.00	0.00	0.00	100.72
46.06	0.00	0.00	0.00	100.81
46.14	0.00	0.07	0.00	100.93
46.89	0.00	0.00	0.00	101.63
46.39	0.00	0.01	0.00	101.23
46.50	0.00	0.04	0.00	101.57
46.23	0.00	0.02	0.00	101.61

Highlights

1. Variably magnetic fractions of cobaltite in the ICB have low Re and total Os contents (ca. 0.4−4 ppb and 14−64 ppt, respectively) but elevated $^{187}Re/^{188}Os$ (600−1800) and highly radiogenic $^{187}Os/^{188}Os$ (17−45) ratios. The Black Pine prospect has too low Re concentrations (<0.3 ppb) to produce any reliable Re-Os data. By contrast, Re-Os data of cobaltite from the quartz-tourmaline-cobaltite breccia Haynes-Stellite deposit are accurate and reproducible providing a minimum aliquot size of 200 mg whereas data for cobaltite from the quartz-tourmaline-cobaltite breccia of the Idaho Zone require a minimum aliquot size of 150 mg.

2. The Re-Os systematics in cobaltite at Haynes-Stellite and at the Idaho Zone are preserved and Re-Os data are regressed using the isochron approach. By contrast, the Re-Os systematics of cobaltite in the Chicago Zone are disturbed and a linear array of mixing with post-depositional hydrothermal fluids is identified in the $^{187}Os/^{188}Os$ vs. $1/^{192}Os$ space. Cobaltite was primarily deposited in the ICB at Haynes Stellite at 1349 ± 76 Ma and was preceded by REE-Y-mineral precipitation at ca. 1370 Ma in connection with the emplacement of the Big Deer Creek granite at ca. 1377 Ma as part of a suite of bimodal magmatic activity between ca. 1383 and 1359 Ma (Aleinikoff et al., 2012). We propose that our age of 1132 ± 240 Ma for cobaltite in the quartz-tourmaline breccia in the Idaho Zone and consideration of other geologic and structural constraints (Bookstrom et al., 2007, 2016) illustrate a remobilization phase of pre-existing cobaltite mineralization along cleavages and folds during the 1190−1006 Ma Grenvillian orogeny.

3. Cobaltite formed at ca. 1349 Ma upon cooling at an inferred temperature of ≤ 300ºC of a mixture of Mesoproterozoic evaporitic brines and magmatic fluids that carried both metals and reduced sulfur. The extremely high initial Os ratio derived from regression of the Re-Os data at Haynes-Stellite (4.7 ± 2.2) is used a tracing tool to identify the source of metals. The most likely scenario considers that metals were derived from lower crustal mafic source rocks of probable Archean age, overlain by oceanic island-arc rocks of Paleoproterozoic age, which are tectonically interlayered with Archean rocks of the Wyoming Province continental nucleus, in the collisional

Great Falls tectonic zone, which projects beneath the Idaho Cobalt Belt. This scenario is

compatible with the sulfur isotopic composition of cobaltite ($+8.0 \pm 0.4‰$) that is best explained

by reduced sulfur being sourced from deep, crustal and highly metamorphosed rocks (Johnson et

al., 2012).

4. We propose that the Re-Os isotope system in cobaltite in the Chicago Zone, located in a middle

greenschist facies metamorphic zone, may have been subjected to temperatures of 400 to 520ºC

proposed for Cretaceous garnet-bearing rocks of a lower amphibolites facies metamorphic domain

above the domain hosting the Chicago Zone. A maximum closing temperature of 400ºC is

inferred for Re-Os in cobaltite. By contrast, cobaltite at Haynes-Stellite and the Idaho Zone, which

underwent middle to lower greenschist Cretaceous metamorphism, preserved Mesoproterozoic

age information delivered by the Re-Os isotope system.