

Economic Geology

Elevated magmatic sulfur and chlorine contents in ore-forming magmas at the Red Chris porphyry Cu-Au deposit, Northern British Columbia, Canada

--Manuscript Draft--

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Abstract:	<p>The Red Chris porphyry Cu-Au deposit is located in the Stikinia island-arc terrane in northwest British Columbia. It is hosted by the Red Stock, which has four phases of porphyry intrusions: P1, P2E, P2L, and P3. New U-Pb dating of zircon shows that these intrusions were emplaced at 211.6 ± 1.3 Ma (MSWD = 0.85), 206.0 ± 1.2 Ma (MSWD = 1.5), 203.6 ± 1.8 Ma (MSWD = 1.5), and 201.7 ± 1.2 Ma (MSWD = 1.05), respectively. The ore-forming event at Red Chris was a short-lived event at 206.1 ± 0.5 Ma (MSWD = 0.96; weighted average age of three Re-Os analyses), implying a duration of <1 m.y., as defined by the uncertainty range. This mineralization age coincides with the emplacement age of the P2E porphyry, and is consistent with cross-cutting relationships that suggest P2E was the main syn-mineralization intrusion. Zircons from P1 to P3 porphyry rocks have consistently high EuN/EuN* ratios (mostly > 0.4), indicating that their associated magmas were moderately oxidized. The magmatic water contents estimated from plagioclase and amphibole compositions suggest H₂O contents of ~5 wt. %. Taken together, the P1 to P3 porphyries are interpreted to be moderately oxidized and hydrous.</p> <p>The four phases of porphyries are differentiated by sulfur and chlorine contents. The SO₃ contents of igneous apatite microphenocrysts from the mineralization-related P2 porphyries are higher (P2E: 0.30 ± 0.13 wt. %, n = 34; P2L: 0.29 ± 0.18 wt. %, n = 100) than those from the pre-mineralization P1 (0.11 ± 0.03 wt. %, n = 34) and post-mineralization P3 porphyries (0.03 ± 0.01 wt. %, n = 13). The chlorine contents in apatite grains from the P2E and P2L porphyries are 1.18 ± 0.37 (n = 34) and 1.47 ± 0.28 wt. % (n = 100), also higher than those from P1 (0.51 ± 0.3 wt. % Cl, n = 34) and P3 (0.02 ± 0.02 wt. % Cl, n = 17). These results imply that the sulfur and chlorine</p>

	<p>contents of the P2E and P2L magmas were higher than in the P1 and P3 magmas, suggesting that elevated magmatic S-Cl contents in the P2 porphyries may have been important for ore-formation. Although the process that caused the increase in sulfur and chlorine is not clear, reverse zoning seen in plagioclase phenocrysts from the P2 porphyry, and the occurrence of more mafic compositions in P2L suggest that recharge of the deeper magma chamber by a relatively S-Cl-rich mafic magma may have triggered the ore-forming hydrothermal event.</p>
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Dear Editor,

Thank you again for giving us the opportunity to improve our paper entitled “Elevated magmatic sulfur and chlorine contents in ore-forming magmas at the Red Chris porphyry Cu-Au deposit, Northern British Columbia, Canada” (ID: SEG-D-18-00013R1). Particularly, we appreciate the very careful review on this manuscript by the Associated Editor, David R. Cooke.

We agree with the comments by Dr. David Cooke and all the corresponding corrections have been done following his instructions. To facilitate review of our revised paper, the version with marked changes is given. We hope our manuscript could be accepted soon.

With thanks and regards,

Jing-Jing Zhu for the co-authors.

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4 **1 Elevated magmatic sulfur and chlorine contents in ore-forming magmas at the Red**
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6 **2 Chris porphyry Cu-Au deposit, Northern British Columbia, Canada**

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39 14 **Abstract**

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44 16 northwest British Columbia. It is hosted by the Red Stock, which has four phases of
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56 22 defined by the uncertainty range. This mineralization age coincides with the emplacement
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4 23 age of the P2E porphyry, and is consistent with cross-cutting relationships that suggest
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7 24 P2E was the main syn-mineralization intrusion.

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9 25 Zircons from P1 to P3 porphyry rocks have consistently high $\text{Eu}_N/\text{Eu}_N^*$ ratios (mostly >
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11 26 0.4), indicating that their associated magmas were moderately oxidized. The magmatic
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14 27 water contents estimated from plagioclase and amphibole compositions suggest H_2O
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16 28 contents of ~5 wt. %. Taken together, the P1 to P3 porphyries are interpreted to be
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19 29 moderately oxidized and hydrous.

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21 30 The four phases of porphyries are differentiated by sulfur and chlorine contents. The
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24 31 SO_3 contents of igneous apatite microphenocrysts from the mineralization-related P2
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26 32 porphyries are higher (P2E: 0.30 ± 0.13 wt. %, $n = 34$; P2L: 0.29 ± 0.18 wt. %, $n = 100$)
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29 33 than those from the pre-mineralization P1 (0.11 ± 0.03 wt. %, $n = 34$) and post-
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31 34 mineralization P3 porphyries (0.03 ± 0.01 wt. %, $n = 13$). The chlorine contents in apatite
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33 35 grains from the P2E and P2L porphyries are 1.18 ± 0.37 ($n = 34$) and 1.47 ± 0.28 wt. %
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35 36 ($n = 100$), also higher than those from P1 (0.51 ± 0.3 wt. % Cl, $n = 34$) and P3 ($0.02 \pm$
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37 37 0.02 wt. % Cl, $n = 17$). These results imply that the sulfur and chlorine contents of the
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40 38 P2E and P2L magmas were higher than in the P1 and P3 magmas, suggesting that
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43 39 elevated magmatic S-Cl contents in the P2 porphyries may have been important for ore-
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45 40 formation. Although the process that caused the increase in sulfur and chlorine is not
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48 41 clear, reverse zoning seen in plagioclase phenocrysts from the P2 porphyry, and the
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51 42 occurrence of more mafic compositions in P2L suggest that recharge of the deeper
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53 43 magma chamber by a relatively S-Cl-rich mafic magma may have triggered the ore-
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55 44 forming hydrothermal event.

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Introduction

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47 Compared to the relatively long-lived magmatic systems represented by a composite
48 batholith (up to 10 m.y.; Matzel et al., 2006; Walker et al., 2007; Harris et al., 2008;
49 Schaltegger et al., 2009; Paterson et al., 2011; Rezeau et al., 2016), the life spans of
50 porphyry magmatic-hydrothermal ore-forming events are much shorter, probably on the
51 order of several hundred thousand years (up to 1 m.y.; Cathles et al., 1997; Hayba and
52 Ingebritsen, 1997; Masterman et al., 2004; von Quadt et al., 2011; Weis et al., 2012;
53 Chiaradia et al., 2013; Correa et al., 2016; Li et al., 2017). In many large magmatic
54 systems, porphyry formation occurs at a relatively late stage in the system's evolution
55 (Creasey, 1977; Titley and Beane, 1981; Candela, 1992; Richards, 2003; Rohrlach and
56 Loucks, 2005; Yang et al., 2009; Correa et al., 2016).

57 Although the multiple phases of porphyry stocks are commonly broadly cogenetic,
58 they may be derived from packets of magma that evolved at different crustal levels over
59 the history of the larger magmatic system (Annen et al., 2006). Understanding why ore-
60 formation is only associated with a specific intrusive suite within these broader systems,
61 and at discrete, commonly singular times, is a focus of this study.

62 It is recognized that magmas with high sulfur (>1000 ppm), chlorine (>3000 ppm),
63 and water (>4 wt. %) contents as well as relatively high oxidation states (higher than the
64 fayalite-magnetite-quartz buffer, ΔFMQ , by 1–2 log $f\text{O}_2$ units) are fertile for the
65 generation of magmatic-hydrothermal porphyry Cu deposits (Burnham, 1979; Candela,
66 1992; Richards, 2003, 2009, 2011, 2015; Wallace, 2005; Chambefort et al., 2008; Simon
67 and Ripley, 2011; Chiaradia et al., 2012; Loucks, 2014; Hou et al., 2015; Lu et al., 2015,
68 2016; Chelle-Michou and Chiaradia, 2017). These ingredients are essential for the

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4 69 transport of Cu (and Au) both in the magma (Zajacz et al., 2012) and in exsolved high
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6 70 temperature, SO₂-rich, saline magmatic fluids (Candela and Holland, 1984; Zajacz et al.,
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9 71 2011). However, whereas many magmas achieve such compositions, ore formation is a
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11 72 relatively rare and discrete event in such systems.

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14 73 Various mechanisms have been proposed that might trigger an ore forming event
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16 74 from a fertile magma source, including recharge of the magma chamber by hot, sulfur-
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18 75 rich melts (Hattori, 1993; Keith et al., 1997; Larocque et al., 2000; Hattori and Keith,
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20 76 2001; Halter et al., 2005; Stern et al., 2007; Nadeau et al., 2010, 2016; Wilkinson, 2013;
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22 77 Tapster et al., 2016; Zhang and Audétat, 2017), fluxing by sulfur gases (Blundy et al.,
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24 78 2015), or increasing the water and oxidation state of magmas during long-term
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26 79 fractionation in deep magma chambers (Ballard et al., 2002; Wang et al., 2014a, 2014b;
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28 80 Dilles et al., 2015; Lu et al., 2015, 2016). The Red Chris porphyry Cu-Au deposit
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30 81 provides an opportunity to test these hypotheses, because ore-formation occurred at a
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32 82 discrete and relatively late stage in the ~10 m.y. history of the associated magmatic
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34 83 system.

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36 84 The Red Chris deposit is located in northwest British Columbia, and contains
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38 85 measured and indicated resources of 1,035 million metric tonnes of ore grading 0.35%
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40 86 Cu and 0.35 g/t Au (Gillstrom et al., 2012). Pre-, syn-, and post-mineralization porphyry
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42 87 intrusions have been recognized based on detailed core logging and petrographic work
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44 88 (Rees et al., 2015), with mineralization occurring in a singular episode during this
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46 89 magmatic history. In this paper, we present detailed geochronological and geochemical
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48 90 data for the porphyry phases at Red Chris, and show that they have similar bulk
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50 91 compositions, including all being relatively hydrous and oxidized. However, the syn-
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92 mineralization porphyry is characterized by plagioclase with reverse zoning and apatite
93 with relatively high sulfur and chlorine contents. We interpret these data to indicate that
94 high magmatic S and Cl contents, in addition to high H₂O contents and oxidation state,
95 were critical for Cu-Au ore-formation at Red Chris. We speculate that there was an
96 injection of relatively mafic (hotter, and more S- and Cl-rich) but cogenetic magma into
97 the mid–upper crustal source magma chamber approximately coincident with
98 emplacement of the syn-mineralization porphyry.

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100 **Geological Setting**

101 Red Chris is situated within the island-arc terrane of Stikinia in the Intermontane Belt
102 of the Canadian Cordillera of British Columbia (Fig. 1; Monger and Irving, 1980; Nelson
103 and Colpron, 2007; Nelson et al., 2013). Stikinia consists primarily of Mesozoic arc-
104 related igneous and sedimentary rocks, formed in response to subduction processes prior
105 to accretion to the ancestral North American margin in the Middle Jurassic (Gabrielse,
106 1991; Ricketts et al., 1992; Lang et al., 1995; Mihalynuk et al., 2004; Nelson and Colpron,
107 2007; Logan and Mihalynuk, 2014). In northwestern Stikinia, arc assemblages comprise
108 the Middle to Late Triassic Stuhini Group, unconformably overlain by Late Triassic and
109 Early to Middle Jurassic volcanic and sedimentary rocks of the Hazelton Group (Fig. 2;
110 Brown et al., 1991; Gabrielse, 1991; Marsden and Thorkelson, 1992; Dostal et al., 1999;
111 Gagnon et al., 2012; Nelson et al., 2013; Logan and Mihalynuk, 2014). The Hazelton
112 Group is overlain by sedimentary rocks of the syn- to post-accretion Middle Jurassic to
113 Early Cretaceous Bowser Lake Group.

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4 114 Several porphyry Cu±Au±Mo deposits occur in the region, hosted by Late Triassic to
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6 115 Early Jurassic arc-related plutons (Fig. 1), including Red Chris which formed in the Late
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8 116 Triassic Red Stock. Collectively, these regional intrusions have ages ranging from ~222
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10 117 Ma to ~180 Ma, recording much of the pre-accretionary history of Stikinia (Lang et al.,
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12 118 1995; McMillan et al., 1995; Scott et al., 2008; Nelson et al., 2013; Logan and Mihalynuk,
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14 119 2014). A significant percentage of the known porphyry Cu deposits, including Red Chris,
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16 120 formed during a 6-m.y. pulse of magmatism between 206 and 200 Ma, with compositions
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18 121 ranging from calc-alkaline to strongly alkaline (Lang et al., 1995; McMillan et al., 1995;
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20 122 Nelson et al., 2013; Bissig and Cooke, 2014; Logan and Mihalynuk, 2014; Micko et al.,
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22 123 2014; Pass et al., 2014).
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125 **Geology of the Red Chris Cu-Au deposit**

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36 126 The Red Chris Cu-Au deposit was discovered in the 1950s and explored
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38 127 intermittently in subsequent decades, with mining beginning in 2015 (Ash et al., 1995,
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40 128 1996; Newell and Peatfield, 1995; Baker et al., 1997; Gillstrom et al., 2012; Rees et al.,
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42 129 2015). The deposit is hosted by the Red Stock, which is the largest of a suite of Late
43
44 130 Triassic to Early Jurassic stocks and dikes that intrude the Stuhini Group in the district
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46 131 (Fig. 2; Friedman and Ash, 1997; Rees et al., 2015). The stock is tabular, elongate in an
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48 132 east to northeast direction, and approximately 6.5 km long by 300 to 1500 m wide (Fig. 2;
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50 133 Ash et al., 1995; Baker et al., 1997; Gillstrom et al., 2012). It has a steep northern contact
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52 134 against Stuhini Group country rocks, but its southeastern margin against Hazelton and
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54 135 Bowser Lake Group strata is poorly exposed, and has locally been truncated by the NE-
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4 136 trending, steeply SE-dipping South Boundary fault (Fig. 2). This fault, and the East zone
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6 137 fault within the stock, probably reflect a long-lived and deep structure which guided the
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9 138 emplacement of the intrusions, mineralization, and subsequent deformation of the Red
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11 139 Stock.

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15 140 The Red Stock is a composite intrusion consisting of several phases of porphyritic
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17 141 diorite to quartz monzonite. Mineralogically, the rocks consist of medium- to coarse-
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19 142 grained amphibole, plagioclase, and minor biotite phenocrysts, with K-feldspar,
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22 143 plagioclase, and quartz in the groundmass. Based on compositional and textural
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24 144 differences and crosscutting relationships, Rees et al. (2015) identified four distinct
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26 145 porphyry phases, P1, P2, P3, and P4. The P1 porphyry is a pre-mineralization
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28 146 leucodiorite which accounts for the main volume of the Red Stock, and is distinguished
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30 147 by sparse anhedral to subhedral amphibole (~10%) and abundant plagioclase (30–40
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32 148 vol. %) phenocrysts with lengths up to 4 mm. The groundmass is composed of fine-
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34 149 grained plagioclase and minor quartz (Fig. 3A).

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40 150 The P2 porphyry is a syn-mineralization quartz monzonite intrusion which is largely
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42 151 unexposed at surface but is observed in drill core to have intruded P1 at depth. It has been
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44 152 subdivided by Rees et al. (2015) into early (P2E), intermediate (P2I), and late (P2L)
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46 153 phases based on vein truncations and chilled margins. The P2 porphyries are generally
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48 154 characterized by tabular subhedral to euhedral amphibole (10–15 vol. %) and plagioclase
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50 155 (35–50 vol. %) phenocrysts. The amphibole crystals are mostly euhedral and larger than
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52 156 in P1, with lengths up to 10 mm. The groundmass comprises K-feldspar, plagioclase, and
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54 157 quartz. In this paper, the P2 subphases are simplified to an early stage (P2E; Fig. 3B) and
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56 158 late stage porphyry (P2L, probably corresponding to P2I and P2L of Rees et al., 2015;
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4 159 Fig. 3C). The P2E porphyry has a crowded plagioclase texture with slightly smaller
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6 160 amphibole phenocrysts than P2L. In contrast, the P2L porphyry is relatively fresh and
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9 161 occurs only as small dikes (Fig. 4).

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12 162 The post-mineralization P3 monzonite porphyry is much less abundant. It is texturally
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15 163 similar to the P2L porphyry with abundant amphibole phenocrysts (15–20 vol. %), but is
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17 164 distinguished by the absence of quartz in the groundmass, which is mainly composed of
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20 165 K-feldspar and plagioclase (Fig. 3D). Although no crosscutting relationships between the
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22 166 P3 and P2L porphyries have been found, the zircon U-Pb dating results (see below)
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24 167 confirm that P3 is younger. The P4 porphyry occurs as rare dikes and is typified by
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27 168 sparse fine-grained amphibole phenocrysts (Rees et al., 2015). No P4 samples were
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30 169 included in this study.

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33 170 Several younger basaltic to andesitic dikes with sparse amphibole phenocrysts cut the
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35 171 Red Stock and the Stuhini Group host rocks (Figs. 3E and F). They postdate the
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37 172 porphyry-stage Cu-Au mineralization (Baker et al., 1997; Rees et al., 2015), but are
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40 173 mildly to strongly altered (Figs. 3E and F) and are crosscut by late quartz-calcite-pyrite
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43 174 veins.

44 45 175 *Hydrothermal alteration*

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49 176 Alteration at Red Chris has been described previously by Baker et al. (1997),
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51 177 Gillstrom et al. (2012), Norris (2012) and Rees et al. (2015). The alteration assemblages
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53 178 are typical of calc-alkaline porphyry Cu systems (Lowell and Guilbert, 1970; Seedorff et
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56 179 al., 2005; Sillitoe, 2010), and consists of early stage potassic alteration, overprinted by
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59 180 chlorite–sericite, sericitic (phyllic), intermediate argillic, and minor late stage propylitic

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4 181 alteration (see paragenetic diagram in Rees et al., 2015). Potassic alteration is expressed
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6 182 by replacement of amphibole phenocrysts by secondary biotite, plagioclase replaced or
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9 183 rimmed by secondary K-feldspar, and by K-feldspar veins (Figs. 3B and 5A–C). It is best
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11 184 preserved in the deeper levels of the deposit where it is spatially associated with the syn-
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14 185 mineralization P2E porphyry; however it locally extends into pre-mineralization P1
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16 186 porphyry wall rocks (Fig. 4). The P2L and P3 porphyries were only weakly affected by
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19 187 potassic alteration (Rees et al., 2015).
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22 188 Chlorite–sericite alteration is characterized by chlorite replacing secondary biotite
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24 189 (Fig. 5C) and sericite replacing feldspar (Fig. 5D). At shallower levels in the system,
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27 190 potassic alteration is completely overprinted by phyllic and intermediate argillic
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29 191 alteration (Gillstrom et al., 2012), characterized by sericite after plagioclase (phyllic; Fig.
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32 192 5D), and illite and kaolinite (intermediate argillic; Norris, 2012). This lower temperature
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34 193 alteration overprint affects all the porphyry phases, but is less pervasive at depth.
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38 194 Propylitic alteration at Red Chris is mainly observed as minor chlorite and epidote in
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40 195 the outer part of the Red Stock, and extends for 100 to 200 m into the Stuhini volcanic
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42 196 country rocks (Gillstrom et al., 2012; Norris, 2012; Rees et al., 2015).
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45 197 *Vein styles and mineralization*

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49 198 Detailed descriptions of vein styles at Red Chris have been given by Norris (2012)
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51 199 and Rees et al. (2015). A-type quartz veins (Gustafson and Hunt, 1975) and stockworks
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53 200 are associated with potassic alteration and host the bulk of the copper-gold mineralization.
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56 201 These veins are most intensely developed around the apex of the principal P2 porphyry
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59 202 body, but extend for hundreds of metres into the P1 wall rocks. Typical A veins contain
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4 203 K-feldspar, biotite, chalcopyrite, bornite, and magnetite, with K-feldspar alteration halos
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6 204 (Figs. 5A–B and G). Copper sulfides also occur as disseminations in the host porphyry.
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9 205 Bornite is more abundant in the apex of P2E, and progressively decreases outwards
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11 206 where chalcopyrite is the dominant Cu-sulfide (Norris, 2012; Rees et al., 2015). At
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14 207 shallower depths in the preserved system, early bornite was sulfidized to chalcopyrite,
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16 208 and pyrite becomes increasingly dominant. Rees et al. (2015) delineated a high-sulfur
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19 209 contour (>4% S) in section above which total sulfide (dominantly pyrite) ranges from 4 to
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21 210 10%. Microscopic native gold and electrum occur as inclusions in bornite (Rees et al.,
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23
24 211 2015). The grades of Cu and Au are positively correlated with quartz vein density
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26 212 (Gillstrom et al., 2012). In high-grade zones at depth (e.g., 4.12 % Cu and 8.83 g/t Au in
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29 213 hole 09-350 from 540 to 692.5 m down-hole depth), vein abundance exceeds 80 vol. % in
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31 214 sheeted arrays (Fig. 5G; Rees et al., 2015).

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34 215 B- and D-type veins (Gustafson and Hunt, 1975) are relatively minor at Red Chris,
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36 216 and host only minor amounts of Cu sulfides. B quartz veins are characterized by
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39 217 relatively straight margins with sulfide centerlines (pyrite and minor chalcopyrite, and
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42 218 locally molybdenite; Fig. 5E). Pyritic D veins have variable widths (1 to 10 mm; Fig. 5F).
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44 219 Carbonate and minor chlorite veins cut all the earlier veins and are generally barren
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47 220 (Norris, 2012).

221 **Sampling and Analytical Methods**

222 Samples of the P1, P2E, P2L, and P3 porphyries were collected from drill core, and
223 descriptions and locations are listed in Digital Appendix Table A1; sampled drill hole
224 locations are also shown on Figure 2. Fourteen least-altered samples of the porphyry
225 intrusions and three samples of late basaltic to andesitic dikes were selected for whole-

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4 226 rock geochemical analysis. Eight of these samples were selected for determination of Nd-
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6 227 Sr isotopes. Three samples of quartz-carbonate-pyrite-molybdenite-chalcopyrite veins
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9 228 (Fig. 5H) were collected for Re-Os dating. Four samples of the P1 (RC13-40), P2E
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11 229 (RC13-107), P2L (RC13-33), and P3 (RC13-78) intrusions were selected for zircon U-Pb
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14 230 dating, Hf isotopic, and trace element analyses. Details of analytical methods are
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16 231 provided in Appendix 1.
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21 233 *Electron microprobe analyses*
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23 234 Primary igneous minerals such as plagioclase and amphibole in the porphyry rocks
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26 235 are widely altered to K-feldspar, sericite, and chlorite (Figs. 3A–B and 5C–D). However,
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29 236 a few least-altered samples of the P2E, P2L, and P3 porphyries contained unaltered
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31 237 plagioclase and amphibole grains. In addition, igneous apatite grains were typically
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34 238 preserved as inclusions within plagioclase and amphibole phenocrysts. Compared with
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36 239 hydrothermal apatite (acicular crystals intergrown with other hydrothermal minerals such
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39 240 as quartz, sericite, chlorite, and sulfides), igneous apatite grains typically showed stubby
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41 241 prismatic habits, as described by Richards et al. (2017). Detailed analytical methods are
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43 242 described in Appendix 1.
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48 244 *Re-Os molybdenite dating*
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50 245 Three samples of molybdenite from quartz-carbonate-pyrite-molybdenite-chalcopyrite
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53 246 veins were collected from the Gully zone (RC13-88 and RC13-103), and the East zone
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56 247 (RC13-82; Digital Appendix Table A1; Fig. 2). A molybdenite mineral separate was
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58 248 produced for each sample by metal-free crushing followed by gravity and magnetic
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4 249 concentration methods. Dating was conducted at the Canadian Centre for Isotopic
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7 250 Microanalysis at the University of Alberta, Canada, using methods described in
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9 251 Appendix 1.

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14 253 **Geochronological Results**

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16 254 *Zircon U-Pb ages of the Red Stock*

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19 255 Zircon U-Pb results are presented in Digital Appendix Table A2 and illustrated in
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21 256 Figure 6; all ages are illustrated and reported with 2σ errors. All the zircons show
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24 257 oscillatory zoning under BSE imaging. Sample RC13-40 was collected from P1
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26 258 leucodiorite porphyry (Digital Appendix Table A1). Analyzed zircon grains form a
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29 259 tightly clustered age population, mostly with low common lead contents. Except for one
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31 260 inherited or xenocryst zircon (apparent $^{206}\text{Pb}/^{238}\text{U}$ age = 261 ± 12 Ma), the twenty-seven
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34 261 grains yielded an intercept age of 211.6 ± 1.3 Ma (MSWD = 0.85; Fig. 6A), similar to the
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36 262 weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 211.8 ± 1.3 Ma (MSWD = 0.8).

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38 263 Sample RC13-107 was collected from P2E quartz monzonite porphyry (Digital
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41 264 Appendix Table A1). One xenocrystic zircon grain yielded an older age (apparent
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43 265 $^{206}\text{Pb}/^{238}\text{U}$ age = 316 ± 10 Ma), but the other twenty-nine zircons contained low amounts
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46 266 of common lead and intersected the concordia line at 206.0 ± 1.2 Ma (MSWD = 1.5; Fig.
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48 267 6B), in good agreement with the weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 205.9 ± 1.5 Ma
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51 268 (MSWD = 1.5).

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53 269 Sample RC13-33 was collected from P2L quartz monzonite porphyry (Digital
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56 270 Appendix Table A1). Two xenocryst grains were found among thirty-one analyzed
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58 271 zircons, with $^{206}\text{Pb}/^{238}\text{U}$ ages of 298 ± 16 Ma and 441 ± 13 Ma. The remaining twenty-

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4 272 nine zircon grains had low common lead contents and yielded an intercept age of $203.6 \pm$
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6 273 1.8 Ma (MSWD = 1.5; Fig. 6C), similar to the weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of $203.3 \pm$
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9 274 1.5 Ma (MSWD = 1.04).

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11 275 Sample RC13-78 was collected from P3 monzonite porphyry (Digital Appendix Table
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14 276 A1). Thirty-two zircons with low common lead contents yielded an intercept age of 201.7
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16 277 ± 1.2 Ma (MSWD = 1.05; Fig. 6D), and a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 201.6 ± 1.2
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19 278 Ma (MSWD = 1.04).

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21 279 All of the zircons have magmatic textures, and the intercept ages above are interpreted
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24 280 to be the crystallization ages of the individual intrusions. The ages are consistent with
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26 281 crosscutting relationships described by Rees et al. (2015), and define a 10 m.y. span of
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29 282 magmatism from 211.6 ± 1.3 Ma (P1) to 201.7 ± 1.2 Ma (P3). The relative ages of the
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31 283 two samples of syn-mineralization P2 porphyry are consistent with crosscutting
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34 284 relationships, but the dates (P2E: 206.0 ± 1.2 Ma; P2L: 203.6 ± 1.8 Ma) overlap within
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36 285 the 2σ uncertainty. Hence the apparent 2.4 m.y. age difference is not statistically robust,
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39 286 and their true ages may in fact be closer as suggested by their close relationship with ore
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41 287 mineral paragenesis.

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46 289 *Re-Os molybdenite ages*

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48 290 The Re-Os model ages for three samples are shown in Table 1. They have relatively
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51 291 high rhenium contents ranging from 497.8 to 1771 ppm, with common ^{187}Os of 1078 to
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53 292 3821 ppb. Although the three samples are from two different mineralization zones (i.e.,
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56 293 East and Gully zones; Fig. 2), they yielded indistinguishable model ages within
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58 294 uncertainty: 206.5 ± 0.8 Ma, 205.7 ± 0.9 Ma, and 205.9 ± 1.1 Ma (± 2 s.d.). The small
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4 295 grain size of the molybdenite (< 1 mm) minimizes the risk of decoupling between Re and
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6 296 ¹⁸⁷Os in these samples (caused by diffusion: Selby and Creaser, 2004). The results
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9 297 yielded a weighted average age for all three samples of 206.1 ± 0.5 Ma (95% confidence
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11 298 level with MSWD = 0.96; Fig. 7). This age is consistent with the age of the syn-
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14 299 mineralization P2E porphyry (206.0 ± 1.2 Ma).
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18 301 **Geochemical and Isotopic Results**

19 302 *Whole-rock major and trace elements*

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23 303 Whole-rock major and trace element compositions for fourteen samples of the Red
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25 304 Stock and three basaltic–andesitic dike samples are listed in Table 2. All the P1 to P3
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27 305 porphyry intrusions have relatively homogeneous major element compositions, but have
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29 306 moderate to high loss-on-ignition values (LOI: 2.3 to 10.9%) reflecting varying degrees
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31 307 of potassic and sericitic alteration. On an LOI-free basis, they are mostly intermediate
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33 308 composition (SiO₂ = 56.71–63.16 wt. %; Digital Appendix Table A3), and straddle the
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35 309 boundary between granodiorite (diorite) and syenite on a Zr/Ti versus Nb/Y diagram (Fig.
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37 310 8). The late basaltic–andesitic dikes have distinct compositions, and plot in the field of
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39 311 diorite and gabbro on Figure 8. On Harker-type diagrams of SiO₂ versus selected major
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41 312 elements, the alkali elements (K₂O and Na₂O; Figs. 9A–B) show significant scatter,
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43 313 likely due to hydrothermal alteration, as confirmed by a rough correlation between alkali
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45 314 contents and LOI. However, other elements such as Fe₂O₃, MgO, TiO₂, and Al₂O₃ show
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47 315 roughly correlated trends with SiO₂ (Figs. 9E–F), suggesting a broadly cogenetic suite,
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49 316 with the exception of the late basaltic to andesitic dikes, which do not plot on the same
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51 317 trends for K₂O and TiO₂.
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4 318 On primitive mantle-normalized extended trace element and chondrite-normalized
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6 319 rare earth element (REE) diagrams (Figs. 10A–B), the porphyritic rocks show largely
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9 320 indistinguishable patterns consistent with typical subduction-related igneous rocks
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11 321 (Hawkesworth et al., 1993; Pearce, 1996), characterized by large-ion lithophile element
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13 322 (LILEs: Rb, Ba, Th, U, K) and light rare earth element enrichments (LREE), low
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15 323 concentrations of high field strength elements (HFSE: Nb, Ta, Zr, Hf, and Ti), relative
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17 324 depletions in compatible elements and middle to heavy rare earth elements (MREE,
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19 325 HREE; La/Yb = 8.9 ± 1.7 , n = 14), and flat to listric-shaped patterns from MREE to
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21 326 HREE. Such listric patterns likely reflect amphibole fractionation from hydrous magmas,
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23 327 and are an indication of magma fertility for porphyry ore formation (Richards and
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25 328 Kerrich, 2007). Most samples have flat or slightly positive Eu anomalies with Eu_N/Eu_N^*
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27 329 of 1.1 ± 0.15 (n = 14; Fig. 10B), also likely reflecting amphibole fractionation and lack of
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29 330 plagioclase fractionation.

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36 331 Three samples of late basaltic to andesitic dikes show distinct trace element patterns,
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38 332 especially for HFSE and REE, which are enriched relative to the porphyries; they are
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40 333 clearly not cogenetic with the earlier porphyries.

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43 334 Excluding two samples with high Sr values that may reflect calcite veining, the
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45 335 porphyries have relatively high Sr/Y (53 ± 23 , n = 12) and V/Sc ratios (9.0 ± 0.9 , n = 14),
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47 336 which overlap the range of fertile rocks for porphyry formation (Richards and Kerrich,
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49 337 2007; Loucks, 2013, 2014).

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55 339 *Whole-rock Nd-Sr isotopes*
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4 340 Nd and Sr isotopic compositions of the porphyritic rocks are very homogeneous (Table
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6 341 4 and Fig. 11). They show a narrow range of initial Sr isotopic ratios from 0.7042 to
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8 342 0.7047 ($(^{87}\text{Sr}/^{86}\text{Sr})_t = 0.7044 \pm 0.0002$, $n = 8$), with relatively high $\epsilon_{\text{Nd}}(t)$ values from 2.4
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10 343 to 3.6 (average = $+3.1 \pm 0.4$, $n = 8$), consistent with a mantle source with minor crustal
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12 344 contamination. The single-stage Nd model ages (T_{DM1}) calculated following the methods
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14 345 of Goldstein et al. (1984) range from 0.74 to 0.88 Ga (average = 0.80 ± 0.05 Ga, $n = 8$),
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16 346 and are indistinguishable for the different porphyry phases. Two basaltic–andesitic dike
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18 347 samples show slightly lower Sr isotopic ratios ($(^{87}\text{Sr}/^{86}\text{Sr})_t = 0.7042$ and 0.7043) and
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20 348 higher $\epsilon_{\text{Nd}}(t)$ ratios ($+3.5$ and $+3.9$) than the porphyry rocks (Fig. 11).
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27 *Zircon Hf isotopes*

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29 350 Zircon Hf isotopic results are listed in Digital Appendix Table A4, and illustrated in
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31 351 Figure 12. The four zircon samples from porphyritic rocks show indistinguishable Hf
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33 352 isotopic compositions and single-stage Hf model ages of 375 ± 52 Ma (range = 264–527
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35 353 Ma, $n = 56$), with relatively high $\epsilon_{\text{Hf}}(t)$ values of 12.0 ± 0.4 (weighted mean value, 95%
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37 354 confidence level; range = 8.1–14.8, $n = 56$). These data suggest that the porphyries shared
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39 355 a common primitive mantle source, consistent with their island arc origin.
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46 **Amphibole, Plagioclase, Zircon, and Apatite Compositions**

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48 358 The compositions of amphibole, plagioclase, zircon, and apatite from samples of least-
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50 359 altered porphyritic rocks are listed in Digital Appendix Tables A5, A6, A7, and A8,
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52 360 respectively. Based on these analyses, water and sulfur contents as well as oxygen
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54 361 fugacity of the magma have been estimated qualitatively and quantitatively.
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4 363 *Amphibole compositions*
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6 364 Amphibole compositions can be used to estimate magmatic oxidation state,
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9 365 crystallization temperatures and pressures, as well as water contents (e.g., Ridolfi et al.,
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11 366 2010; Zhang et al., 2012; Wang et al., 2014b). However, fresh amphibole grains were
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14 367 only observed in the P2E and P2L porphyries. All of the analyzed grains are calcic
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16 368 amphiboles, and two generations of amphibole were recognized in both P2E and P2L.
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18 369 Early stage grains occur as phenocrysts and are characterized by higher Al contents
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21 370 (range = 9.35–12.86 and 8.79–13.46 wt. % Al₂O₃ for P2E and P2L, respectively; Digital
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23 371 Appendix Table A5); they are mostly classified as magnesio-ferri-hornblende and
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25 372 magnesio-hastingsite (including potassic-magnesio-hastingsite; Fig. 13A). The late stage
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27 373 grains were developed in the groundmass or as recrystallized phenocrysts, typified by
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29 374 lower Al concentrations (range = 1.62–4.76 and 4.33–7.53 wt. % Al₂O₃ for P2E and P2L,
30
31 375 respectively; Digital Appendix Table A5). They are classified as magnesio-ferri-
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33 376 hornblende or rare actinolite (Fig. 13A). The early stage amphibole grains are intergrown
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35 377 with plagioclase phenocrysts (Fig. 14), and igneous apatite crystals occur as inclusions
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37 378 within amphibole (Fig. 14B). This indicates that these three minerals are broadly coeval.

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43 379 The crystallization temperature, magmatic water contents, and oxygen fugacities were
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45 380 estimated for the two stages of amphiboles using the spreadsheet of Ridolfi et al. (2010),
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47 381 and the crystallization pressures were calculated following the equation of Mutch et al.
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49 382 (2016). The Al-in-hornblende geobarometer used is applicable to granitoids that contain
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51 383 amphibole, plagioclase, quartz, and alkali feldspar, broadly consistent with the mineral
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53 384 assemblage within P2 porphyry rocks. The calculated results are listed in Digital
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55 385 Appendix Table A5 and illustrated in Figure 13 (B–C).
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4 386 Amphibole phenocrysts from the P2E and P2L porphyry samples have similar
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6 387 compositions, with calculated crystallization pressures from 3.5 to 6.7 kbar (average =
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8 388 4.6 ± 0.8 kbar, $n = 51$), temperatures from 855 to 983 °C (average = 900 ± 30 °C, $n = 51$),
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10 389 Δ FMQ values from 0.5 to 1.8 (average = 1.1 ± 0.3 , $n = 51$), and H₂O contents in melts
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12 390 from 4.0 to 6.1 wt. % (average = 5.1 ± 0.4 wt. %, $n = 51$).
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16 391 Late stage (low-Al) amphiboles in P2E porphyry samples yielded crystallization
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18 392 pressures from 0.6 to 1.4 kbar (average = 1.1 ± 0.3 kbar, $n = 12$), temperatures from 637°
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20 393 to 774 °C (average = 719 ± 34 °C, $n = 12$), Δ FMQ values from 2.6 to 3.3 (average = $3 \pm$
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22 394 0.2 , $n = 12$), and magmatic water contents from 3.6 to 5.1 wt. % (average = 4 ± 0.4 Ma, n
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24 395 = 12). Low-Al amphibole grains from P2L samples yielded slightly higher crystallization
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26 396 pressures (2.0 ± 0.4 MPa; range = 1.3–2.6 kbar, $n = 8$) and temperatures (800 ± 24 °C;
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28 397 range = 751–800 °C, $n = 8$), slightly lower Δ FMQ values (2.4 ± 0.3 ; range = 1.1–2.9, $n =$
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30 398 8), but similar H₂O contents (4.0 ± 0.2 ; range = 3.7–4.4, $n = 8$) to those calculated for
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32 399 P2E.
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39 400 There are clear trends of increasing oxidation state and decreasing crystallization
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41 401 pressures, temperatures, and magmatic water contents from early to late stage amphiboles
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43 402 in both P2E and P2L porphyries (Figs. 13B–D). These trends are consistent with the
44
45 403 high-Al amphibole phenocrysts having crystallized at depth before final crystallization of
46
47 404 the magma (and low-Al amphibole) after emplacement at shallow levels (e.g., Rutherford
48
49 405 and Devine, 2003). The decreasing magmatic water contents might be attributed to
50
51 406 degassing during ascent and crystallization, which can also result in a small increase in
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53 407 oxygen fugacity (Mathez, 1984; Candela, 1986; Burgisser and Scaillet, 2007; Zimmer et
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55 408 al., 2010; Bell and Simon, 2011; Dilles et al., 2015).
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6 410 *Plagioclase compositions*

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9 411 Plagioclase phenocrysts from all P1 porphyry samples studied have been altered, but
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11 412 partially unaltered phenocrysts were found in some P2 and P3 porphyry samples.

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14 413 Analyses with $K_2O > 1$ wt. % are interpreted to reflect potassic alteration and have been
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16 414 excluded. Plagioclase grains from the P2E porphyry (sample RC13-13) are classified as
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18 415 oligoclase ($\geq An_{17}$) to labradorite ($\leq An_{70}$), with an average anorthite (An) content of $44 \pm$
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20 416 12 mol % ($n = 28$). Nearly all these crystals exhibited reverse zoning in An content, with
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22 417 compositional ranges up to ~ 16 mol %, and FeO contents that correlate positively with
23
24 418 X_{An} values (Digital Appendix Table A6; Fig. 14A). Plagioclase grains from the P2L
25
26 419 porphyry (samples RC13-26 and RC13-33) have compositions similar to P2E (An mol %
27
28 420 $= 48 \pm 5$, $n = 38$), and also show reverse zoning with amplitudes up to ~ 19 mol % An that
29
30 421 correlate with FeO contents (Fig. 14B). Such reverse zoning with positive An-FeO
31
32 422 correlations were observed both in partially altered (i.e., sericite alteration; Fig. 14A) and
33
34 423 relatively fresh plagioclase crystals (Fig. 14B), indicating that it was not likely to have
35
36 424 been generated by hydrothermal alterations. In contrast, plagioclase grains from the P3
37
38 425 porphyry (RC13-78) display relatively uniform An compositions (An mol % $= 51 \pm 5$, n
39
40 426 $= 37$), with no clear zoning in X_{An} and FeO (Fig. 14C).

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42
43 427 Plagioclase compositions (An proportions) are very sensitive to dissolved water
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45 428 content and temperature of the melt, and thus can be used as a hygrometer if corrected for
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47 429 temperature (Mathez, 1973; Lange, 2009; Waters and Lange, 2015). We used apatite
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49 430 saturation temperatures (AST) following the formula of Piccoli and Candela (1994) to
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51 431 calculate magma temperatures for P2 and P3 samples (Table 4; Digital Appendix Table
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4 432 A3). The whole-rock SiO₂ and P₂O₅ values were assumed to approximate the melt
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6 433 contents at the time of crystallization, although this assumption can be challenged
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9 434 (Piccoli and Candela, 1994, 2002). Nevertheless, the calculated temperatures are
10
11 435 reasonable for magmas of intermediate, hydrous composition, and are also in agreement
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14 436 with the estimated crystallization temperatures of early stage amphibole (Digital
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16 437 Appendix Table A5): 932 °C for P2E (RC13-13), 892 °C (RC13-26) and 905 °C (RC13-
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18 438 33) for P2L, and 928 °C for P3 (RC13-78; Table 4).

21 439 The updated spreadsheet of Waters and Lange (2015) has been used to calculate
22
23 440 magmatic water contents (Digital Appendix Table A9). The calculated values are
24
25 441 relatively uniform and indistinguishable between the different porphyry phases: 5.2 ± 0.2
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27 442 ($n = 28$) for P2E, 5.5 ± 0.2 ($n = 38$) for P2L, and 5.1 ± 0.1 wt. % ($n = 37$) for P3. For the
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29 443 P2E and P2L porphyries, the magmatic water contents calculated from plagioclase and
30
31 444 early stage amphibole compositions agree well with each other (Digital Appendix Tables
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33 445 A5 and A8).

37 446

40 447 *Zircon trace element compositions*

43 448 Twenty trace element spot analyses were obtained for zircons from samples of P1 to
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45 449 P3 porphyries (Digital Appendix Table A7; Fig. 15). It is common to encounter small
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47 450 mineral inclusions, especially of apatite, titanite, in zircon during LA-ICP-MS analyses
48
49 451 (e.g., Lu et al., 2016). We have taken Ca > 200 ppm or La > 0.3 ppm as an indication of
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51 452 apatite contamination, and Ti > 20 ppm to reflect titanite contamination, and such data
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53 453 were excluded (Digital Appendix Table A7). The remaining analyses have low LREE
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55 454 and elevated HREE contents, with small negative Eu and strongly positive Ce anomalies
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4 455 (Fig. 15). These REE patterns and total contents ($\Sigma\text{REE} = 351\text{--}1125$ ppm), as well as
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6 456 Th/U ratios above 0.1 (0.26–0.63) (Digital Appendix Table A7) are typical of igneous
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9 457 zircons from relatively oxidized magmas (Hoskin and Schaltegger, 2003).

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11 458 Zircon crystals from the P1 to P3 porphyries show slightly different calculated Ti-in-
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13
14 459 zircon temperatures and Eu anomalies ($\text{Eu}_\text{N}/\text{Eu}_\text{N}^* = \text{Eu}_\text{N}/(\text{Sm}_\text{N} \times \text{Gd}_\text{N})^{0.5}$; Fig. 16; Digital
15
16 460 Appendix Table A7). Titanium-in-zircon temperatures were calculated using the equation
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18
19 461 of Ferry and Watson (2007), where it is assumed that $\log a_{\text{SiO}_2} = 1$ because of the
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21 462 existence of quartz in the host porphyries, and $\log a_{\text{TiO}_2} = 0.7$ due to the presence of
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24 463 titanite (a common accessory mineral in these rocks). Zircon crystals from the P2E
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26 464 porphyry (sample RC13-107) have the highest calculated temperatures (average of $760 \pm$
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28 465 29 °C; range = $723^\circ\text{--}819^\circ\text{C}$, $n = 14$) with relatively uniform $\text{Eu}_\text{N}/\text{Eu}_\text{N}^*$ values from 0.5 to
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31 466 0.7 (average = 0.6 ± 0.1 ; $n = 14$). Zircon crystals from the P3 porphyry (sample RC13-78)
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33
34 467 show the lowest calculated temperatures (average = 726 ± 24 °C; range = $693^\circ\text{--}758^\circ\text{C}$, n
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36 468 = 18) and lowest $\text{Eu}_\text{N}/\text{Eu}_\text{N}^*$ ratios (0.5 ± 0.1 ; range = 0.3–0.6, $n = 18$). Zircon crystals
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38 469 from the P1 and P2L porphyries have intermediate calculated temperatures (P1: $736 \pm$
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41 470 26 °C, range = $699^\circ\text{--}810^\circ\text{C}$, $n = 16$; P2L: 730 ± 24 °C, range = $705^\circ\text{--}795^\circ\text{C}$, $n = 17$),
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43
44 471 and a similar wide range of $\text{Eu}_\text{N}/\text{Eu}_\text{N}^*$ values (P1: 0.3–1.0, average = 0.6 ± 0.2 ; P2L: 0.3–
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46 472 0.9, average = 0.6 ± 0.1).

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48 473 Trace element compositions in zircon are sensitive to magmatic water content and
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51 474 oxygen fugacity, and have been used to distinguish between fertile and infertile suites in
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53 475 porphyry Cu \pm Au \pm Mo systems (Ballard et al., 2002; Liang et al., 2006; Qiu et al., 2014;
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55 476 Wang et al., 2014b; Dilles et al., 2015; Shen et al., 2015; Lu et al., 2016; Xu et al., 2016;
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57
58 477 Loader et al., 2017). In hydrous magmas, early plagioclase crystallization is suppressed
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4 478 (Moore and Carmichael, 1998), leading to no pronounced depletion of Eu in melts (Eu^{2+}
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6 479 substitutes for Ca^{2+} in plagioclase). On the other hand, in oxidized magmas, Eu is
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8
9 480 predominantly present as Eu^{3+} and is partitioned into zircon along with other REE
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11 481 (Ballard et al., 2002; Hoskin and Schaltegger, 2003; Trail et al., 2011; Dilles et al., 2015).
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13
14 482 Therefore, the small negative Eu anomalies in zircon samples (as indicated by $\text{Eu}_N/\text{Eu}_N^*$
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16 483 ratios >0.3 and mostly > 0.4 ; Fig. 16A), and slightly positive anomalies in whole rocks
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18 484 (Fig. 10) indicate relatively high magmatic water contents and/or oxidation states
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20
21 485 (Ballard et al., 2002; Dilles et al., 2015; Lu et al., 2016).

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23
24 486 Ballard et al. (2002) defined the zircon $\text{Ce}^{4+}/\text{Ce}^{3+}$ ratio as an indicator of magmatic
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26 487 oxidation state. However, Dilles et al. (2015) and Lu et al. (2016) argue that these values
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28 488 are difficult to estimate accurately because the abundances of the adjacent elements, La
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31 489 and Pr, which are used as a baseline to calculate the magnitude of the Ce anomaly, are
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33 490 low and close to the analytical detection limit. Confirming this reservation, $\text{Ce}^{4+}/\text{Ce}^{3+}$
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35
36 491 ratios and $f\text{O}_2$ values calculated for our samples using the equation of Trail et al. (2011)
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38 492 yielded unrealistic and widely variable ΔFMQ values (-9 to +5).
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42 43 494 *Apatite compositions*

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45 495 The SO_3 , Cl, and F analyses of igneous apatites are listed in Digital Appendix Table
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48 496 A8 and illustrated in Figures 17 and 18. The results show that igneous apatites from the
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51 497 P2 porphyries have higher sulfur and chlorine concentrations than those from P1 and P3
52
53 498 rocks. The high SO_3 contents in apatites from the P2 porphyries (P2E: 0.11 to 0.8 wt. %
54
55 499 SO_3 , average = 0.30 ± 0.13 wt. %, $n = 34$; P2L: 0.07 to 1.2 wt. % SO_3 , average = $0.29 \pm$
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57
58 500 0.18 wt. %, $n = 100$) are similar to values from global porphyry Cu deposits (Streck and
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4 501 Dilles, 1998; Imai, 2002; Li et al., 2012; Pan et al., 2016; Richards et al., 2017). In
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6 502 comparison, apatite crystals from the P1 porphyry (RC13-39) have lower sulfur contents
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8 503 (0.05 to 0.19 wt. % SO₃; average = 0.11 ± 0.03 wt. %, n = 34), whereas those from the P3
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10 504 porphyry (RC13-78) have the lowest sulfur concentrations (from 0.05 wt. % to below the
11
12 505 analytical detection limit of SO₃; average of analyses above limit of detection = 0.03 ±
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14 506 0.01, n = 13). Although the ranges for P1 and P2L porphyries overlap at their outer limits,
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16 507 a t-test shows that the two populations are significantly different (t = 5.87, p = 0).

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19 508 Similarly, the Cl contents in apatite crystals from P1 and P3 porphyries are
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21 509 significantly lower (0.51 ± 0.3 wt. %, n = 34, and 0.1 ± 0.06 wt. %, n = 17, respectively)
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23 510 compared to values from the P2 porphyries (P2E: 1.18 ± 0.37 wt. %, n = 34; P2L: 1.47 ±
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25 511 0.28 wt. %, n = 100). There is a rough positive correlation between sulfur and chlorine
26
27 512 contents of apatites from the four porphyries (Fig. 18A), with relatively constant molar
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29 513 S/Cl ratios (P1: 0.13 ± 0.08, n = 34; P2E: 0.13 ± 0.08, n = 34; P2L: 0.10 ± 0.14, n = 100;
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31 514 P3: 0.16 ± 0.09, n = 13; Table 4). Detailed analyses also show that single apatite crystals
32
33 515 are commonly zoned in sulfur, with decreasing core-to-rim SO₃ contents (and small
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35 516 decreases in Cl; Fig. 17C), possibly reflecting progressive degassing of SO₂ (and Cl)
36
37 517 from the melt during crystallization (Richards et al., 2017).

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39 518 Fluorine contents in apatites from the P1 and P3 porphyries are similar (P1: 3.64 ±
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41 519 0.48 wt. %, n = 34; P3: 3.6 ± 0.4, n = 17) and distinctly higher than in the P2E (2.33 ±
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43 520 0.29 wt. %, n = 34) and P2L (2.32 ± 0.23 wt. %, n = 100) porphyries.

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45 521 Some hydrothermal apatite grains were also analyzed for comparison with igneous
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47 522 grains, and they show wide variations in sulfur and chlorine content (Digital Appendix
48
49 523 Table A8), as observed by other researchers (e.g., Li et al., 2012; Richards et al., 2017).

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4 524 The sulfur partition coefficient between apatite and melt is jointly controlled by
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6 525 temperature, oxygen fugacity, and the S content in the silicate melt (Peng et al., 1997;
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9 526 Parat and Holtz, 2005; Parat et al., 2011; Webster and Piccoli, 2015; Konecke et al.,
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11 527 2017a, b), and no formula currently exists to accurately calculate magmatic sulfur
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14 528 concentrations from apatite SO₃ contents. However, two semi-quantitative formulae can
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16 529 be used to estimate relative magmatic S content (Peng et al., 1997; Parat et al., 2011).
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19 530 Using the equation of Peng et al. (1997), the magmas related to the P2E and P2L
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21 531 porphyry intrusions are calculated to have sulfur concentrations of 0.03 ± 0.01 wt. % (n =
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23 532 34) and 0.02 ± 0.01 wt. % (n = 100), significantly higher than the corresponding S
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26 533 contents for P1 (0.010 ± 0.003 wt. % S, n = 34) and P3 magmas (0.003 ± 0.001 wt. % S,
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28 534 n = 13; Table 4). The results calculated by the method of Parat et al. (2011) are more
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31 535 variable, and yielded lower absolute values than those from Peng et al. (1997). However,
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33 536 they also support the conclusion that the P2E and P2L magmas had significantly higher
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36 537 sulfur contents than the P1 and P3 magmas (Table 4).

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38 538 The chlorine partition coefficient between apatite and melt is a complex function of
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41 539 magma composition and Cl concentration in the melt (Zhu and Sverjensky, 1991; Piccolli
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43 540 and Candela, 1994; Mathez and Webster, 2005; Webster et al., 2009; Chelle-Michou and
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46 541 Chiaradia, 2017). Mathez and Webster (2005) proposed a value of 0.8 as the partition
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48 542 coefficient (mass ratio) between apatite and basaltic melt (51.1 wt. % SiO₂). Based on
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51 543 this semi-quantitative equation (and recognizing that these magmas were not basaltic), we
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53 544 estimate Cl concentrations in the P2E and P2L magmas of 1.48 ± 0.46 wt.% (n = 34) and
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55 545 1.83 ± 0.34 wt.% (n = 100), respectively, significantly higher than for P1 (0.63 ± 0.38
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58 546 wt.%, n = 34) and P3 (0.12 ± 0.07 wt.%; Table 4).

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548 **Discussion**

549 *Magmatic duration and timing of ore formation at Red Chris*

550 Four phases of the Red Stock, P1, P2E, P2L, and P3, have been dated at 211.6 ± 1.3
551 Ma, 206.0 ± 1.2 Ma, 203.6 ± 1.8 Ma, and 201.7 ± 1.2 Ma, respectively, spanning a period
552 of ~10 m.y. (Fig. 6; Digital Appendix Table A2). These ages are consistent with
553 crosscutting relationships described by Rees et al. (2015). Re-Os dating of molybdenite
554 intergrown with chalcopyrite in quartz veins yielded a weighted mean age of 206.1 ± 0.5
555 Ma (Fig. 7), reflecting the timing of mineralization and in good agreement with the
556 emplacement age of the mineralized P2E quartz monzonite porphyry. Ore formation
557 occurred at a relatively late stage in the ~10 m.y. evolution of the Red Stock, and over a
558 relatively brief period time (<1 m.y., as defined by the uncertainty on the average of three
559 Re-Os analyses). This relatively late and short duration of mineralization is similar to the
560 timing observed in many other porphyry deposits (e.g., Cathles et al., 1997; Masterman et
561 al., 2004; von Quadt et al., 2011; Chiaradia et al., 2013; Correa et al., 2016; Li et al.,
562 2017).

563 The age of mineralization at ~206 Ma is contemporaneous with the major pulse of
564 mineralization in the Stikine and Quesnel terranes, representing a particularly fertile pre-
565 accretionary magmatic event (Nelson and Colpron, 2007; Nelson et al., 2013; Logan and
566 Mihalynuk, 2014).

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568 *Igneous geochemistry and petrogenesis*

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4 569 Whole-rock major and trace element compositions of P1 to P3 porphyry rocks are
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6 570 almost identical to each other, and show characteristics of subduction-related magmatic
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9 571 rocks with high LILE and low HFSE concentrations (Figs. 9 and 10; Hawkesworth et al.,
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11 572 1993; Pearce, 1996). The high Sr/Y ratios (53 ± 23 , $n = 12$) and slightly positive Eu
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13 573 anomalies ($\text{Eu}_n/\text{Eu}_n^* = 1.1 \pm 0.15$; $n = 14$) indicate the suppression of plagioclase
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15 574 crystallization and early crystallization of amphibole from hydrous and oxidized magmas
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17 575 (Richards and Kerrich, 2007; Loucks, 2014). Rees et al. (2015) concluded from their
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19 576 analysis of petrochemistry that the Red Stock evolved from subalkaline (P1, P2) to
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21 577 marginally alkaline (P3), and classified the overall system as high-K calc-alkalic.

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25 578 All the porphyritic rocks from P1 to P3 show mantle-like, relatively low ($^{87}\text{Sr}/^{86}\text{Sr}$)_t
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27 579 (0.7044 ± 0.0002) and high $\epsilon_{\text{Nd}}(t)$ ($+3.1 \pm 0.4$) values, and similar single-stage Nd model
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29 580 ages (0.80 ± 0.05 Ga). Zircon crystals from these rocks yield positive zircon $\epsilon_{\text{Hf}}(t)$ values
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31 581 ($+8.1$ to $+14.8$) and single-stage Hf model ages of 375 ± 52 Ma (Figs. 11 and 12). These
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33 582 data indicate minimal involvement of ancient crustal components in the petrogenesis of
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35 583 these magmas, as observed for other Mesozoic igneous rocks in the Stikine terrane, and
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37 584 are consistent with an oceanic island arc setting (Samson et al., 1989; Logan and
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39 585 Mihalynuk, 2014).

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43 586 Late basaltic to andesitic dikes have similar Nd–Sr isotopic compositions to the
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45 587 porphyritic rocks, but show distinct trace element and REE patterns, and lower Sr/Y and
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47 588 La/Yb ratios, suggesting that they are not cogenetic.

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55 590 *Oxygen fugacity and sulfur-chlorine-water contents of the porphyry magmas*
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4 591 Magmatic oxidation state, and sulfur, chlorine, and water contents are fundamental
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6 592 factors for the formation of porphyry Cu deposits (e.g., Burnham, 1979; Candela, 1992;
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9 593 Richards, 2003; Loucks, 2014). It is difficult to measure the original magmatic oxidation
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11 594 state in altered plutonic rocks, but normalized Ce and Eu anomalies in zircons can be
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14 595 used to provide qualitative estimates (Ballard et al., 2002; Dilles et al., 2015; Lu et al.,
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16 596 2016). Zircon grains from the P1 to P3 porphyries display large positive Ce and small
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19 597 negative Eu anomalies (Eu_N/Eu_N^* mostly > 0.4 ; Figs. 15 and 16), similar to zircons from
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21 598 other mineralized porphyry systems worldwide (Wang et al., 2014b; Dilles et al., 2015;
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24 599 Lu et al., 2016). These data are also consistent with estimates of magmatic oxidation state
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26 600 from early stage amphibole phenocrysts from the P2E and P2L porphyries, which yielded
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29 601 ΔFMQ values of 1.5 ± 0.2 ($n = 7$) and 1.1 ± 0.3 ($n = 44$). Consequently, we conclude that
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31 602 the four phases of magma in the Red Stock were all at least moderately oxidized.

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34 603 It is similarly difficult to estimate original magmatic water contents from altered
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36 604 plutonic rocks. However, the presence of amphibole phenocrysts in the P1 to P3
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39 605 porphyries suggest that the magmas contained at least 4 wt. % H_2O (Burnham, 1979,
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41 606 Naney, 1983; Merzbacher and Egglar, 1984; Rutherford and Devine, 1988; Ridolfi et al.,
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43
44 607 2010). This observation is supported by: (1) high whole-rock Sr/Y ratios (53 ± 23 , $n =$
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46 608 12), which reflect the suppression of plagioclase relative to amphibole crystallization
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49 609 under hydrous conditions (Moore and Carmichael, 1998; Richards and Kerrich, 2007); (2)
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51 610 high $10000 \times (Eu_N/Eu_N^*)/Y$ (> 1) and low Dy/Yb (< 0.3) ratios in zircons (Digital Appendix
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54 611 Table A7; Lu et al., 2016); (3) high whole-rock V/Sc ratios (9.0 ± 0.9 , $n = 14$) (Loucks,
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56 612 2014); and (4) estimates of magmatic water content from plagioclase phenocryst
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58 613 compositions, which indicate values of ~ 5 wt.% H_2O .

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4 614 The compositions of igneous apatite inclusions in plagioclase and amphibole
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6 615 phenocrysts have been used to estimate magmatic sulfur and chlorine contents. The
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9 616 results suggest that apatite microphenocrysts from the P2E and P2L porphyries had
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11 617 higher S and Cl contents than for P1 and P3 (Fig. 18). The SO₃ contents in apatite are
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14 618 controlled by both magmatic sulfur concentration and oxidation state (Peng et al., 1997;
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16 619 Parat and Holtz, 2005; Webster and Piccoli, 2015). Sulfur occurs in the apatite structure
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19 620 mainly as S⁶⁺ and its content will increase in relatively oxidized environments (Boyce et
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21 621 al., 2010; Parat et al., 2011; Konecke et al., 2017b). The magmas associated with the four
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24 622 porphyry phases at Red Chris were all moderately oxidized, and therefore the differences
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26 623 in apatite sulfur content likely mainly reflect variations in the sulfur fugacity in the
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29 624 associated melts. The data indicate that the P2E and P2L porphyry magmas contained
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31 625 significantly higher S contents than the P1 and P3 magmas (Fig. 18). Models used to
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33 626 estimate magmatic sulfur content in equilibrium with apatite (Peng et al., 1997) are not
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36 627 strictly accurate because of uncertainties in the details of exchange reactions (Streck and
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38 628 Dilles, 1998; Mao et al., 2016) but are expected to be correct in relative terms. Our data
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41 629 yield higher values for the P2E and P2L porphyries (0.02–0.03 wt. % S) compared to
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43 630 estimates for P1 (0.01 wt. % S) and P3 (0.003 wt. % S). Consequently, we conclude that
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46 631 the syn-mineral porphyry (P2E and P2L) magmas were more S-rich than the pre-
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48 632 mineralization P1 and especially the post-mineralization P3 porphyry magmas.

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50 633 Similarly, the magmatic Cl contents calculated from apatite compositions in the P2E
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52 634 and P2L porphyries are also higher than in the P1 and P3 porphyries, whereas F contents
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55 635 are lower (Table 4). Volatile species such as S, Cl, and F are differentially affected by
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58 636 degassing during ascent and crystallization of magmas: S and Cl will be preferentially

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4 637 lost to the vapor phase relative to F (Webster et al., 2014; Stock et al., 2016), leading to
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6 638 high F/Cl and F/S ratios in degassed magmas. The data shown in Figure 18 suggest that
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9 639 the P2E and P2L were S-Cl-rich, whereas the P1 and P3 were S-Cl-poor but enriched in F.

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11 640 These differences could reflect either loss of volatiles from the source magma
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13 641 chamber (presumed to be an underlying batholith) prior to emplacement of the P1 and P3
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15 642 magmas but not in the case of P2, or conversely addition of a volatile phase or a volatile-
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17 643 rich magma to the magma chamber immediately prior to emplacement of P2. Volatile
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19 644 loss is expected to be a general condition of upper crustal magma emplacement, and this
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21 645 may well be a factor contributing to the low Cl/F ratios in the late P3 magmas, as well as
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23 646 the negatively correlated trend for F and Cl in P1. The P2 magma was emplaced after P1,
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25 647 and so might be expected to be similarly degassed if this represents a comagmatic
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27 648 sequence. However, the 6–8 m.y. age difference between P1 (211.6 ± 1.3 Ma) and P2
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29 649 (P2E: 206.0 ± 1.2 Ma; P2L: 203.6 ± 1.8 Ma) indicates that P1 and P2 are not directly
30
31 650 comagmatic. One explanation is that the P2 event represents a pulse of S-Cl-rich magma
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33 651 injected at a relatively late stage into the mid-crustal magma reservoir, leading to a pulse
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35 652 of fluid exsolution and injection of P2 magma into the shallower level Red Stock, where
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37 653 syn-P2, quartz-vein hosted mineralization formed the Red Chris Cu-Au deposit. The
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39 654 post-mineralization P3 (201.7 ± 1.2 Ma) and P4 porphyries may represent the final stages
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41 655 of intrusive activity emanating from the now degassed underlying magma chamber.

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43 656 Although both P2E and P2L porphyries are S-Cl-rich, apatites from P2L are
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45 657 marginally the most enriched. Samples of P2L are also the most mafic (SiO_2 -poor) in the
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47 658 suite, and it is therefore tempting to speculate that the source magma chamber was
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49 659 recharged with more mafic, S-Cl- H_2O -rich magma during the P2 stage. The solubility of
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4 660 S and Cl is higher in oxidized mafic magmas compared to felsic melts (> 1000 ppm S
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6 661 and > 3000 ppm Cl; Carmichael and Ghiorso, 1986; Webster, 1997; Webster et al., 1999;
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9 662 Hattori and Keith, 2001; Jugo et al., 2005; Jugo, 2009; Chelle-Michou and Chiaradia,
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11 663 2017). Therefore, a key step in porphyry ore formation at Red Chris may have been
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14 664 recharge of the batholithic system by a less evolved, although broadly cogenetic, magma
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16 665 (cf. Hattori and Keith, 2001; Steinberger et al., 2013; Large et al., 2018).

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19 666 Support for this recharge model is provided by high magmatic temperatures
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21 667 calculated from zircon grains from some P2 samples (Fig. 16), and reverse zoning
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24 668 observed in plagioclase phenocrysts from the P2E and P2L porphyries, but which is not
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26 669 observed in P3 (Fig. 14C). In detail, zircons from the P2E porphyry show the highest Ti-
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28 670 in-zircon temperatures and Eu_N/Eu_N^* values, and P3 the lowest values. This might reflect
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31 671 an influx of higher temperature melt during the evolution of the P2E magma, and greater
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33 672 degrees of plagioclase fractionation from the late P3 magma. The compositional ranges
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36 673 of up to ~19 mol % An for plagioclase are higher than the upper limit caused by chemical
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38 674 diffusion (~10 mol % An; Pearce and Kolisnik, 1990). Furthermore, these reverse zoning
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41 675 patterns are matched by FeO contents in the phenocrysts (Figs. 14A and 14B), which
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43 676 suggests that the zonation reflects changes in magma composition (as opposed to simply
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46 677 changes in pressure, temperature, or water content; Ginibre and Wörner, 2007; Lange,
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48 678 2009; Ustunisik et al., 2014; Waters and Lange, 2015). We therefore interpret these
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51 679 changes to reflect late-stage (rim) growth from a more mafic magma that mixed into a
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53 680 more evolved resident magma.

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Conclusions

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4 683 Zircon U-Pb ages indicate that the Red Stock was emplaced over a period of ~10 m.y.
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6 684 (211.6 ± 1.3 Ma, 206.0 ± 1.2 Ma, 203.6 ± 1.8 Ma, and 201.7 ± 1.2 Ma for the P1, P2E,
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8 685 P2L, and P3 porphyries, respectively). The ore-forming event at Red Chris occurred
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10 686 relatively late in this magmatic history, synchronous with P2E at 206.1 ± 0.5 Ma, and
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12 687 over a short period of <1 m.y. (as constrained by the error on the average of three Re-Os
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14 688 analyses). The four phases of porphyry and their related magmas were chemically fairly
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16 689 homogeneous and isotopically primitive, consistent with an island arc origin. The
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18 690 magmas were all moderately oxidized and hydrous (~5 wt. H₂O), but the P2 magmas had
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20 691 distinctly higher S and Cl contents. Combined with evidence for reverse zonation in
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22 692 plagioclase phenocrysts from P2 porphyries, and more mafic compositions in P2L, we
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24 693 suggest that the deeper source magma chamber was recharged at the time of
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26 694 emplacement of the P2 porphyries by injection of a more mafic S-Cl-rich magma. This
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28 695 recharge process may have triggered ore formation by causing voluminous exsolution of
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30 696 metalliferous hydrothermal fluids.
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18 19 712 **References**

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1195 **Figure Captions**

1196 Fig. 1. Major terranes in the south of the Canadian Cordillera, showing Triassic to
1197 Jurassic magmatic belts and major associated porphyry deposits in the Stikine and
1198 Quesnel terranes (modified from Nelson and Colpron, 2007; Nelson et al., 2013; original
1199 graphics file provided by Joanne Nelson, British Columbia Geological Survey of Canada).
1200 The age of the Red Chris deposit is from this study (Table 1), and the ages for the other
1201 porphyry deposits are from Brown and Kahlert (1986; Red Mountain), Mortensen et al.
1202 (1995; Mt. Polley), Scott et al. (2008; Schaft Creek), Duuring et al. (2009; Kemess
1203 South), Taseko Mines Limited (2013; Gibraltar), Bath et al. (2014; Lorraine), Byrne and
1204 Tosdal (2014; Galore Creek), Devine et al. (2014; Lorraine), Logan and Mihalynuk (2014;
1205 Highland Valley, Copper Mountaine, Afton/Ajax, and Brenda), and Jago et al. (2014; Mt.
1206 Milligan).

1207
1208 Fig. 2. Simplified geological map of the Red Stock and Red Chris Cu-Au deposit,
1209 showing the main mineralized zoned (named) and the locations of sampled drill holes
1210 (modified from Rees et al., 2015). Universal Transverse Mercator coordinates are based
1211 on the WGS84 datum.

1212
1213 Fig. 3. Hand specimen photographs of samples of the P1, P2E, P2L, and P3 porphyry
1214 intrusions, and two late basaltic to andesitic dikes. (A) P1 porphyry with anhedral to
1215 subhedral altered amphibole and plagioclase phenocrysts, crosscut by pyrite-quartz veins
1216 (sample RC13-35). (B) P2E porphyry showing crowded texture with chloritized
1217 amphibole and plagioclase phenocrysts; plagioclase grains are rimmed or replaced by

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1218 secondary K-feldspar. The brick-red color is due to fine-grained hematite (sample RC13-
1219 107). (C) P2L porphyry composed of abundant fresh amphibole and plagioclase
1220 phenocrysts, with quartz in the groundmass (sample RC13-32). (D) The P3 porphyry is
1221 similar to P2L, but is distinguished by the absence of quartz in the groundmass (sample
1222 RC13-78). (E) Andesitic dike with anhedral to subhedral amphibole phenocrysts, crosscut
1223 by a small calcite vein (sample RC13-62). (F) Basaltic dike with chloritized amphibole
1224 phenocrysts (sample RC13-106). See Digital Appendix Table A1 for sample locations.

1225
1226 Fig. 4. West-southwest–east-northeast cross section A–A’, and north-northwest–south-
1227 southwest cross section B–B’ (location of sections shown in Fig. 2), modified from
1228 Gillstrom et al. (2012) and Rees et al. (2015). Copper equivalent-grade zones, drill holes,
1229 and the boundary between potassic and post-potassic zones are shown: Cu equivalent (%)
1230 = $\text{Cu (\%)} + 0.486 \times \text{Au (g/t)}$.

1231
1232 Fig. 5. Hand specimen photographs and photomicrographs of alteration and vein minerals.
1233 (A, B) P2E porphyry (samples RC13-81 and RC13-75) with potassic alteration and A-
1234 type quartz veins comprising magnetite, secondary K-feldspar, and disseminated bornite
1235 and pyrite. Potassic alteration is characterized by secondary K-feldspar veins and
1236 selvages around A-veins. Late unmineralized carbonate veins cut the A veins. (C)
1237 Amphibole phenocryst altered to secondary biotite, which has then been altered to
1238 chlorite, reflecting potassic alteration overprinted by chlorite–sericite alteration (plane-
1239 polarized transmitted light; P2E: sample RC13-30). (D) Plagioclase phenocryst partially
1240 overprinted by sericite (cross-polarized transmitted light; P2E porphyry: sample RC13-

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1241 11). (E) B-type quartz veins with centerline pyrite in P2L porphyry (sample RC13-44). (F)
1242 Pyritic D vein in P1 porphyry (sample RC13-40). (G) High-grade ore in sheeted A-type
1243 quartz-chalcopyrite-K-feldspar veins (sample RC13-31 in P2E porphyry). (H) Quartz-
1244 carbonate-pyrite-molybdenite-chalcopyrite vein in P2E porphyry (reflected light; sample
1245 RC13-88). Abbreviations: Amp = amphibole; Bi = biotite; Bn = bornite; Cbn = carbonate;
1246 Chl = chlorite; Cpy = chalcopyrite; Kfs = K-feldspar; Mo = molybdenite; Mt = magnetite;
1247 Pl = plagioclase; Py = pyrite; Qtz = quartz. See Digital Appendix Table A1 for sample
1248 locations.

1249
1250 Fig. 6. Zircon U–Pb Tera-Wasserburg concordia diagrams for (A) P1, (B) P2E, (C) P2L,
1251 and (D) P3 porphyry samples dated by LA-MC-ICP-MS. Uncertainty ellipses and
1252 calculated ages are shown at 2σ .

1253
1254 Fig. 7. Weighted mean Re-Os model age of three molybdenite vein samples from the Red
1255 Chris Cu-Au deposit.

1256
1257 Fig. 8. Zr/Ti vs. Nb/Y discrimination diagram (Winchester and Floyd, 1977) for porphyry
1258 and basaltic–andesitic dike samples from Red Chris.

1259
1260 Fig. 9. Selected whole-rock major element variation diagrams for porphyry and basaltic–
1261 andesitic dike samples from Red Chris: (A) K₂O, (B) Na₂O, (C) TiO₂, (D) Al₂O₃, (E)
1262 total Fe₂O₃, and (F) MgO vs. SiO₂.

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1264 Fig. 10. (A) Primitive mantle-normalized trace element, and (B) chondrite-normalized
1265 rare earth element diagrams for porphyry and basaltic–andesitic dike samples from Red
1266 Chris. The normalization values for primitive mantle chondrite are from Sun and
1267 McDonough (1989).

1268
1269 Fig. 11. $\epsilon_{Nd}(t)$ vs. initial $^{87}Sr/^{86}Sr$ ratios for porphyry and basaltic–andesitic dike samples
1270 from Red Chris, calculated at $t = 200$ Ma. All the samples fall in the field of Mesozoic
1271 igneous rocks in the Stikine island arc terrane, clearly different from the Late Cretaceous
1272 to Eocene plutons in the Northern Coast Plutonic Complex, which were derived from
1273 evolved crust. The depleted MORB mantle field is from Pilet et al. (2011); the Stikinia
1274 Mesozoic igneous rock field is from Samson et al. (1989); the Northern Coast Plutonic
1275 Complex field is from Samson et al. (1991) and Patchett et al. (1998); all data are re-
1276 calculated at 200 Ma.

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1278 Fig. 12. Histogram and relative probability curve for zircon $\epsilon_{Hf}(t)$ values from (A) P1, (B)
1279 P2E, (C) P2L, and (D) P3 porphyry samples.

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1281 Fig. 13. Classification diagram and plots of oxidation state, temperature, pressure, and
1282 magmatic water content estimated from amphibole compositions from P2E and P2L
1283 porphyry samples at Red Chris. A. $C(Al^{VI} + Fe^{3+} + 2Ti^{4+})$ (apfu) vs. $A(Na^{+} + K^{+})$ (apfu). B.
1284 ΔFMQ vs. temperature. C. ΔFMQ vs. pressure. D. ΔFMQ vs. magmatic water content.
1285 The classification diagram for calcic amphibole is given by the Excel spreadsheet of

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1286 Locock (2014), and the superscript C and A represent C and A cations following the
1287 general amphibole formula ($AB_2C_5T_8O_{22}W_2$), respectively. Note that potassic-magnesian-
1288 hastingsite is included in the field of magnesian-hastingsite. ΔFMQ values, temperatures,
1289 and magmatic water content were calculated from the spreadsheet of Ridolfi et al. (2010).
1290 Amphibole crystallization pressures were calculated using the equation of Mutch et al.
1291 (2016). The ΔFMQ values were calculated following the equation of Myers and Eugster
1292 (1983): $\log fO_2 = -24,441.9/T (K) + 8.290 (\pm 0.167)$. Abbreviation: apfu = atoms per
1293 formula unit.

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1295 Fig. 14. Photomicrographs (cross-polarized transmitted light), Backscattered electron
1296 (BSE) images, and electron microprobe analysis profiles for FeO and anorthite proportion
1297 (X_{An}) for representative plagioclase crystals from the P2E, P2L, and P3 porphyries at Red
1298 Chris. Red circles on photomicrographs and white circles on BSE images denote the
1299 analyzed spots. The error bars for X_{An} and FeO analyses are smaller than the size of the
1300 symbols. Abbreviations: Amp = amphibole; Ap = apatite; Pl = plagioclase; Ser = sericite.

1301

1302 Fig. 15. Chondrite-normalized REE patterns for zircons from (A) P1, (B), P2E, (C) P2L,
1303 and (D) P3 porphyry samples from Red Chris. Normalization values are from Sun and
1304 McDonough (1989).

1305

1306 Fig. 16. Zircon Eu_N/Eu_N^* vs. temperature diagram. Eu_N/Eu_N^* is the europium anomaly,
1307 calculated as $Eu_N/Eu_N^* = Eu_N/(Sm_N \times Gd_N)^{0.5}$, using the chondrite normalization values

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1308 of Sun and McDonough (1989). Oxidized suites have zircon Eu_N/Eu_N^* values >0.4
1309 (Dilles et al., 2015).

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1311 Fig. 17. Backscattered electron images of apatite crystals in samples from (A) P1 (RC13-
1312 39), (B) P2E (RC13-107), (C) P2L (RC13-33), and (D) P3 (RC13-78). Concentrations of
1313 SO_3 and Cl in apatite crystals are shown in wt. % (SO_3/Cl); red circles represent the
1314 analyzed spots. Higher concentrations are observed in apatites from P2E and P2L; some
1315 apatite microphenocrysts from P2 porphyries show zoning from SO_3 -Cl-rich cores to
1316 SO_3 -Cl-poorer rims (C).

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1318 Fig. 18. Plots of (A) S, and (B) F vs. Cl contents for apatite microphenocrysts from P1 to
1319 P3 porphyry samples at Red Chris. Abbreviation: apfu = atoms per formula unit. Data
1320 from Digital Appendix Table A8.

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4 **1 Elevated magmatic sulfur and chlorine contents in ore-forming magmas at the Red**
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6 **2 Chris porphyry Cu-Au deposit, Northern British Columbia, Canada**

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40 **14 Abstract**

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42 15 The Red Chris porphyry Cu-Au deposit is located in the Stikinia island-arc terrane in
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44 16 northwest British Columbia, ~~it~~ It is hosted by the Red Stock, which has four phases of
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46 17 porphyry intrusions: P1, P2E, P2L, and P3. New U-Pb dating of zircon shows that these
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48 18 intrusions were emplaced at 211.6 ± 1.3 Ma (MSWD = 0.85), 206.0 ± 1.2 Ma (MSWD =
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50 19 1.5), 203.6 ± 1.8 Ma (MSWD = 1.5), and 201.7 ± 1.2 Ma (MSWD = 1.05), respectively.
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52 20 The ore-forming event at Red Chris was a short-lived event at 206.1 ± 0.5 Ma (MSWD =
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54 21 0.96 ; weighted average age of three Re-Os analyses), implying a duration of <1 m.y., as
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56 22 defined by the uncertainty range. This mineralization age coincides with the emplacement
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4 23 age of the P2E porphyry, and is consistent with cross-cutting relationships that suggest
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7 24 P2E was the main syn-mineralization intrusion.

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9 25 Zircons from P1 to P3 porphyry rocks have consistently high $\text{Eu}_N/\text{Eu}_N^*$ ratios (mostly >
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11 26 0.4), indicating that their associated magmas were moderately oxidized. The magmatic
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14 27 water contents estimated from plagioclase and amphibole compositions suggest H_2O
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16 28 contents of ~5 wt. %. Taken together, the P1 to P3 porphyries are interpreted to be
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19 29 moderately oxidized and hydrous.

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21 30 The four phases of porphyries are differentiated by sulfur and chlorine contents. The
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24 31 SO_3 contents of igneous apatite microphenocrysts from the mineralization-related P2
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26 32 porphyries are higher (P2E: 0.30 ± 0.13 wt. %, $n = 34$; P2L: 0.29 ± 0.18 wt. %, $n = 100$)
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29 33 than those from the pre-mineralization P1 (0.11 ± 0.03 wt. %, $n = 34$) and post-
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31 34 mineralization P3 porphyries (0.03 ± 0.01 wt. %, $n = 13$). The chlorine contents in apatite
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33 35 grains from the P2E and P2L porphyries are 1.18 ± 0.37 ($n = 34$) and 1.47 ± 0.28 wt. %
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35 36 ($n = 100$), also higher than those from P1 (0.51 ± 0.3 wt. % Cl, $n = 34$) and P3 ($0.02 \pm$
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37 37 0.02 wt. % Cl, $n = 17$). These results imply that the sulfur and chlorine contents of the
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40 38 P2E and P2L magmas were higher than in the P1 and P3 magmas, suggesting that
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43 39 elevated magmatic S-Cl contents in the P2 porphyries may have been important for ore-
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45 40 formation. Although the process that caused the increase in sulfur and chlorine is not
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48 41 clear, reverse zoning seen in plagioclase phenocrysts from the P2 porphyry, and the
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51 42 occurrence of more mafic compositions in P2L suggest that recharge of the deeper
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53 43 magma chamber by a relatively S-Cl-rich mafic magma may have triggered the ore-
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55 44 forming hydrothermal event.

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Introduction

Compared to the relatively long-lived magmatic systems represented by a composite batholith (up to 10 m.y.; Matzel et al., 2006; Walker et al., 2007; Harris et al., 2008; Schaltegger et al., 2009; Paterson et al., 2011; Rezeau et al., 2016), the life spans of porphyry magmatic-hydrothermal ore-forming events are much shorter, probably on the order of several ~~10⁵~~-hundred thousand years (up to 1 m.y.; Cathles et al., 1997; Hayba and Ingebritsen, 1997; Masterman et al., 2004; von Quadt et al., 2011; Weis et al., 2012; Chiaradia et al., 2013; Correa et al., 2016; Li et al., 2017). In many large magmatic systems, porphyry formation occurs at a relatively late stage in the system's evolution (Creasey, 1977; Titley and Beane, 1981; Candela, 1992; Richards, 2003; Rohrlach and Loucks, 2005; Yang et al., 2009; Correa et al., 2016).

Although the multiple phases of porphyry stocks are commonly broadly cogenetic, they may be derived from packets of magma that evolved at different crustal levels over the history of the larger magmatic system (Annen et al., 2006). Understanding why ore-formation is only associated with a specific intrusive suite within these broader systems, and at discrete, commonly singular times, is a focus of this study.

It is recognized that magmas with high sulfur (>1000 ppm), chlorine (>3000 ppm), and water (>4 wt. %) contents as well as relatively high oxidation states (higher than the fayalite-magnetite-quartz buffer, ΔFMQ , by 1–2 log $f\text{O}_2$ units) are fertile for the generation of magmatic-hydrothermal porphyry Cu deposits (Burnham, 1979; Candela, 1992; Richards, 2003, 2009, 2011, 2015; Wallace, 2005; Chambefort et al., 2008; Simon and Ripley, 2011; Chiaradia et al., 2012; Loucks, 2014; Hou et al., 2015; Lu et al., 2015, 2016; Chelle-Michou and Chiaradia, 2017). These ingredients are essential for the

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4 69 transport of Cu (and Au) both in the magma (Zajacz et al., 2012) and in exsolved high
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6 70 temperature, SO₂-rich, saline magmatic fluids (Candela and Holland, 1984; Zajacz et al.,
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9 71 2011). However, whereas many magmas achieve such compositions, ore formation is a
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11 72 relatively rare and discrete event in such systems.
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14 73 Various mechanisms have been proposed that might trigger an ore forming event
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16 74 from a fertile magma source, including recharge of the magma chamber by hot, sulfur-
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18 75 rich melts (Hattori, 1993; Keith et al., 1997; Larocque et al., 2000; Hattori and Keith,
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20 76 2001; Halter et al., 2005; Stern et al., 2007; Nadeau et al., 2010, 2016; Wilkinson, 2013;
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22 77 Tapster et al., 2016; Zhang and Audétat, 2017), fluxing by sulfur gases (Blundy et al.,
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24 78 2015), or increasing the water and oxidation state of magmas during long-term
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26 79 fractionation in deep magma chambers (Ballard et al., 2002; Wang et al., 2014a, 2014b;
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28 80 Dilles et al., 2015; Lu et al., 2015, 2016). The Red Chris porphyry Cu-Au deposit
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30 81 provides an opportunity to test these hypotheses, because ore-formation occurred at a
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32 82 discrete and relatively late stage in the ~10 m.y. history of the associated magmatic
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34 83 system.
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40 84 The Red Chris deposit is located in northwest British Columbia, and contains
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42 85 measured and indicated resources of 1,035 million metric tonnes of ore grading 0.35%
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44 86 Cu and 0.35 g/t Au (Gillstrom et al., 2012). Pre-, syn-, and post-mineralization porphyry
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46 87 intrusions have been recognized based on detailed core logging and petrographic work
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48 88 (Rees et al., 2015), with mineralization occurring in a singular episode during this
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50 89 magmatic history. In this paper, we present detailed geochronological and geochemical
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52 90 data for the porphyry phases at Red Chris, and show that they have similar bulk
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54 91 compositions, including all being relatively hydrous and oxidized. However, the syn-
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92 mineralization porphyry is characterized by plagioclase with reverse zoning and apatite
93 with relatively high sulfur and chlorine contents. We interpret these data to indicate that
94 high magmatic S and Cl contents, in addition to high H₂O contents and oxidation state,
95 were critical for Cu-Au ore-formation at Red Chris. We speculate that there was an
96 injection of relatively mafic (hotter, and more S- and Cl-rich) but cogenetic magma into
97 the mid–upper crustal source magma chamber approximately coincident with
98 emplacement of the syn-mineralization porphyry.

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100 **Geological Setting**

101 Red Chris is situated within the island-arc terrane of Stikinia in the Intermontane Belt
102 of the Canadian Cordillera of British Columbia (Fig. 1; Monger and Irving, 1980; Nelson
103 and Colpron, 2007; Nelson et al., 2013). Stikinia consists primarily of Mesozoic arc-
104 related igneous and sedimentary rocks, formed in response to subduction processes prior
105 to accretion to the ancestral North American margin in the Middle Jurassic (Gabrielse,
106 1991; Ricketts et al., 1992; Lang et al., 1995; Mihalynuk et al., 2004; Nelson and Colpron,
107 2007; Logan and Mihalynuk, 2014). In northwestern Stikinia, arc assemblages comprise
108 the Middle to Late Triassic Stuhini Group, unconformably overlain by Late Triassic and
109 Early to Middle Jurassic volcanic and sedimentary rocks of the Hazelton Group (Fig. 2;
110 Brown et al., 1991; Gabrielse, 1991; Marsden and Thorkelson, 1992; Dostal et al., 1999;
111 Gagnon et al., 2012; Nelson et al., 2013; Logan and Mihalynuk, 2014). The Hazelton
112 Group is overlain by sedimentary rocks of the syn- to post-accretion Middle Jurassic to
113 Early Cretaceous Bowser Lake Group.

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4 114 Several porphyry Cu±Au±Mo deposits occur in the region, hosted by Late Triassic to
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7 115 Early Jurassic arc-related plutons (Fig. 1), including Red Chris which formed in the Late
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9 116 Triassic Red Stock. Collectively, these regional intrusions have ages ranging from ~222
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11 117 Ma to ~180 Ma, recording much of the pre-accretionary history of Stikinia (Lang et al.,
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13 118 1995; McMillan et al., 1995; Scott et al., 2008; Nelson et al., 2013; Logan and Mihalynuk,
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15 119 2014). A significant percentage of the known porphyry Cu deposits, including Red Chris,
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17 120 formed during a 6-m.y. pulse of magmatism between 206 and 200 Ma, with compositions
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19 121 ranging from calc-alkaline to strongly alkaline (Lang et al., 1995; McMillan et al., 1995;
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21 122 Nelson et al., 2013; Bissig and Cooke, 2014; Logan and Mihalynuk, 2014; Micko et al.,
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23 123 2014; Pass et al., 2014).
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125 **Geology of the Red Chris Cu-Au deposit**

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36 126 The Red Chris Cu-Au deposit was discovered in the 1950s and explored
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38 127 intermittently in subsequent decades, with mining beginning in 2015 (Ash et al., 1995,
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40 128 1996; Newell and Peatfield, 1995; Baker et al., 1997; Gillstrom et al., 2012; Rees et al.,
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42 129 2015). The deposit is hosted by the Red Stock, which is the largest of a suite of Late
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44 130 Triassic to Early Jurassic stocks and dikes that intrude the Stuhini Group in the district
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46 131 (Fig. 2; Friedman and Ash, 1997; Rees et al., 2015). The stock is tabular, elongate in an
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48 132 east to northeast direction, and approximately 6.5 km long by 300 to 1500 m wide (Fig. 2;
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50 133 Ash et al., 1995; Baker et al., 1997; Gillstrom et al., 2012). It has a steep northern contact
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52 134 against Stuhini Group country rocks, but its southeastern margin against Hazelton and
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54 135 Bowser Lake Group strata is poorly exposed, and has locally been truncated by the NE-
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4 136 trending, steeply SE-dipping South Boundary fault (Fig. 2). This fault, and the East zone
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6 137 fault within the stock, probably reflect a long-lived and deep structure which guided the
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9 138 emplacement of the intrusions, mineralization, and subsequent deformation of the Red
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11 139 Stock.

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15 140 The Red Stock is a composite intrusion consisting of several phases of porphyritic
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17 141 diorite to quartz monzonite. Mineralogically, the rocks consist of medium- to coarse-
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19 142 grained amphibole, plagioclase, and minor biotite phenocrysts, with K-feldspar,
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22 143 plagioclase, and quartz in the groundmass. Based on compositional and textural
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24 144 differences and crosscutting relationships, Rees et al. (2015) identified four distinct
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26 145 porphyry phases, P1, P2, P3, and P4. The P1 porphyry is a pre-mineralization
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28 146 leucodiorite which accounts for the main volume of the Red Stock, and is distinguished
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30 147 by sparse anhedral to subhedral amphibole (~10%) and abundant plagioclase (30–40
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32 148 vol. %) phenocrysts with lengths up to 4 mm. The groundmass is composed of fine-
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35 149 grained plagioclase and minor quartz (Fig. 3A).

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40 150 The P2 porphyry is a syn-mineralization quartz monzonite intrusion which is largely
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42 151 unexposed at surface but is observed in drill core to have intruded P1 at depth. It has been
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44 152 subdivided by Rees et al. (2015) into early (P2E), intermediate (P2I), and late (P2L)
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46 153 phases based on vein truncations and chilled margins. The P2 porphyries are generally
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48 154 characterized by tabular subhedral to euhedral amphibole (10–15 vol. %) and plagioclase
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50 155 (35–50 vol. %) phenocrysts. The amphibole crystals are mostly euhedral and larger than
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52 156 in P1, with lengths up to 10 mm. The groundmass comprises K-feldspar, plagioclase, and
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55 157 quartz. In this paper, the P2 subphases are simplified to an early stage (P2E; Fig. 3B) and
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57 158 late stage porphyry (P2L, probably corresponding to P2I and P2L of Rees et al., 2015;
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4 159 Fig. 3C). The P2E porphyry has a crowded plagioclase texture with slightly smaller
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6 160 amphibole phenocrysts than P2L. In contrast, the P2L porphyry is relatively fresh and
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9 161 occurs only as small dikes (Fig. 4).

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12 162 The post-mineralization P3 monzonite porphyry is much less abundant. It is texturally
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15 163 similar to the P2L porphyry with abundant amphibole phenocrysts (15–20 vol. %), but is
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17 164 distinguished by the absence of quartz in the groundmass, which is mainly composed of
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20 165 K-feldspar and plagioclase (Fig. 3D). Although no crosscutting relationships between the
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22 166 P3 and P2L porphyries have been found, the zircon U-Pb dating results (see below)
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24 167 confirm that P3 is younger. The P4 porphyry occurs as rare dikes and is typified by
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27 168 sparse fine-grained amphibole phenocrysts (Rees et al., 2015). No P4 samples were
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30 169 included in this study.

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33 170 Several younger basaltic to andesitic dikes with sparse amphibole phenocrysts cut the
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35 171 Red Stock and the Stuhini Group host rocks (Figs. 3E and F). They postdate the
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37 172 porphyry-stage Cu-Au mineralization (Baker et al., 1997; Rees et al., 2015), but are
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40 173 mildly to strongly altered (Figs. 3E and F) and are crosscut by late quartz-calcite-pyrite
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43 174 veins.

44 45 175 *Hydrothermal alteration*

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49 176 Alteration at Red Chris has been described previously by Baker et al. (1997),
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51 177 Gillstrom et al. (2012), Norris (2012) and Rees et al. (2015). The alteration assemblages
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53 178 are typical of calc-alkaline porphyry Cu systems (Lowell and Guilbert, 1970; Seedorff et
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56 179 al., 2005; Sillitoe, 2010), and consists of early stage potassic alteration, overprinted by
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59 180 chlorite-sericite, sericitic (phyllic), intermediate argillic, and minor late stage propylitic

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4 181 ~~alteration (see paragenetic diagram in Rees et al., 2015)~~Alteration at Red Chris is typical
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6 182 ~~of calc-alkaline porphyry Cu systems (Lowell and Guilbert, 1970; Seedorff et al., 2005;~~
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8 183 ~~Sillitoe, 2010), and consists of early stage potassic alteration, overprinted by chlorite–~~
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10 184 ~~sericite, sericitic (phyllic), intermediate argillic, and minor late stage propylitic alteration~~
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12 185 ~~(see paragenetic diagram in figure 13 of Rees et al., 2015; Baker et al., 1997; Gillstrom et~~
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14 186 ~~al., 2012; Norris, 2012).~~ Potassic alteration is expressed by replacement of amphibole
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16 187 phenocrysts by secondary biotite, plagioclase replaced or rimmed by secondary K-
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18 188 feldspar, and by K-feldspar veins (Figs. 3B and 5A–C). It is best preserved in the deeper
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20 189 levels of the deposit where it is spatially associated with the syn-mineralization P2E
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22 190 porphyry; however it locally extends into pre-mineralization P1 porphyry wall rocks (Fig.
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24 191 4). The P2L and P3 porphyries ~~are~~were only weakly affected by potassic alteration (Rees
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26 192 et al., 2015).

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34 193 Chlorite–sericite alteration is characterized by chlorite replacing secondary biotite
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36 194 (Fig. 5C) and sericite replacing feldspar (Fig. 5D). At shallower levels in the system,
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38 195 potassic alteration is completely overprinted by phyllic and intermediate argillic
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40 196 alteration (Gillstrom et al., 2012), characterized by sericite after plagioclase (phyllic; Fig.
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42 197 5D), and illite and kaolinite (intermediate argillic; Norris, 2012). This lower temperature
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44 198 alteration overprint affects all the porphyry phases, but is less pervasive at depth.

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49 199 Propylitic alteration at Red Chris is mainly observed as minor chlorite and epidote in
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51 200 the outer part of the Red Stock, and extends for 100 to 200 m into the Stuhini volcanic
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53 201 country rocks (Gillstrom et al., 2012; Norris, 2012; Rees et al., 2015).

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58 202 *Vein styles and mineralization*

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4 203 Detailed descriptions of vein styles at Red Chris have been given by Norris (2012)
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6 204 and Rees et al. (2015). A-type quartz veins (Gustafson and Hunt, 1975) and stockworks
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9 205 are associated with potassic alteration and host the bulk of the copper-gold mineralization.
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11 206 These veins are most intensely developed around the apex of the principal P2 porphyry
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14 207 body, but extend for hundreds of metres into the P1 wall rocks (~~see figure 13 in Rees et~~
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16 208 ~~al., 2015~~). Typical A veins contain K-feldspar, biotite, chalcopyrite, bornite, and
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19 209 magnetite, with K-feldspar alteration halos (Figs. 5A–B and G). Copper sulfides also
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21 210 occur as disseminations in the host porphyry. Bornite is more abundant in the apex of
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23 211 P2E, and progressively decreases outwards where chalcopyrite is the dominant Cu-
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25 212 sulfide (Norris, 2012; Rees et al., 2015). At shallower depths in the preserved system,
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27 213 early bornite was sulfidized to chalcopyrite, and pyrite becomes increasingly dominant.
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29 214 Rees et al. (2015) delineated a high-sulfur contour (>4% S) in section above which total
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31 215 sulfide (dominantly pyrite) ranges from 4 to 10%. Microscopic native gold and electrum
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33 216 occur as inclusions in bornite (Rees et al., 2015). The grades of Cu and Au are positively
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35 217 correlated with quartz vein density (Gillstrom et al., 2012). In high-grade zones at depth
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37 218 (e.g., 4.12 % Cu and 8.83 g/t Au in hole 09-350 from 540 to 692.5 m down-hole depth),
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39 219 vein abundance exceeds 80 vol. % in sheeted arrays (Fig. 5G; Rees et al., 2015).
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46 220 B- and D-type veins (Gustafson and Hunt, 1975) are relatively minor at Red Chris,
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48 221 and host only minor amounts of Cu sulfides. B quartz veins are characterized by
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50 222 relatively straight margins with sulfide centerlines (pyrite and minor chalcopyrite, and
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52 223 locally molybdenite; Fig. 5E). Pyritic D veins have variable widths (1 to 10 mm; Fig. 5F).
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54 224 Carbonate and minor chlorite veins cut all the earlier veins and are generally barren
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56 225 (Norris, 2012).
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4 **226 Sampling and Analytical Methods**

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6 **227** Samples of the P1, P2E, P2L, and P3 porphyries were collected from drill core, and
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9 **228** descriptions and locations are listed in Digital Appendix Table A1; sampled drill hole
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11 **229** locations are also shown on Figure 2. Fourteen least-altered samples of the porphyry
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14 **230** intrusions and three samples of late basaltic to andesitic dikes were selected for whole-
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16 **231** rock geochemical analysis. Eight of these samples were selected for determination of Nd-
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18 **232** Sr isotopes. Three samples of quartz-carbonate-pyrite-molybdenite-chalcopyrite veins
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21 **233** (Fig. 5H) were collected for Re-Os dating. Four samples of the P1 (RC13-40), P2E
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23 **234** (RC13-107), P2L (RC13-33), and P3 (RC13-78) intrusions were selected for zircon U-Pb
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26 **235** dating, Hf isotopic, and trace element analyses. Details of analytical methods are
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29 **236** provided in Appendix 1.

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31 **237**
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33 **238** *Electron microprobe analyses*

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36 **239** Primary igneous minerals such as plagioclase and amphibole in the porphyry rocks
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38 **240** are widely altered to K-feldspar, sericite, and chlorite (Figs. 3A–B and 5C–D). However,
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41 **241** a few least-altered samples of the P2E, P2L, and P3 porphyries contained unaltered
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43 **242** plagioclase and amphibole grains. In addition, igneous apatite grains were typically
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45 **243** preserved as inclusions within plagioclase and amphibole phenocrysts. Compared with
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48 **244** hydrothermal apatite (acicular crystals intergrown with other hydrothermal minerals such
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50 **245** as quartz, sericite, chlorite, and sulfides), igneous apatite grains typically showed stubby
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53 **246** prismatic habits, as described by Richards et al. (2017). Detailed analytical methods are
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55 **247** described in Appendix 1.

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58 **248**
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4 249 *Re-Os molybdenite dating*

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6 250 Three samples of molybdenite from quartz-carbonate-pyrite-molybdenite-chalcopyrite
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9 251 veins were collected from the Gully zone (RC13-88 and RC13-103), and the East zone
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11 252 (RC13-82; Digital Appendix Table A1; Fig. 2). A molybdenite mineral separate was
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13
14 253 produced for each sample by metal-free crushing followed by gravity and magnetic
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16 254 concentration methods. Dating was conducted at the Canadian Centre for Isotopic
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19 255 Microanalysis at the University of Alberta, Canada, using methods described in
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21 256 Appendix 1.
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26 258 **Geochronological Results**

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28 259 *Zircon U-Pb ages of the Red Stock*

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31 260 Zircon U-Pb results are presented in Digital Appendix Table A2 and illustrated in
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33 261 Figure 6; all ages are illustrated and reported with 2σ errors. All the zircons show
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36 262 oscillatory zoning under BSE imaging. Sample RC13-40 was collected from P1
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38 263 leucodiorite porphyry (Digital Appendix Table A1). Analyzed zircon grains form a
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41 264 tightly clustered age population, mostly with low common lead contents. Except for one
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43 265 inherited or xenocryst zircon (apparent $^{206}\text{Pb}/^{238}\text{U}$ age = 261 ± 12 Ma), the twenty-seven
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45
46 266 grains yielded an intercept age of 211.6 ± 1.3 Ma (MSWD = 0.85; Fig. 6A), similar to the
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48 267 weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 211.8 ± 1.3 Ma (MSWD = 0.8).
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51 268 Sample RC13-107 was collected from P2E quartz monzonite porphyry (Digital
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53 269 Appendix Table A1). One xenocrystic zircon grain yielded an older age (apparent
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55 270 $^{206}\text{Pb}/^{238}\text{U}$ age = 316 ± 10 Ma), but the other twenty-nine zircons contained low amounts
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58 271 of common lead and intersected the concordia line at 206.0 ± 1.2 Ma (MSWD = 1.5; Fig.
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4 272 6B), in good agreement with the weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 205.9 ± 1.5 Ma
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6 273 (MSWD = 1.5).

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9 274 Sample RC13-33 was collected from P2L quartz monzonite porphyry (Digital
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11 Appendix Table A1). Two xenocryst grains were found among thirty-one analyzed
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13 275 zircons, with $^{206}\text{Pb}/^{238}\text{U}$ ages of 298 ± 16 Ma and 441 ± 13 Ma. The remaining twenty-
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15 276 nine zircon grains had low common lead contents and yielded an intercept age of $203.6 \pm$
16
17 277 1.8 Ma (MSWD = 1.5; Fig. 6C), similar to the weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of $203.3 \pm$
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19 278 1.5 Ma (MSWD = 1.04).
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23 280 Sample RC13-78 was collected from P3 monzonite porphyry (Digital Appendix Table
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25 281 A1). Thirty-two zircons with low common lead contents yielded an intercept age of 201.7
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27 ± 1.2 Ma (MSWD = 1.05; Fig. 6D), and a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 201.6 ± 1.2
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29 282 Ma (MSWD = 1.04).
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33 284 All of the zircons have magmatic textures, and the intercept ages above are interpreted
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35 285 to be the crystallization ages of the individual intrusions. The ages are consistent with
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37 286 crosscutting relationships described by Rees et al. (2015), and define a 10 m.y. span of
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39 287 magmatism from 211.6 ± 1.3 Ma (P1) to 201.7 ± 1.2 Ma (P3). The relative ages of the
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41 288 two samples of syn-mineralization P2 porphyry are consistent with crosscutting
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43 289 relationships, but the dates (P2E: 206.0 ± 1.2 Ma; P2L: 203.6 ± 1.8 Ma) overlap within
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45 290 the 2σ uncertainty. Hence the apparent 2.4 m.y. age difference is not statistically robust,
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47 291 and their true ages may in fact be closer as suggested by their close relationship with ore
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49 292 mineral paragenesis.
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58 294 *Re-Os molybdenite ages*
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4 295 The Re-Os model ages for three samples are shown in Table 1. They have relatively
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6 296 high rhenium contents ranging from 497.8 to 1771 ppm, with common ^{187}Os of 1078 to
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9 297 3821 ppb. Although the three samples are from two different mineralization zones (i.e.,
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11 298 East and Gully zones; Fig. 2), they yielded indistinguishable model ages within
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13
14 299 uncertainty: 206.5 ± 0.8 Ma, 205.7 ± 0.9 Ma, and 205.9 ± 1.1 Ma (± 2 s.d.). The small
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16 300 grain size of the molybdenite (< 1 mm) minimizes the risk of decoupling between Re and
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19 301 ^{187}Os in these samples (caused by diffusion: Selby and Creaser, 2004). The results
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21 302 yielded a weighted average age for all three samples of 206.1 ± 0.5 Ma (95% confidence
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24 303 level with MSWD = 0.96; Fig. 7). This age is consistent with the age of the syn-
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26 304 mineralization P2E porphyry (206.0 ± 1.2 Ma).

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30 306 **Geochemical and Isotopic Results**

31 307 *Whole-rock major and trace elements*

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33 308 Whole-rock major and trace element compositions for fourteen samples of the Red
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36 309 Stock and three basaltic–andesitic dike samples are listed in Table 2. All the P1 to P3
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39 310 porphyry intrusions have relatively homogeneous major element compositions, but have
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42 311 moderate to high loss-on-ignition values (LOI: 2.3 to 10.9%) reflecting varying degrees
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45 312 of potassic and sericitic alteration. On an LOI-free basis, they are mostly intermediate
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48 313 composition ($\text{SiO}_2 = 56.71\text{--}63.16$ wt. %; Digital Appendix Table A3), and straddle the
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51 314 boundary between granodiorite (diorite) and syenite on a Zr/Ti versus Nb/Y diagram (Fig.
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53 315 8). The late basaltic–andesitic dikes have distinct compositions, and plot in the field of
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56 316 diorite and gabbro on Figure 8. On Harker-type diagrams of SiO_2 versus selected major
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58 317 elements, the alkali elements (K_2O and Na_2O ; Figs. 9A–B) show significant scatter,
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4 318 likely due to hydrothermal alteration, as confirmed by a rough correlation between alkali
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6 319 contents and LOI. However, other elements such as Fe₂O₃, MgO, TiO₂, and Al₂O₃ show
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9 320 roughly correlated trends with SiO₂ (Figs. 9E–F), suggesting a broadly cogenetic suite,
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11 321 with the exception of the late basaltic to andesitic dikes, which do not plot on the same
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14 322 trends for K₂O and TiO₂.

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16 323 On primitive mantle-normalized extended trace element and chondrite-normalized
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18 324 rare earth element (REE) diagrams (Figs. 10A–B), the porphyritic rocks show largely
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21 325 indistinguishable patterns consistent with typical subduction-related igneous rocks
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24 326 (Hawkesworth et al., 1993; Pearce, 1996), characterized by large-ion lithophile element
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26 327 (LILEs: Rb, Ba, Th, U, K) and light rare earth element enrichments (LREE), low
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28 328 concentrations of high field strength elements (HFSE: Nb, Ta, Zr, Hf, and Ti), relative
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31 329 depletions in compatible elements and middle to heavy rare earth elements (MREE,
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33 330 HREE; La/Yb = 8.9 ± 1.7 , n = 14), and flat to listric-shaped patterns from MREE to
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36 331 HREE. Such listric patterns likely reflect amphibole fractionation from hydrous magmas,
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38 332 and are an indication of magma fertility for porphyry ore formation (Richards and
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41 333 Kerrich, 2007). Most samples have flat or slightly positive Eu anomalies with Eu_N/Eu_N^*
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43 334 of 1.1 ± 0.15 (n = 14; Fig. 10B), also likely reflecting amphibole fractionation and lack of
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45
46 335 plagioclase fractionation.

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48 336 Three samples of late basaltic to andesitic dikes show distinct trace element patterns,
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51 337 especially for HFSE and REE, which are enriched relative to the porphyries; they are
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53 338 clearly not cogenetic with the earlier porphyries.

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55 339 Excluding two samples with high Sr values that may reflect calcite veining, the
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58 340 porphyries have relatively high Sr/Y (53 ± 23 , n = 12) and V/Sc ratios (9.0 ± 0.9 , n = 14),
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4 341 which overlap the range of fertile rocks for porphyry formation (Richards and Kerrich,
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6 342 2007; Loucks, 2013, 2014).
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11 344 *Whole-rock Nd-Sr isotopes*
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14 345 Nd and Sr isotopic compositions of the porphyritic rocks are very homogeneous (Table
15
16 346 4 and Fig. 11). They show a narrow range of initial Sr isotopic ratios from 0.7042 to
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18 347 0.7047 ($(^{87}\text{Sr}/^{86}\text{Sr})_t = 0.7044 \pm 0.0002$, $n = 8$), with relatively high $\epsilon_{\text{Nd}}(t)$ values from 2.4
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21 348 to 3.6 (average = $+3.1 \pm 0.4$, $n = 8$), consistent with a mantle source with minor crustal
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23 349 contamination. The single-stage Nd model ages (T_{DM1}) calculated following the methods
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25 350 of Goldstein et al. (1984) range from 0.74 to 0.88 Ga (average = 0.80 ± 0.05 Ga, $n = 8$),
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27 351 and are indistinguishable for the different porphyry phases. Two basaltic–andesitic dike
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29 352 samples show slightly lower Sr isotopic ratios ($(^{87}\text{Sr}/^{86}\text{Sr})_t = 0.7042$ and 0.7043) and
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31 353 higher $\epsilon_{\text{Nd}}(t)$ ratios ($+3.5$ and $+3.9$) than the porphyry rocks (Fig. 11).
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36 354 *Zircon Hf isotopes*
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39 355 Zircon Hf isotopic results are listed in Digital Appendix Table A4, and illustrated in
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41 356 Figure 12. The four zircon samples from porphyritic rocks show indistinguishable Hf
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43 357 isotopic compositions and single-stage Hf model ages of 375 ± 52 Ma (range = 264–527
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45 358 Ma, $n = 56$), with relatively high $\epsilon_{\text{Hf}}(t)$ values of 12.0 ± 0.4 (weighted mean value, 95%
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47 359 confidence level; range = 8.1–14.8, $n = 56$). These data suggest that the porphyries shared
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49 360 a common primitive mantle source, consistent with their island arc origin.
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56 362 **Amphibole, Plagioclase, Zircon, and Apatite Compositions**
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4 363 The compositions of amphibole, plagioclase, zircon, and apatite from samples of least-
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6 364 altered porphyritic rocks are listed in Digital Appendix Tables A5, A6, A7, and A8,
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8
9 365 respectively. Based on these analyses, water and sulfur contents as well as oxygen
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11 366 fugacity of the magma have been estimated qualitatively and quantitatively.
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16 368 *Amphibole compositions*
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19 369 Amphibole compositions can be used to estimate magmatic oxidation state,
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21 370 crystallization temperatures and pressures, as well as water contents (e.g., Ridolfi et al.,
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23 371 2010; Zhang et al., 2012; Wang et al., 2014b). However, fresh amphibole grains were
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25 372 only observed in the P2E and P2L porphyries. All of the analyzed grains are calcic
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27 373 amphiboles, and two generations of amphibole were recognized in both P2E and P2L.
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30 374 Early stage grains occur as phenocrysts and are characterized by higher Al contents
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32 375 (range = 9.35–12.86 and 8.79–13.46 wt. % Al_2O_3 for P2E and P2L, respectively; Digital
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34 376 Appendix Table A5); they are mostly classified as magnesio-ferri-hornblende and
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36 377 magnesio-hastingsite (including potassic-magnesio-hastingsite; Fig. 13A). The late stage
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38 378 grains were developed in the groundmass or as recrystallized phenocrysts, typified by
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40 379 lower Al concentrations (range = 1.62–4.76 and 4.33–7.53 wt. % Al_2O_3 for P2E and P2L,
41
42 380 respectively; Digital Appendix Table A5). They are classified as magnesio-ferri-
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44 381 hornblende or rare actinolite (Fig. 13A). The early stage amphibole grains are intergrown
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46 382 with plagioclase phenocrysts (Fig. 14), and igneous apatite crystals occur as inclusions
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48 383 within amphibole (Fig. 14B). This indicates that these three minerals are broadly coeval.
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53 384 The crystallization temperature, magmatic water contents, and oxygen fugacities were
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55 385 estimated for the two stages of amphiboles using the spreadsheet of Ridolfi et al. (2010),
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4 386 and the crystallization pressures were calculated following the equation of Mutch et al.
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6 387 (2016). The Al-in-hornblende geobarometer used is applicable to granitoids that contain
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8 388 amphibole, plagioclase, quartz, and alkali feldspar, broadly consistent with the mineral
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10 389 assemblage within P2 porphyry rocks. The calculated results are listed in Digital
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14 390 Appendix Table A5 and illustrated in Figure 13 (B–C).

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16 391 Amphibole phenocrysts from the P2E and P2L porphyry samples have similar
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18 392 compositions, with calculated crystallization pressures from 3.5 to 6.7 kbar (average =
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20
21 393 4.6 ± 0.8 kbar, $n = 51$), temperatures from 855 to 983 °C (average = 900 ± 30 °C, $n = 51$),
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23 394 Δ FMQ values from 0.5 to 1.8 (average = 1.1 ± 0.3 , $n = 51$), and H₂O contents in melts
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26 395 from 4.0 to 6.1 wt. % (average = 5.1 ± 0.4 wt. %, $n = 51$).

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28 396 Late stage (low-Al) amphiboles in P2E porphyry samples yielded crystallization
29
30 397 pressures from 0.6 to 1.4 kbar (average = 1.1 ± 0.3 kbar, $n = 12$), temperatures from 637°
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32 398 to 774 °C (average = 719 ± 34 °C, $n = 12$), Δ FMQ values from 2.6 to 3.3 (average = $3 \pm$
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34 399 0.2 , $n = 12$), and magmatic water contents from 3.6 to 5.1 wt. % (average = 4 ± 0.4 wt. %, n
35
36 400 = 12). Low-Al amphibole grains from P2L samples yielded slightly higher crystallization
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38 401 pressures (2.0 ± 0.4 MPa; range = 1.3–2.6 kbar, $n = 8$) and temperatures (800 ± 24 °C;
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40 402 range = 751–800 °C, $n = 8$), slightly lower Δ FMQ values (2.4 ± 0.3 ; range = 1.1–2.9, $n =$
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42 403 8), but similar H₂O contents (4.0 ± 0.2 ; range = 3.7–4.4, $n = 8$) to those calculated for
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44 404 P2E.

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46 405 There are clear trends of increasing oxidation state and decreasing crystallization
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48 406 pressures, temperatures, and magmatic water contents from early to late stage amphiboles
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50
51 407 in both P2E and P2L porphyries (Figs. 13B–D). These trends are consistent with the
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54 408 high-Al amphibole phenocrysts having crystallized at depth before final crystallization of
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4 409 the magma (and low-Al amphibole) after emplacement at shallow levels (e.g., Rutherford
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6 410 and Devine, 2003). The decreasing magmatic water contents might be attributed to
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9 411 degassing during ascent and crystallization, which can also result in a small increase in
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11 412 oxygen fugacity (Mathez, 1984; Candela, 1986; Burgisser and Scaillet, 2007; Zimmer et
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13 413 al., 2010; Bell and Simon, 2011; Dilles et al., 2015).

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19 415 *Plagioclase compositions*

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21 416 Plagioclase phenocrysts from all P1 porphyry samples studied have been altered, but
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23 417 partially unaltered phenocrysts were found in some P2 and P3 porphyry samples.

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25 418 Analyses with $K_2O > 1$ wt. % are interpreted to reflect potassic alteration and have been
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27 419 excluded. Plagioclase grains from the P2E porphyry (sample RC13-13) are classified as
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29 420 oligoclase ($\geq An_{17}$) to labradorite ($\leq An_{70}$), with an average anorthite (An) content of $44 \pm$
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31 421 12 mol % ($n = 28$). Nearly all these crystals exhibited reverse zoning in An content, with
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33 422 compositional ranges up to ~ 16 mol %, and FeO contents that correlate positively with
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35 423 X_{An} values (Digital Appendix Table A6; Fig. 14A). Plagioclase grains from the P2L
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37 424 porphyry (samples RC13-26 and RC13-33) have compositions similar to P2E (An mol %
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39 425 = 48 ± 5 , $n = 38$), and also show reverse zoning with amplitudes up to ~ 19 mol % An that
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41 426 correlate with FeO contents (Fig. 14B). Such reverse zoning with positive An-FeO
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43 427 correlations were observed both in partially altered (i.e., sericite alteration; Fig. 14A) and
44
45 428 relatively fresh plagioclase crystals (Fig. 14B), indicating that it was not likely to have
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47 429 been generated by hydrothermal alterations. In contrast, plagioclase grains from the P3
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49 430 porphyry (RC13-78) display relatively uniform An compositions (An mol % = 51 ± 5 , n
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51 431 = 37), with no clear zoning in X_{An} and FeO (Fig. 14C).

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4 432 Plagioclase compositions (An proportions) are very sensitive to dissolved water
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6 433 content and temperature of the melt, and thus can be used as a hygrometer if corrected for
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8
9 434 temperature (Mathez, 1973; Lange, 2009; Waters and Lange, 2015). We used apatite
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11 435 saturation temperatures (AST) following the formula of Piccoli and Candela (1994) to
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13 436 calculate magma temperatures for P2 and P3 samples (Table 4; Digital Appendix Table
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15
16 437 A3). The whole-rock SiO₂ and P₂O₅ values were assumed to approximate the melt
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18 438 contents at the time of crystallization, although this assumption can be challenged
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21 439 (Piccoli and Candela, 1994, 2002). Nevertheless, the calculated temperatures are
22
23 440 reasonable for magmas of intermediate, hydrous composition, and are also in agreement
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25 441 with the estimated crystallization temperatures of early stage amphibole (Digital
26
27
28 442 Appendix Table A5): 932 °C for P2E (RC13-13), 892 °C (RC13-26) and 905 °C (RC13-
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30 443 33) for P2L, and 928 °C for P3 (RC13-78; Table 4).

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32
33 444 The updated spreadsheet of Waters and Lange (2015) has been used to calculate
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35 445 magmatic water contents (Digital Appendix Table A9). The calculated values are
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37 446 relatively uniform and indistinguishable between the different porphyry phases: 5.2 ± 0.2
38
39 447 ($n = 28$) for P2E, 5.5 ± 0.2 ($n = 38$) for P2L, and 5.1 ± 0.1 wt. % ($n = 37$) for P3. For the
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41 448 P2E and P2L porphyries, the magmatic water contents calculated from plagioclase and
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43 449 early stage amphibole compositions agree well with each other (Digital Appendix Tables
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46 450 A5 and A8).

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52 452 *Zircon trace element compositions*

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55 453 Twenty trace element spot analyses were obtained for zircons from samples of P1 to
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58 454 P3 porphyries (Digital Appendix Table A7; Fig. 15). It is common to encounter small

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4 455 mineral inclusions, especially of apatite, titanite, in zircon during LA-ICP-MS analyses
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6 456 (e.g., Lu et al., 2016). We have taken Ca > 200 ppm or La > 0.3 ppm as an indication of
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9 457 apatite contamination, and Ti > 20 ppm to reflect titanite contamination, and such data
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11 458 were excluded (Digital Appendix Table A7). The remaining analyses have low LREE
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13
14 459 and elevated HREE contents, with small negative Eu and strongly positive Ce anomalies
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16 460 (Fig. 15). These REE patterns and total contents ($\Sigma\text{REE} = 351\text{--}1125$ ppm), as well as
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19 461 Th/U ratios above 0.1 (0.26–0.63) (Digital Appendix Table A7) are typical of igneous
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21 462 zircons from relatively oxidized magmas (Hoskin and Schaltegger, 2003).

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23
24 463 Zircon crystals from the P1 to P3 porphyries show slightly different calculated Ti-in-
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26 464 zircon temperatures and Eu anomalies ($\text{Eu}_\text{N}/\text{Eu}_\text{N}^* = \text{Eu}_\text{N}/(\text{Sm}_\text{N} \times \text{Gd}_\text{N})^{0.5}$; Fig. 16; Digital
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28
29 465 Appendix Table A7). Titanium-in-zircon temperatures were calculated using the equation
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31 466 of Ferry and Watson (2007), where it is assumed that $\log a_{\text{SiO}_2} = 1$ because of the
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33 467 existence of quartz in the host porphyries, and $\log a_{\text{TiO}_2} = 0.7$ due to the presence of
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35
36 468 titanite (a common accessory mineral in these rocks). Zircon crystals from the P2E
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38
39 469 porphyry (sample RC13-107) have the highest calculated temperatures (average of $760 \pm$
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41 470 29 °C; range = $723^\circ\text{--}819^\circ\text{C}$, $n = 14$) with relatively uniform $\text{Eu}_\text{N}/\text{Eu}_\text{N}^*$ values from 0.5 to
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43
44 471 0.7 (average = 0.6 ± 0.1 ; $n = 14$). Zircon crystals from the P3 porphyry (sample RC13-78)
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46 472 show the lowest calculated temperatures (average = 726 ± 24 °C; range = $693^\circ\text{--}758^\circ\text{C}$, n
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48 473 = 18) and lowest $\text{Eu}_\text{N}/\text{Eu}_\text{N}^*$ ratios (0.5 ± 0.1 ; range = 0.3–0.6, $n = 18$). Zircon crystals
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50
51 474 from the P1 and P2L porphyries have intermediate calculated temperatures (P1: $736 \pm$
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53 475 26 °C, range = $699^\circ\text{--}810^\circ\text{C}$, $n = 16$; P2L: 730 ± 24 °C, range = $705^\circ\text{--}795^\circ\text{C}$, $n = 17$),
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55
56 476 and a similar wide range of $\text{Eu}_\text{N}/\text{Eu}_\text{N}^*$ values (P1: 0.3–1.0, average = 0.6 ± 0.2 ; P2L: 0.3–
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58 477 0.9, average = 0.6 ± 0.1).

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4 478 Trace element compositions in zircon are sensitive to magmatic water content and
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7 479 oxygen fugacity, and have been used to distinguish between fertile and infertile suites in
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10 480 porphyry Cu ± Au ± Mo systems (Ballard et al., 2002; Liang et al., 2006; Qiu et al., 2014;
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12 481 Wang et al., 2014b; Dilles et al., 2015; Shen et al., 2015; Lu et al., 2016; Xu et al., 2016;
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14 482 Loader et al., 2017). In hydrous magmas, early plagioclase crystallization is suppressed
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16 483 (Moore and Carmichael, 1998), leading to no pronounced depletion of Eu in melts (Eu²⁺
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18 484 substitutes for Ca²⁺ in plagioclase). On the other hand, in oxidized magmas, Eu is
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21 485 predominantly present as Eu³⁺ and is partitioned into zircon along with other REE
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24 486 (Ballard et al., 2002; Hoskin and Schaltegger, 2003; Trail et al., 2011; Dilles et al., 2015).
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26 487 Therefore, the small negative Eu anomalies in zircon samples (as indicated by Eu_N/Eu_N^{*}
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28 488 ratios >0.3 and mostly > 0.4; Fig. 16A), and slightly positive anomalies in whole rocks
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31 489 (Fig. 10) indicate relatively high magmatic water contents and/or oxidation states
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34 490 (Ballard et al., 2002; Dilles et al., 2015; Lu et al., 2016).

35
36 491 Ballard et al. (2002) defined the zircon Ce⁴⁺/Ce³⁺ ratio as an indicator of magmatic
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38 492 oxidation state. However, Dilles et al. (2015) and Lu et al. (2016) argue that these values
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41 493 are difficult to estimate accurately because the abundances of the adjacent elements, La
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43 494 and Pr, which are used as a baseline to calculate the magnitude of the Ce anomaly, are
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45 495 low and close to the analytical detection limit. Confirming this reservation, Ce⁴⁺/Ce³⁺
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48 496 ratios and *f*O₂ values calculated for our samples using the equation of Trail et al. (2011)
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51 497 yielded unrealistic and widely variable ΔFMQ values (-9 to +5).

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55 499 *Apatite compositions*
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4 500 The SO₃, Cl, and F analyses of igneous apatites are listed in Digital Appendix Table
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6 501 A8 and illustrated in Figures 17 and 18. The results show that igneous apatites from the
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8 502 P2 porphyries have higher sulfur and chlorine concentrations than those from P1 and P3
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10 503 rocks. The high SO₃ contents in apatites from the P2 porphyries (P2E: 0.11 to 0.8 wt. %
11
12 504 SO₃, average = 0.30 ± 0.13 wt. %, n = 34; P2L: 0.07 to 1.2 wt. % SO₃, average = 0.29 ±
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14 505 0.18 wt. %, n = 100) are similar to values from global porphyry Cu deposits (Streck and
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16 506 Dilles, 1998; Imai, 2002; Li et al., 2012; Pan et al., 2016; Richards et al., 2017). In
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18 507 comparison, apatite crystals from the P1 porphyry (RC13-39) have lower sulfur contents
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20 508 (0.05 to 0.19 wt. % SO₃; average = 0.11 ± 0.03 wt. %, n = 34), whereas those from the P3
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22 509 porphyry (RC13-78) have the lowest sulfur concentrations (from 0.05 wt. % to below the
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24 510 analytical detection limit of SO₃; average of analyses above limit of detection = 0.03 ±
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26 511 0.01, n = 13). Although the ranges for P1 and P2L porphyries overlap at their outer limits,
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28 512 a t-test shows that the two populations are significantly different (t = 5.87, p = 0).

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31 513 Similarly, the Cl contents in apatite crystals from P1 and P3 porphyries are
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33 514 significantly lower (0.51 ± 0.3 wt. %, n = 34, and 0.1 ± 0.06 wt. %, n = 17, respectively)
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35 515 compared to values from the P2 porphyries (P2E: 1.18 ± 0.37 wt. %, n = 34; P2L: 1.47 ±
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37 516 0.28 wt. %, n = 100). There is a rough positive correlation between sulfur and chlorine
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39 517 contents of apatites from the four porphyries (Fig. 18A), with relatively constant molar
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41 518 S/Cl ratios (P1: 0.13 ± 0.08, n = 34; P2E: 0.13 ± 0.08, n = 34; P2L: 0.10 ± 0.14, n = 100;
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43 519 P3: 0.16 ± 0.09, n = 13; Table 4). Detailed analyses also show that single apatite crystals
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45 520 are commonly zoned in sulfur, with decreasing core-to-rim SO₃ contents (and small
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47 521 decreases in Cl; Fig. 17C), possibly reflecting progressive degassing of SO₂ (and Cl)
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49 522 from the melt during crystallization (Richards et al., 2017).
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4 523 Fluorine contents in apatites from the P1 and P3 porphyries are similar (P1: $3.64 \pm$
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6 524 0.48 wt. %, $n = 34$; P3: 3.6 ± 0.4 , $n = 17$) and distinctly higher than in the P2E ($2.33 \pm$
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9 525 0.29 wt. %, $n = 34$) and P2L (2.32 ± 0.23 wt. %, $n = 100$) porphyries.

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11 526 Some hydrothermal apatite grains were also analyzed for comparison with igneous
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14 527 grains, and they show wide variations in sulfur and chlorine content (Digital Appendix
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16 528 Table A8), as observed by other researchers (e.g., Li et al., 2012; Richards et al., 2017).

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19 529 The sulfur partition coefficient between apatite and melt is jointly controlled by
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21 530 temperature, oxygen fugacity, and the S content in the silicate melt (Peng et al., 1997;
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24 531 Parat and Holtz, 2005; Parat et al., 2011; Webster and Piccoli, 2015; Konecke et al.,
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26 532 2017a, b), and no formula currently exists to accurately calculate magmatic sulfur
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29 533 concentrations from apatite SO_3 contents. However, two semi-quantitative formulae can
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31 534 be used to estimate relative magmatic S content (Peng et al., 1997; Parat et al., 2011).
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33 535 Using the equation of Peng et al. (1997), the magmas related to the P2E and P2L
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36 536 porphyry intrusions are calculated to have sulfur concentrations of 0.03 ± 0.01 wt. % ($n =$
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38 537 34) and 0.02 ± 0.01 wt. % ($n = 100$), significantly higher than the corresponding S
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41 538 contents for P1 (0.010 ± 0.003 wt. % S, $n = 34$) and P3 magmas (0.003 ± 0.001 wt. % S,
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43 539 $n = 13$; Table 4). The results calculated by the method of Parat et al. (2011) are more
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45
46 540 variable, and yielded lower absolute values than those from Peng et al. (1997). However,
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48 541 they also support the conclusion that the P2E and P2L magmas had significantly higher
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51 542 sulfur contents than the P1 and P3 magmas (Table 4).

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53 543 The chlorine partition coefficient between apatite and melt is a complex function of
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55 544 magma composition and Cl concentration in the melt (Zhu and Sverjensky, 1991; Piccoli
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58 545 and Candela, 1994; Mathez and Webster, 2005; Webster et al., 2009; Chelle-Michou and

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4 546 Chiaradia, 2017). Mathez and Webster (2005) proposed a value of 0.8 as the partition
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6 547 coefficient (mass ratio) between apatite and basaltic melt (51.1 wt. % SiO₂). Based on
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8 548 this semi-quantitative equation (and recognizing that these magmas were not basaltic), we
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11 549 estimate Cl concentrations in the P2E and P2L magmas of 1.48 ± 0.46 wt.% (n = 34) and
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14 550 1.83 ± 0.34 wt.% (n = 100), respectively, significantly higher than for P1 (0.63 ± 0.38
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16 551 wt.%, n = 34) and P3 (0.12 ± 0.07 wt.%; Table 4).
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21 553 **Discussion**

23 554 *Magmatic duration and timing of ore formation at Red Chris*

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26 555 Four phases of the Red Stock, P1, P2E, P2L, and P3, have been dated at 211.6 ± 1.3
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28 556 Ma, 206.0 ± 1.2 Ma, 203.6 ± 1.8 Ma, and 201.7 ± 1.2 Ma, respectively, spanning a period
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31 557 of ~10 m.y. (Fig. 6; Digital Appendix Table A2). These ages are consistent with
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33 558 crosscutting relationships described by Rees et al. (2015). Re-Os dating of molybdenite
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35 559 intergrown with chalcopyrite in quartz veins yielded a weighted mean age of 206.1 ± 0.5
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38 560 Ma (Fig. 7), reflecting the timing of mineralization and in good agreement with the
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41 561 emplacement age of the mineralized P2E quartz monzonite porphyry. Ore formation
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43 562 occurred at a relatively late stage in the ~10 m.y. evolution of the Red Stock, and over a
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46 563 relatively brief period time (<1 m.y., as defined by the uncertainty on the average of three
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48 564 Re-Os analyses). This relatively late and short duration of mineralization is similar to the
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51 565 timing observed in many other porphyry deposits (e.g., Cathles et al., 1997; Masterman et
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53 566 al., 2004; von Quadt et al., 2011; Chiaradia et al., 2013; Correa et al., 2016; Li et al.,
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55 567 2017).
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4 568 The age of mineralization at ~206 Ma is contemporaneous with the major pulse of
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6 569 mineralization in the Stikine and Quesnel terranes, representing a particularly fertile pre-
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9 570 accretionary magmatic event (Nelson and Colpron, 2007; Nelson et al., 2013; Logan and
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11 571 Mihalynuk, 2014).

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16 573 *Igneous geochemistry and petrogenesis*

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19 574 Whole-rock major and trace element compositions of P1 to P3 porphyry rocks are
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21 575 almost identical to each other, and show characteristics of subduction-related magmatic
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23 576 rocks with high LILE and low HFSE concentrations (Figs. 9 and 10; Hawkesworth et al.,
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25 577 1993; Pearce, 1996). The high Sr/Y ratios (53 ± 23 , $n = 12$) and slightly positive Eu
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27 578 anomalies ($Eu_n/Eu_n^* = 1.1 \pm 0.15$; $n = 14$) indicate the suppression of plagioclase
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29 579 crystallization and early crystallization of amphibole from hydrous and oxidized magmas
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31 580 (Richards and Kerrich, 2007; Loucks, 2014). Rees et al. (2015) concluded from their
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33 581 analysis of petrochemistry that the Red Stock evolved from subalkaline (P1, P2) to
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35 582 marginally alkaline (P3), and classified the overall system as high-K calc-alkalic.

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38 583 All the porphyritic rocks from P1 to P3 show mantle-like, relatively low $(^{87}\text{Sr}/^{86}\text{Sr})_t$
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40 584 (0.7044 ± 0.0002) and high $\epsilon_{\text{Nd}}(t)$ ($+3.1 \pm 0.4$) values, and similar single-stage Nd model
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42 585 ages (0.80 ± 0.05 Ga). Zircon crystals from these rocks yield positive zircon $\epsilon_{\text{Hf}}(t)$ values
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44 586 ($+8.1$ to $+14.8$) and single-stage Hf model ages of 375 ± 52 Ma (Figs. 11 and 12). These
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46 587 data indicate minimal involvement of ancient crustal components in the petrogenesis of
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48 588 these magmas, as observed for other Mesozoic igneous rocks in the Stikine terrane, and
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50 589 are consistent with an oceanic island arc setting (Samson et al., 1989; Logan and
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52 590 Mihalynuk, 2014).

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4 591 Late basaltic to andesitic dikes have similar Nd–Sr isotopic compositions to the
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6 592 porphyritic rocks, but show distinct trace element and REE patterns, and lower Sr/Y and
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9 593 La/Yb ratios, suggesting that they are not cogenetic.
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14 595 *Oxygen fugacity and sulfur-chlorine-water contents of the porphyry magmas*

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16 596 Magmatic oxidation state, and sulfur, chlorine, and water contents are fundamental
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18 597 factors for the formation of porphyry Cu deposits (e.g., Burnham, 1979; Candela, 1992;
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21 598 Richards, 2003; Loucks, 2014). It is difficult to measure the original magmatic oxidation
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23 599 state in altered plutonic rocks, but normalized Ce and Eu anomalies in zircons can be
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26 600 used to provide qualitative estimates (Ballard et al., 2002; Dilles et al., 2015; Lu et al.,
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28
29 601 2016). Zircon grains from the P1 to P3 porphyries display large positive Ce and small
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31 602 negative Eu anomalies (Eu_N/Eu_N^* mostly > 0.4 ; Figs. 15 and 16), similar to zircons from
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33 603 other mineralized porphyry systems worldwide (Wang et al., 2014b; Dilles et al., 2015;
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36 604 Lu et al., 2016). These data are also consistent with estimates of magmatic oxidation state
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38 605 from early stage amphibole phenocrysts from the P2E and P2L porphyries, which yielded
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41 606 ΔFMQ values of 1.5 ± 0.2 ($n = 7$) and 1.1 ± 0.3 ($n = 44$). Consequently, we conclude that
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43 607 the four phases of magma in the Red Stock were all at least moderately oxidized.
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46 608 It is similarly difficult to estimate original magmatic water contents from altered
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48 609 plutonic rocks. However, the presence of amphibole phenocrysts in the P1 to P3
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51 610 porphyries suggest that the magmas contained at least 4 wt. % H_2O (Burnham, 1979,
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53 611 Naney, 1983; Merzbacher and Egger, 1984; Rutherford and Devine, 1988; Ridolfi et al.,
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55
56 612 2010). This observation is supported by: (1) high whole-rock Sr/Y ratios (53 ± 23 , $n =$
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58 613 12), which reflect the suppression of plagioclase relative to amphibole crystallization
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4 614 under hydrous conditions (Moore and Carmichael, 1998; Richards and Kerrich, 2007); (2)
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6 615 high $10000 \times (\text{Eu}_N/\text{Eu}_N^*)/Y$ (>1) and low Dy/Yb (<0.3) ratios in zircons (Digital Appendix
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9 616 Table A7; Lu et al., 2016); (3) high whole-rock V/Sc ratios (9.0 ± 0.9 , $n = 14$) (Loucks,
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11 617 2014); and (4) estimates of magmatic water content from plagioclase phenocryst
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14 618 compositions, which indicate values of ~ 5 wt.% H₂O.

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16 619 The compositions of igneous apatite inclusions in plagioclase and amphibole
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18 620 phenocrysts have been used to estimate magmatic sulfur and chlorine contents. The
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21 621 results suggest that apatite microphenocrysts from the P2E and P2L porphyries had
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24 622 higher S and Cl contents than for P1 and P3 (Fig. 18). The SO₃ contents in apatite are
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26 623 controlled by both magmatic sulfur concentration and oxidation state (Peng et al., 1997;
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28 624 Parat and Holtz, 2005; Webster and Piccoli, 2015). Sulfur occurs in the apatite structure
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31 625 mainly as S⁶⁺ and its content will increase in relatively oxidized environments (Boyce et
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34 626 al., 2010; Parat et al., 2011; Konecke et al., 2017b). The magmas associated with the four
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36 627 porphyry phases at Red Chris were all moderately oxidized, and therefore the differences
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38 628 in apatite sulfur content likely mainly reflect variations in the sulfur fugacity in the
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41 629 associated melts. The data indicate that the P2E and P2L porphyry magmas contained
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44 630 significantly higher S contents than the P1 and P3 magmas (Fig. 18). Models used to
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46 631 estimate magmatic sulfur content in equilibrium with apatite (Peng et al., 1997) are not
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48 632 strictly accurate because of uncertainties in the details of exchange reactions (Streck and
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51 633 Dilles, 1998; Mao et al., 2016) but are expected to be correct in relative terms. Our data
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53 634 yield higher values for the P2E and P2L porphyries (0.02–0.03 wt. % S) compared to
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55 635 estimates for P1 (0.01 wt. % S) and P3 (0.003 wt. % S). Consequently, we conclude that

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4 636 the syn-mineral porphyry (P2E and P2L) magmas were more S-rich than the pre-
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7 637 mineralization P1 and especially the post-mineralization P3 porphyry magmas.

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9 638 Similarly, the magmatic Cl contents calculated from apatite compositions in the P2E
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11 639 and P2L porphyries are also higher than in the P1 and P3 porphyries, whereas F contents
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14 640 are lower (Table 4). Volatile species such as S, Cl, and F are differentially affected by
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16 641 degassing during ascent and crystallization of magmas: S and Cl will be preferentially
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19 642 lost to the vapor phase relative to F (Webster et al., 2014; Stock et al., 2016), leading to
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21 643 high F/Cl and F/S ratios in degassed magmas. The data shown in Figure 18 suggest that
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24 644 the P2E and P2L were S-Cl-rich, whereas the P1 and P3 were S-Cl-poor but enriched in F.

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26 645 These differences could reflect either loss of volatiles from the source magma
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29 646 chamber (presumed to be an underlying batholith) prior to emplacement of the P1 and P3
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31 647 magmas but not in the case of P2, or conversely addition of a volatile phase or a volatile-
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34 648 rich magma to the magma chamber immediately prior to emplacement of P2. Volatile
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36 649 loss is expected to be a general condition of upper crustal magma emplacement, and this
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39 650 may well be a factor contributing to the low Cl/F ratios in the late P3 magmas, as well as
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41 651 the negatively correlated trend for F and Cl in P1. The P2 magma was emplaced after P1,
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44 652 and so might be expected to be similarly degassed if this represents a comagmatic
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46 653 sequence. However, the 6–8 m.y. age difference between P1 (211.6 ± 1.3 Ma) and P2
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48 654 (P2E: 206.0 ± 1.2 Ma; P2L: 203.6 ± 1.8 Ma) indicates that P1 and P2 are not directly
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51 655 comagmatic. One explanation is that the P2 event represents a pulse of S-Cl-rich magma
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53 656 injected at a relatively late stage into the mid-crustal magma reservoir, leading to a pulse
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56 657 of fluid exsolution and injection of P2 magma into the shallower level Red Stock, where
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58 658 syn-P2, quartz-vein hosted mineralization formed the Red Chris Cu-Au deposit. The

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4 659 post-mineralization P3 (201.7 ± 1.2 Ma) and P4 porphyries may represent the final stages
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6 660 of intrusive activity emanating from the now degassed underlying magma chamber.
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9 661 Although both P2E and P2L porphyries are S-Cl-rich, apatites from P2L are
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11 662 marginally the most enriched. Samples of P2L are also the most mafic (SiO_2 -poor) in the
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14 663 suite, and it is therefore tempting to speculate that the source magma chamber was
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16 664 recharged with more mafic, S-Cl- H_2O -rich magma during the P2 stage. The solubility of
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19 665 S and Cl is higher in oxidized mafic magmas compared to felsic melts (> 1000 ppm S
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21 666 and > 3000 ppm Cl; Carmichael and Ghiorso, 1986; Webster, 1997; Webster et al., 1999;
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24 667 Hattori and Keith, 2001; Jugo et al., 2005; Jugo, 2009; Chelle-Michou and Chiaradia,
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26 668 2017). Therefore, a key step in porphyry ore formation at Red Chris may have been
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29 669 recharge of the batholithic system by a less evolved, although broadly cogenetic, magma
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31 670 (cf. Hattori and Keith, 2001; Steinberger et al., 2013; Large et al., 2018).
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33 671 Support for this recharge model is provided by high magmatic temperatures
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36 672 calculated from zircon grains from some P2 samples (Fig. 16), and reverse zoning
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38 673 observed in plagioclase phenocrysts from the P2E and P2L porphyries, but which is not
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41 674 observed in P3 (Fig. 14C). In detail, zircons from the P2E porphyry show the highest Ti-
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43 675 in-zircon temperatures and $\text{Eu}_N/\text{Eu}_N^*$ values, and P3 the lowest values. This might reflect
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46 676 an influx of higher temperature melt during the evolution of the P2E magma, and greater
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48 677 degrees of plagioclase fractionation from the late P3 magma. The compositional ranges
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51 678 of up to ~ 19 mol % An for plagioclase are higher than the upper limit caused by chemical
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53 679 diffusion (~ 10 mol % An; Pearce and Kolisnik, 1990). Furthermore, these reverse zoning
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56 680 patterns are matched by FeO contents in the phenocrysts (Figs. 14A and 14B), which
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58 681 suggests that the zonation reflects changes in magma composition (as opposed to simply
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4 682 changes in pressure, temperature, or water content; Ginibre and Wörner, 2007; Lange,
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6 683 2009; Ustunisik et al., 2014; Waters and Lange, 2015). We therefore interpret these
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9 684 changes to reflect late-stage (rim) growth from a more mafic magma that mixed into a
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11 685 more evolved resident magma.
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687 **Conclusions**

688 Zircon U-Pb ages indicate that the Red Stock was emplaced over a period of ~10 m.y.
689 (211.6 ± 1.3 Ma, 206.0 ± 1.2 Ma, 203.6 ± 1.8 Ma, and 201.7 ± 1.2 Ma for the P1, P2E,
690 P2L, and P3 porphyries, respectively). The ore-forming event at Red Chris occurred
691 relatively late in this magmatic history, synchronous with P2E at 206.1 ± 0.5 Ma, and
692 over a short period of <1 m.y. (as constrained by the error on the average of three Re-Os
693 analyses). The four phases of porphyry and their related magmas were chemically fairly
694 homogeneous and isotopically primitive, consistent with an island arc origin. The
695 magmas were all moderately oxidized and hydrous (~5 wt. H₂O), but the P2 magmas had
696 distinctly higher S and Cl contents. Combined with evidence for reverse zonation in
697 plagioclase phenocrysts from P2 porphyries, and more mafic compositions in P2L, we
698 suggest that the deeper source magma chamber was recharged at the time of
699 emplacement of the P2 porphyries by injection of a more mafic S-Cl-rich magma. This
700 recharge process may have triggered ore formation by causing voluminous exsolution of
701 metalliferous hydrothermal fluids.

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7
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17
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23
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34 716

35 36 717 **References**

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1200 Figure Captions

1201 Fig. 1. Major terranes in the south of the Canadian Cordillera, showing Triassic to
1202 Jurassic magmatic belts and major associated porphyry deposits in the Stikine and
1203 Quesnel terranes (modified from Nelson and Colpron, 2007; Nelson et al., 2013; original
1204 graphics file provided by Joanne Nelson, British Columbia Geological Survey of Canada).
1205 The age of the Red Chris deposit is from this study (Table 1), and the ages for the other
1206 porphyry deposits are from Brown and Kahlert (1986; Red Mountain), Mortensen et al.
1207 (1995; Mt. Polley), Scott et al. (2008; Schaft Creek), Duuring et al. (2009; Kemess
1208 South), Taseko Mines Limited (2013; Gibraltar), Bath et al. (2014; Lorraine), Byrne and
1209 Tosdal (2014; Galore Creek), Devine et al. (2014; Lorraine), Logan and Mihalynuk (2014;
1210 Highland Valley, Copper Mountaine, Afton/Ajax, and Brenda), and Jago et al. (2014; Mt.
1211 Milligan).

1212
1213 Fig. 2. Simplified geological map of the Red Stock and Red Chris Cu-Au deposit,
1214 showing the main mineralized zoned (named) and the locations of sampled drill holes
1215 (modified from Rees et al., 2015). Universal Transverse Mercator coordinates are based
1216 on the WGS84 datum.

1217
1218 Fig. 3. Hand specimen photographs of samples of the P1, P2E, P2L, and P3 porphyry
1219 intrusions, and two late basaltic to andesitic dikes. (A) P1 porphyry with anhedral to
1220 subhedral altered amphibole and plagioclase phenocrysts, crosscut by pyrite-quartz veins
1221 (sample RC13-35). (B) P2E porphyry showing crowded texture with chloritized
1222 amphibole and plagioclase phenocrysts; plagioclase grains are rimmed or replaced by

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1223 secondary K-feldspar. The brick-red color is due to fine-grained hematite (sample RC13-
1224 107). (C) P2L porphyry composed of abundant fresh amphibole and plagioclase
1225 phenocrysts, with quartz in the groundmass (sample RC13-32). (D) The P3 porphyry is
1226 similar to P2L, but is distinguished by the absence of quartz in the groundmass (sample
1227 RC13-78). (E) Andesitic dike with anhedral to subhedral amphibole phenocrysts, crosscut
1228 by a small calcite vein (sample RC13-62). (F) Basaltic dike with chloritized amphibole
1229 phenocrysts (sample RC13-106). See Digital Appendix Table A1 for sample locations.

1230

1231 Fig. 4. West-southwest–east-northeast cross section A–A’, and north-northwest–south-
1232 southwest cross section B–B’ (location of sections shown in Fig. 2), modified from
1233 Gillstrom et al. (2012) and Rees et al. (2015). Copper equivalent-grade zones, drill holes,
1234 and the boundary between potassic and post-potassic zones are shown: Cu equivalent (%)
1235 = $\text{Cu (\%)} + 0.486 \times \text{Au (g/t)}$.

1236

1237 Fig. 5. Hand specimen photographs and photomicrographs of alteration and vein minerals.
1238 (A, B) P2E porphyry (samples RC13-81 and RC13-75) with potassic alteration and A-
1239 type quartz veins comprising magnetite, secondary K-feldspar, and disseminated bornite
1240 and pyrite. Potassic alteration is characterized by secondary K-feldspar veins and
1241 selvages around A-veins. Late unmineralized carbonate veins cut the A veins. (C)
1242 Amphibole phenocryst altered to secondary biotite, which has then been altered to
1243 chlorite, reflecting potassic alteration overprinted by chlorite–sericite alteration (plane-
1244 polarized transmitted light; P2E: sample RC13-30). (D) Plagioclase phenocryst partially
1245 overprinted by sericite (cross-polarized transmitted light; P2E porphyry: sample RC13-

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1246 11). (E) B-type quartz veins with centerline pyrite in P2L porphyry (sample RC13-44). (F)
1247 Pyritic D vein in P1 porphyry (sample RC13-40). (G) High-grade ore in sheeted A-type
1248 quartz-chalcopyrite-K-feldspar veins (sample RC13-31 in P2E porphyry). (H) Quartz-
1249 carbonate-pyrite-molybdenite-chalcopyrite vein in P2E porphyry (reflected light; sample
1250 RC13-88). Abbreviations: Amp = amphibole; Bi = biotite; Bn = bornite; Cbn = carbonate;
1251 Chl = chlorite; Cpy = chalcopyrite; Kfs = K-feldspar; Mo = molybdenite; Mt = magnetite;
1252 Pl = plagioclase; Py = pyrite; Qtz = quartz. See Digital Appendix Table A1 for sample
1253 locations.

1254
1255 Fig. 6. Zircon U–Pb Tera-Wasserburg concordia diagrams for (A) P1, (B) P2E, (C) P2L,
1256 and (D) P3 porphyry samples dated by LA-MC-ICP-MS. Uncertainty ellipses and
1257 calculated ages are shown at 2σ .

1258
1259 Fig. 7. Weighted mean Re-Os model age of three molybdenite vein samples from the Red
1260 Chris Cu-Au deposit.

1261
1262 Fig. 8. Zr/Ti vs. Nb/Y discrimination diagram (Winchester and Floyd, 1977) for porphyry
1263 and basaltic–andesitic dike samples from Red Chris.

1264
1265 Fig. 9. Selected whole-rock major element variation diagrams for porphyry and basaltic–
1266 andesitic dike samples from Red Chris: (A) K₂O, (B) Na₂O, (C) TiO₂, (D) Al₂O₃, (E)
1267 total Fe₂O₃, and (F) MgO vs. SiO₂.

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1269 Fig. 10. (A) Primitive mantle-normalized trace element, and (B) chondrite-normalized
1270 rare earth element diagrams for porphyry and basaltic–andesitic dike samples from Red
1271 Chris. The normalization values for primitive mantle chondrite are from Sun and
1272 McDonough (1989).

1273

1274 Fig. 11. $\epsilon_{Nd}(t)$ vs. initial $^{87}Sr/^{86}Sr$ ratios for porphyry and basaltic–andesitic dike samples
1275 from Red Chris, calculated at $t = 200$ Ma. All the samples fall in the field of Mesozoic
1276 igneous rocks in the Stikine island arc terrane, clearly different from the Late Cretaceous
1277 to Eocene plutons in the Northern Coast Plutonic Complex, which were derived from
1278 evolved crust. The depleted MORB mantle field is from Pilet et al. (2011); the Stikinia
1279 Mesozoic igneous rock field is from Samson et al. (1989); the Northern Coast Plutonic
1280 Complex field is from Samson et al. (1991) and Patchett et al. (1998); all data are re-
1281 calculated at 200 Ma.

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1283 Fig. 12. Histogram and relative probability curve for zircon $\epsilon_{Hf}(t)$ values from (A) P1, (B)
1284 P2E, (C) P2L, and (D) P3 porphyry samples.

1285

1286 Fig. 13. Classification diagram and plots of oxidation state, temperature, pressure, and
1287 magmatic water content estimated from amphibole compositions from P2E and P2L
1288 porphyry samples at Red Chris. A. $C(Al^{VI} + Fe^{3+} + 2Ti^{4+})$ (apfu) vs. $A(Na^{+} + K^{+})$ (apfu). B.
1289 ΔFMQ vs. temperature. C. ΔFMQ vs. pressure. D. ΔFMQ vs. magmatic water content.

1290 The classification diagram for calcic amphibole is given by the Excel spreadsheet of

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1291 Locock (2014), and the superscript C and A represent C and A cations following the
1292 general amphibole formula ($AB_2C_5T_8O_{22}W_2$), respectively. Note that potassic-magnesian-
1293 hastingsite is included in the field of magnesian-hastingsite. ΔFMQ values, temperatures,
1294 and magmatic water content were calculated from the spreadsheet of Ridolfi et al. (2010).
1295 Amphibole crystallization pressures were calculated using the equation of Mutch et al.
1296 (2016). The ΔFMQ values were calculated following the equation of Myers and Eugster
1297 (1983): $\log fO_2 = -24,441.9/T (K) + 8.290 (\pm 0.167)$. Abbreviation: apfu = atoms per
1298 formula unit.

1299
1300 Fig. 14. Photomicrographs (cross-polarized transmitted light), Backscattered electron
1301 (BSE) images, and electron microprobe analysis profiles for FeO and anorthite proportion
1302 (X_{An}) for representative plagioclase crystals from the P2E, P2L, and P3 porphyries at Red
1303 Chris. Red circles on photomicrographs and white circles on BSE images denote the
1304 analyzed spots. The error bars for X_{An} and FeO analyses are smaller than the size of the
1305 symbols. Abbreviations: Amp = amphibole; Ap = apatite; Pl = plagioclase; Ser = sericite.

1306
1307 Fig. 15. Chondrite-normalized REE patterns for zircons from (A) P1, (B), P2E, (C) P2L,
1308 and (D) P3 porphyry samples from Red Chris. Normalization values are from Sun and
1309 McDonough (1989).

1310
1311 Fig. 16. Zircon Eu_N/Eu_N^* vs. temperature diagram. Eu_N/Eu_N^* is the europium anomaly,
1312 calculated as $Eu_N/Eu_N^* = Eu_N/(Sm_N \times Gd_N)^{0.5}$, using the chondrite normalization values

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1313 of Sun and McDonough (1989). Oxidized suites have zircon Eu_N/Eu_N^* values >0.4
1314 (Dilles et al., 2015).

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1316 Fig. 17. Backscattered electron images of apatite crystals in samples from (A) P1 (RC13-
1317 39), (B) P2E (RC13-107), (C) P2L (RC13-33), and (D) P3 (RC13-78). Concentrations of
1318 SO_3 and Cl in apatite crystals are shown in wt. % (SO_3/Cl); red circles represent the
1319 analyzed spots. Higher concentrations are observed in apatites from P2E and P2L; some
1320 apatite microphenocrysts from P2 porphyries show zoning from SO_3 -Cl-rich cores to
1321 SO_3 -Cl-poorer rims (C).

1322

1323 Fig. 18. Plots of (A) S, and (B) F vs. Cl contents for apatite microphenocrysts from P1 to
1324 P3 porphyry samples at Red Chris. Abbreviation: apfu = atoms per formula unit. Data
1325 from Digital Appendix Table A8.

Figure 1

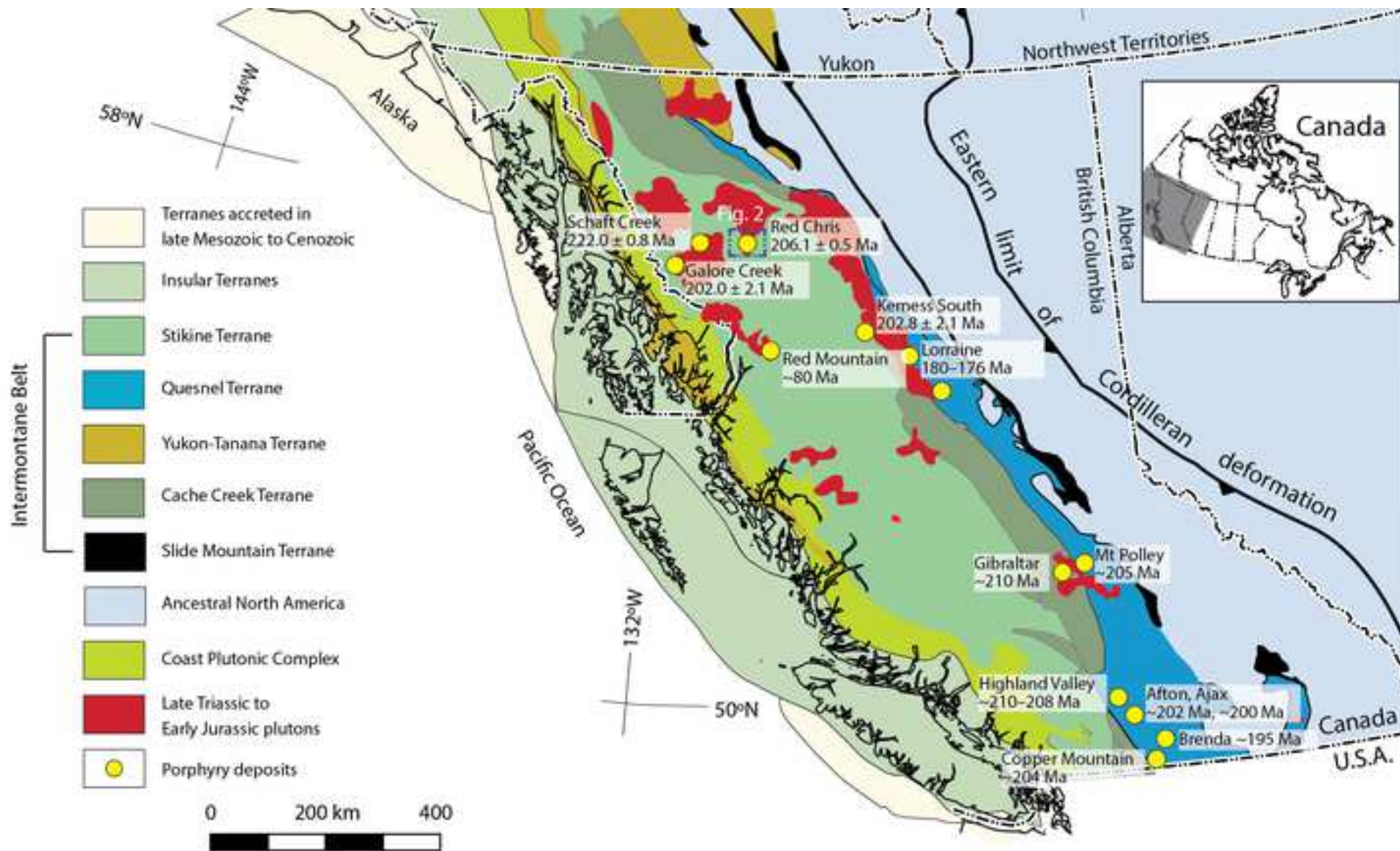
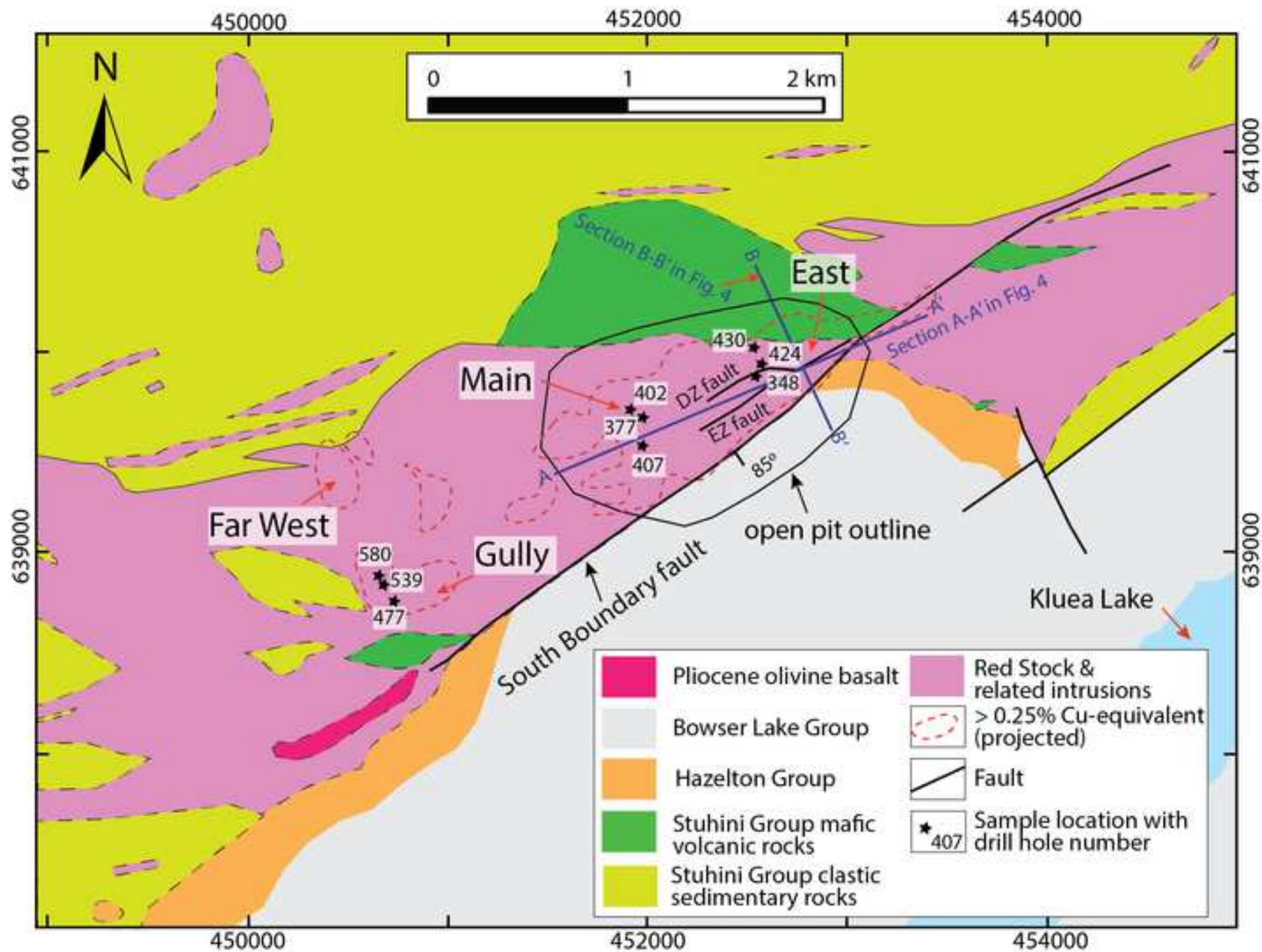
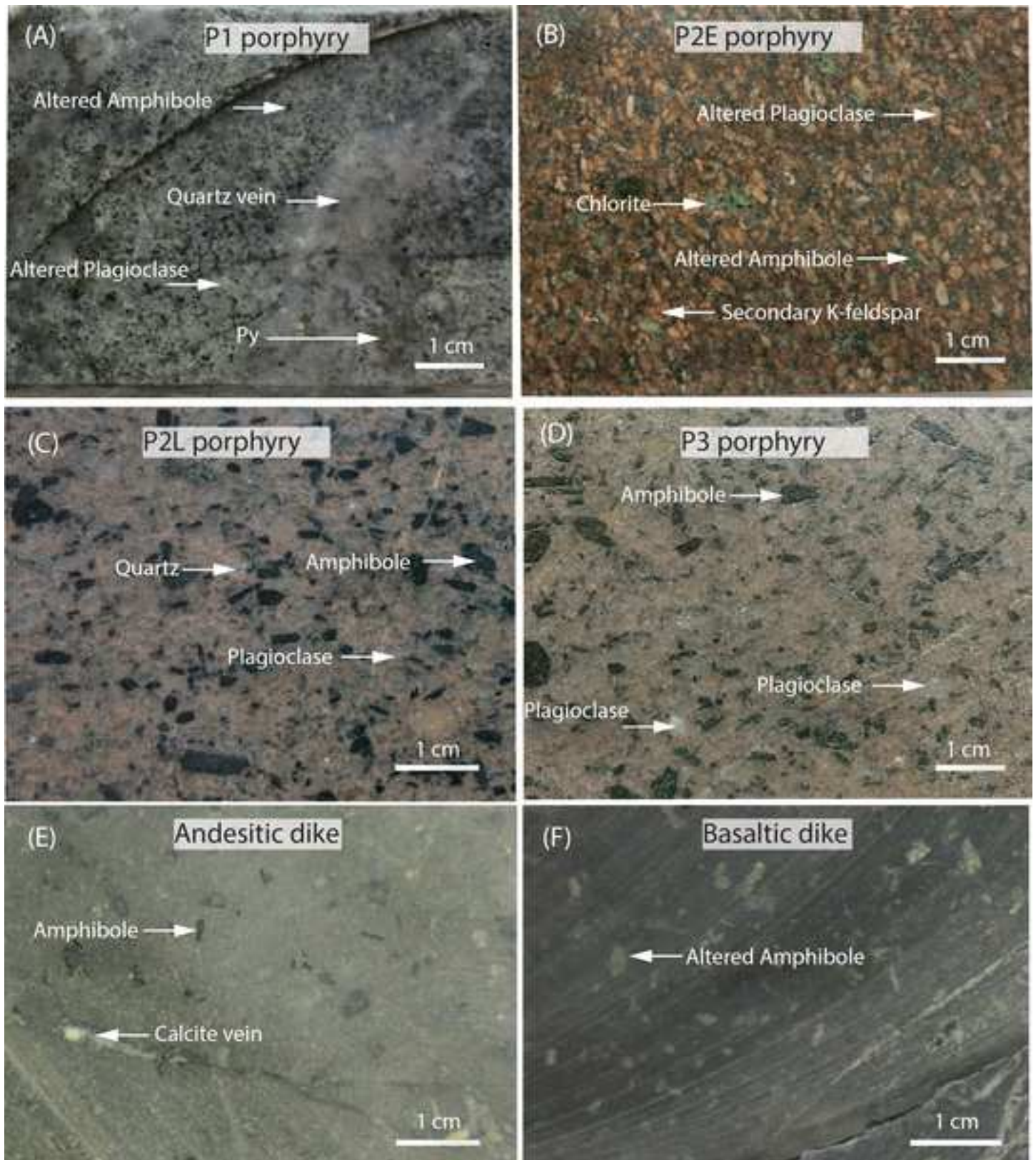
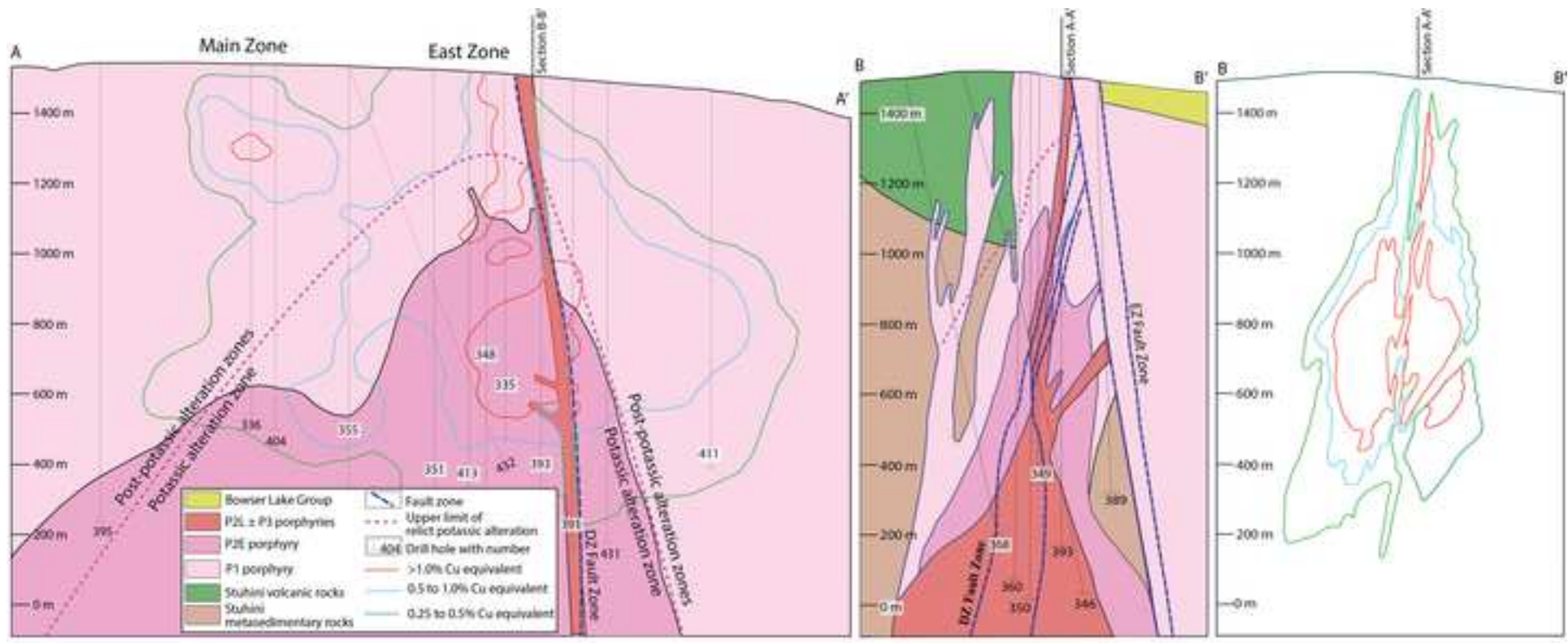
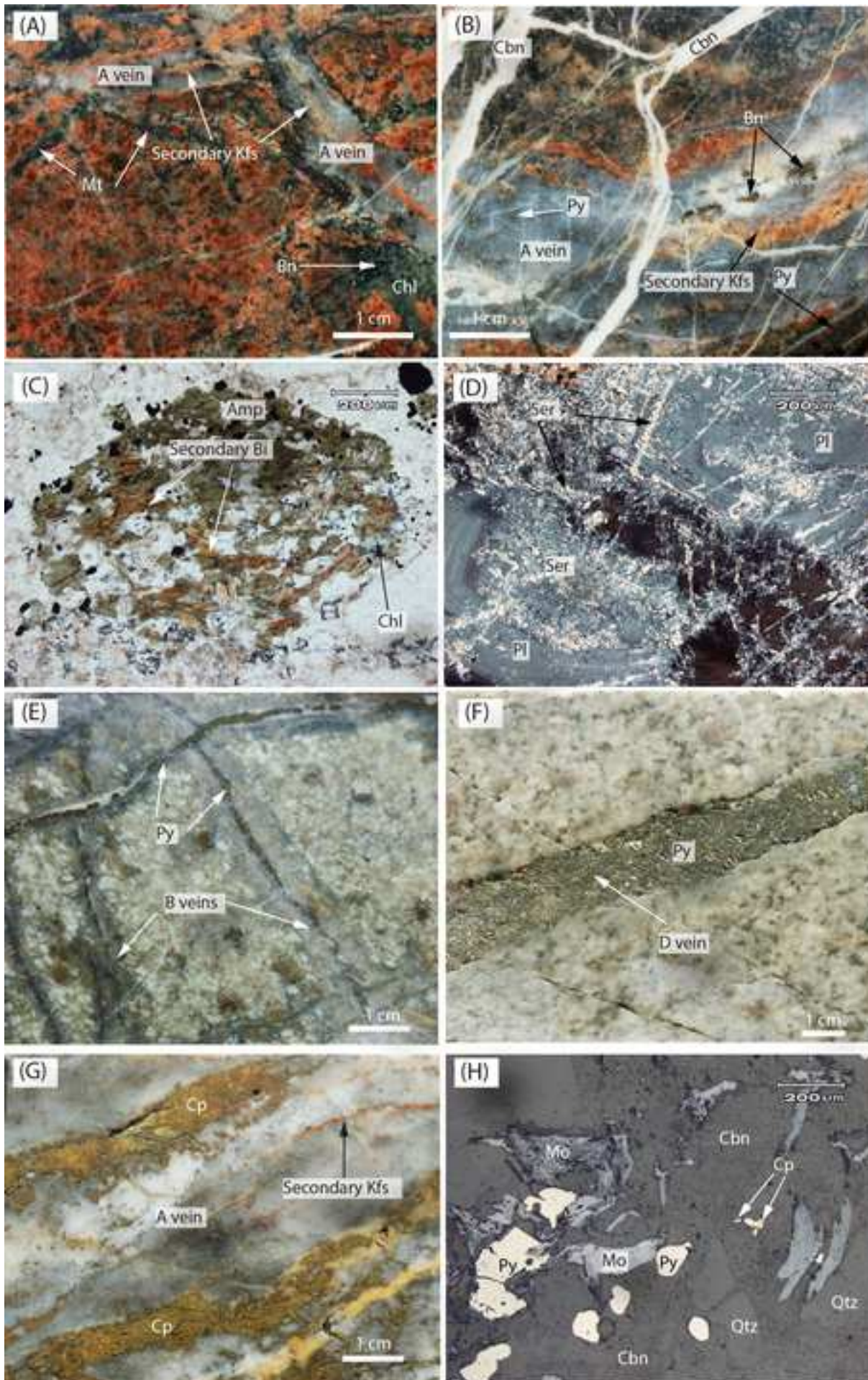


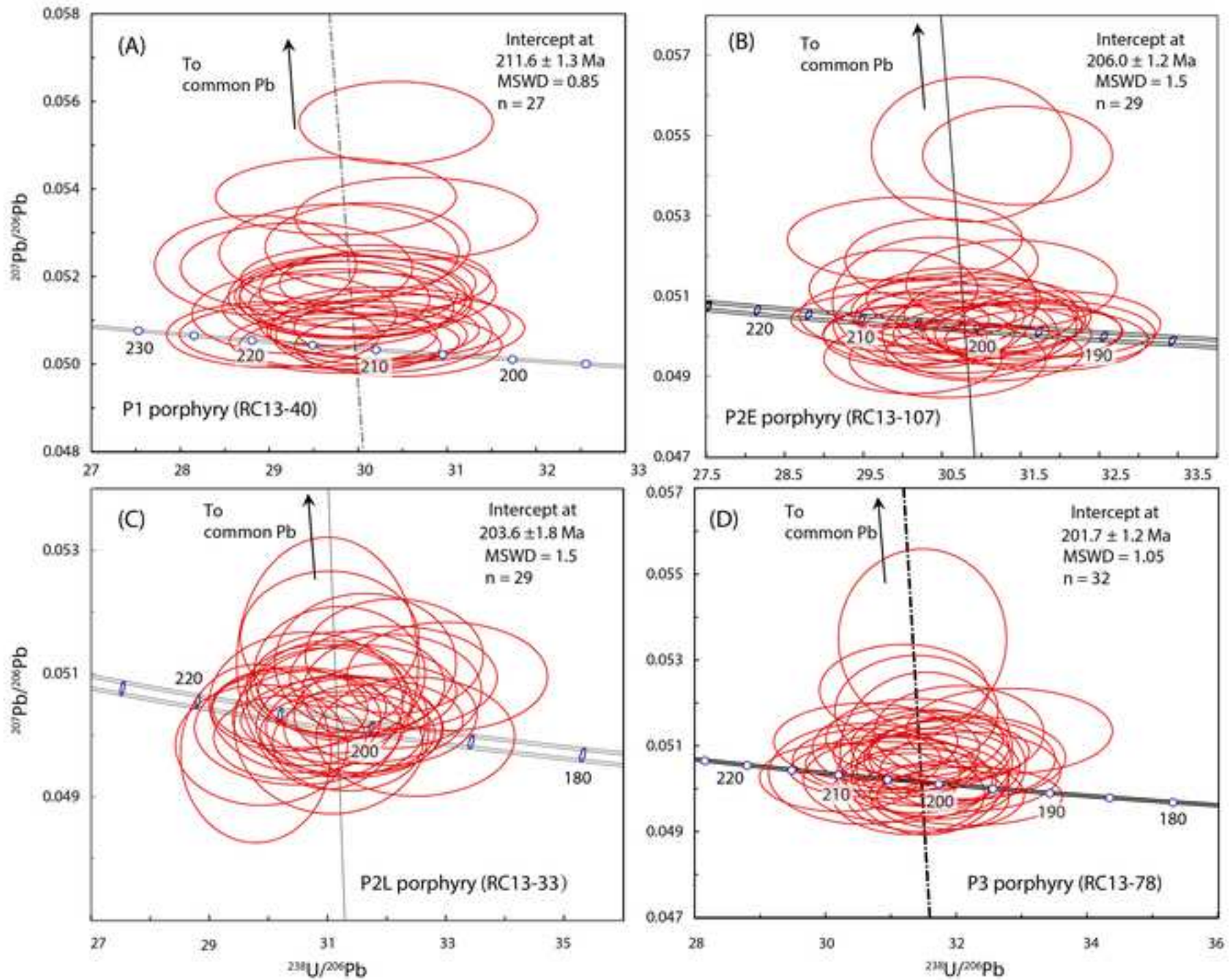
Figure 2











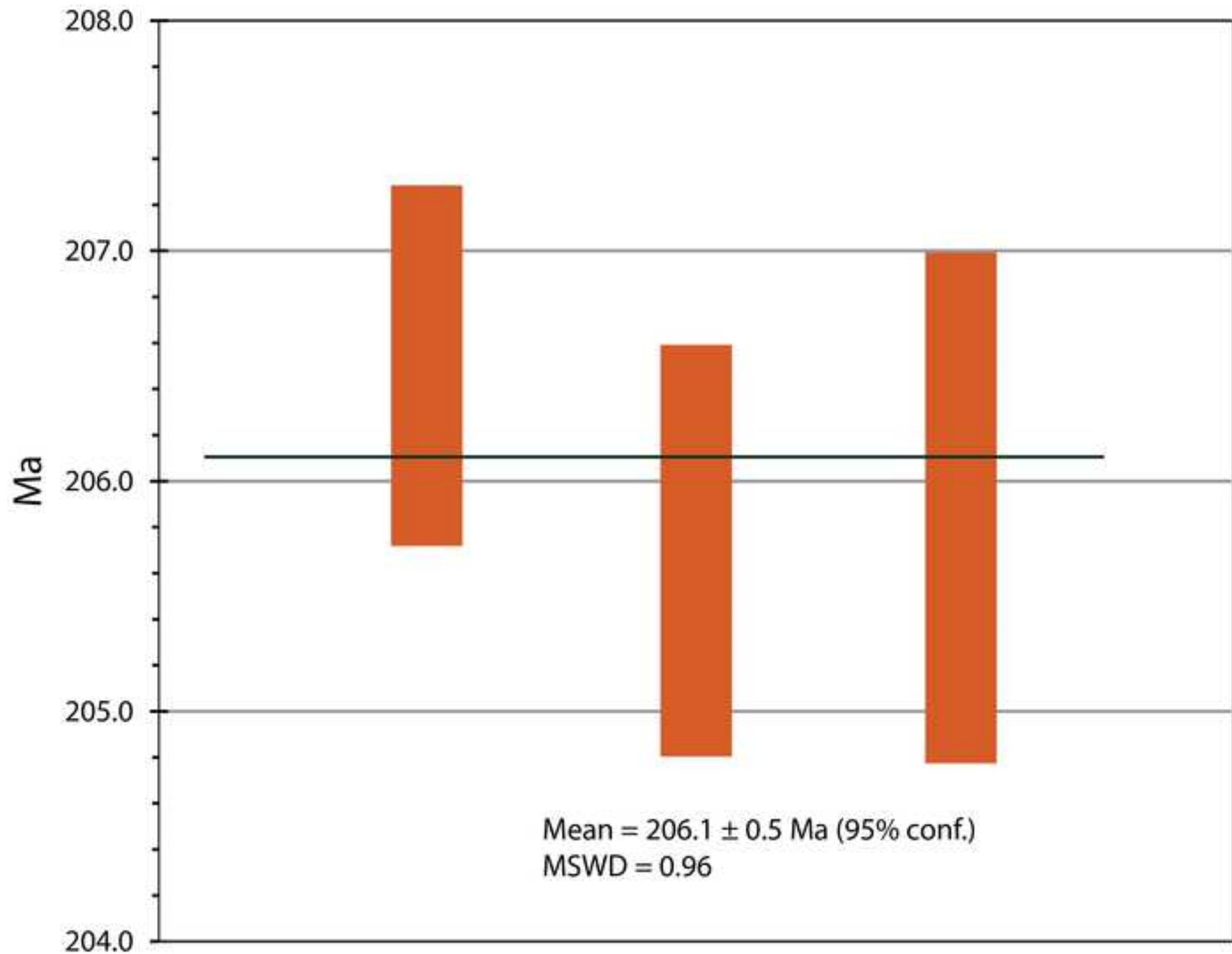
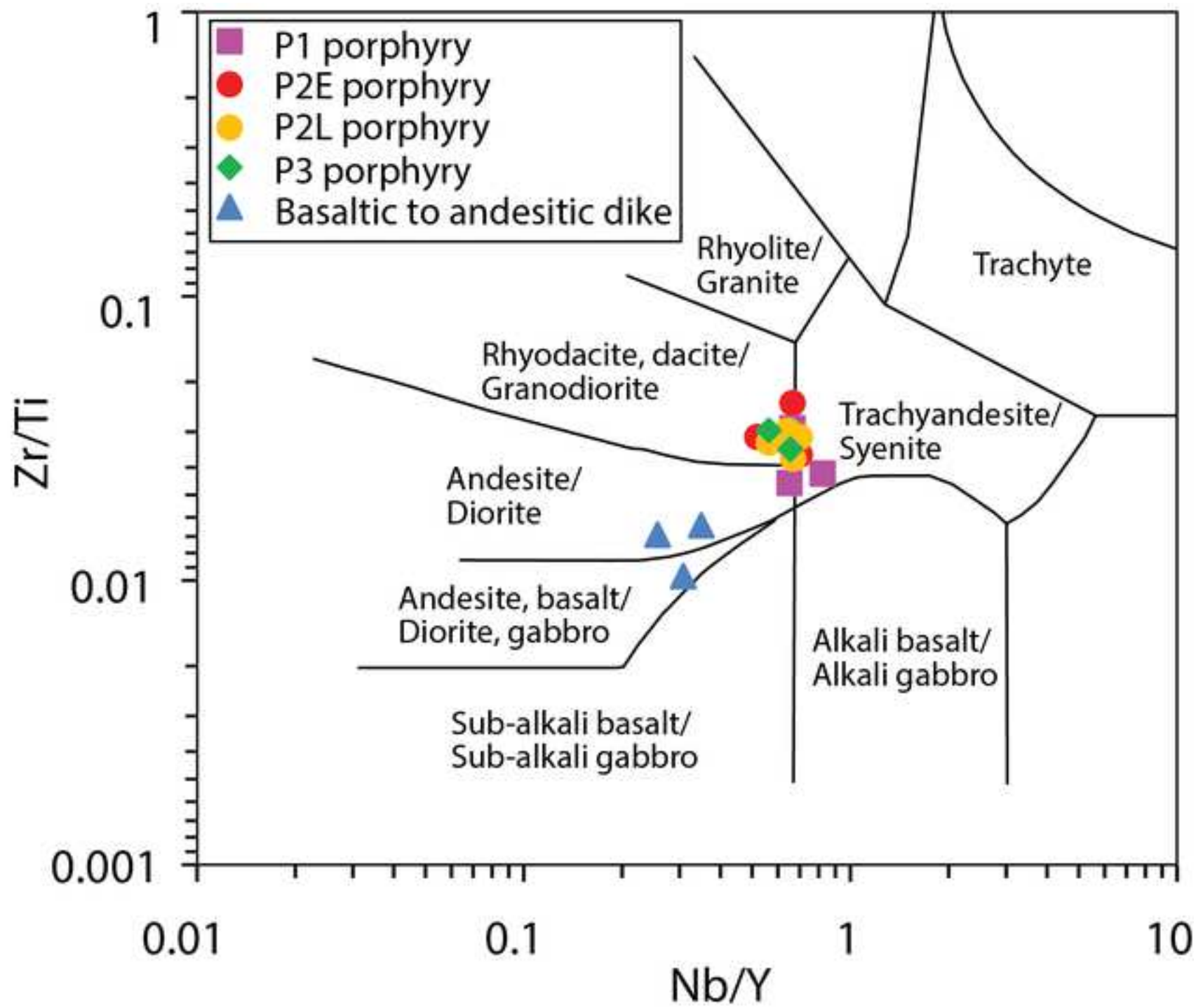
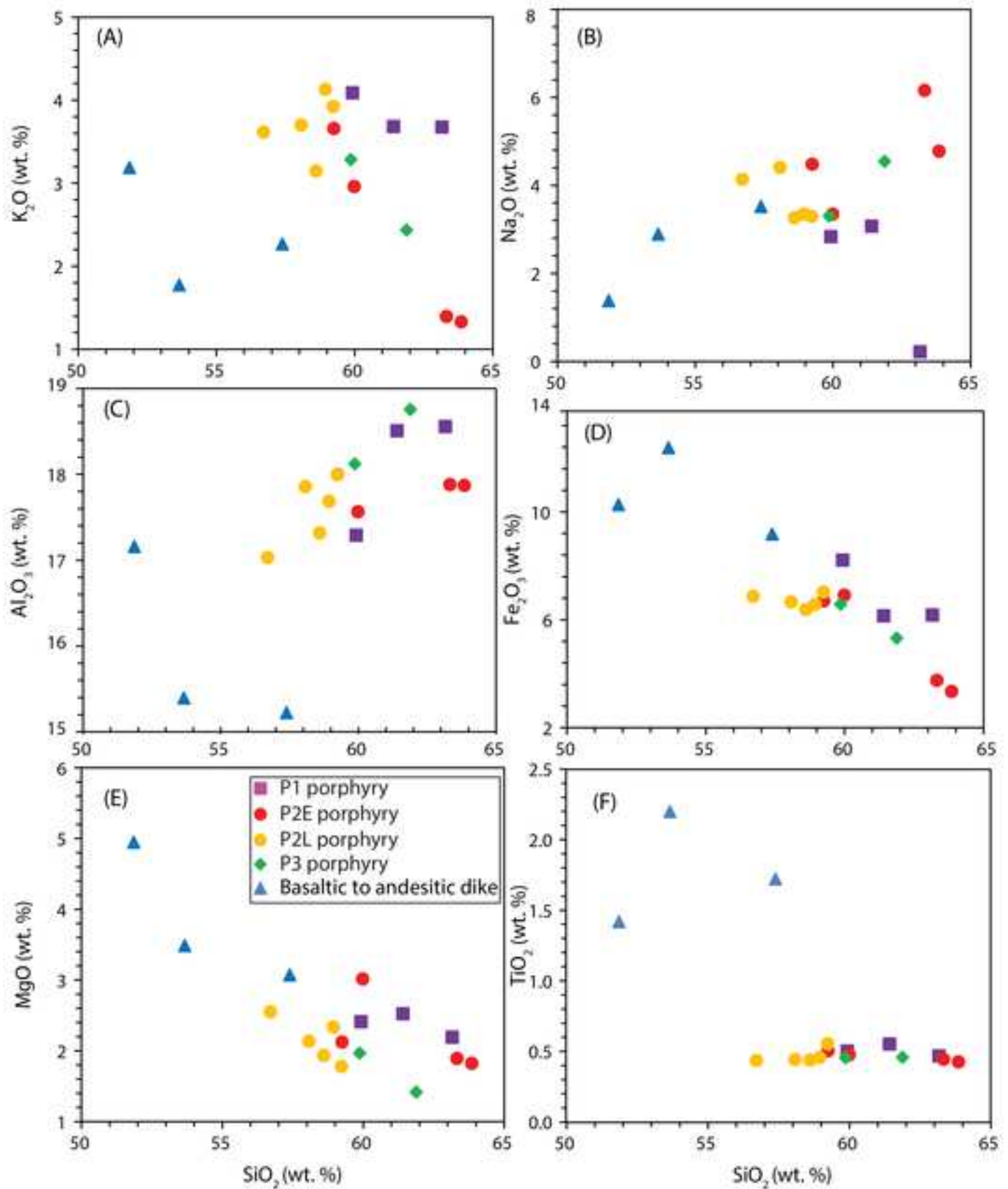
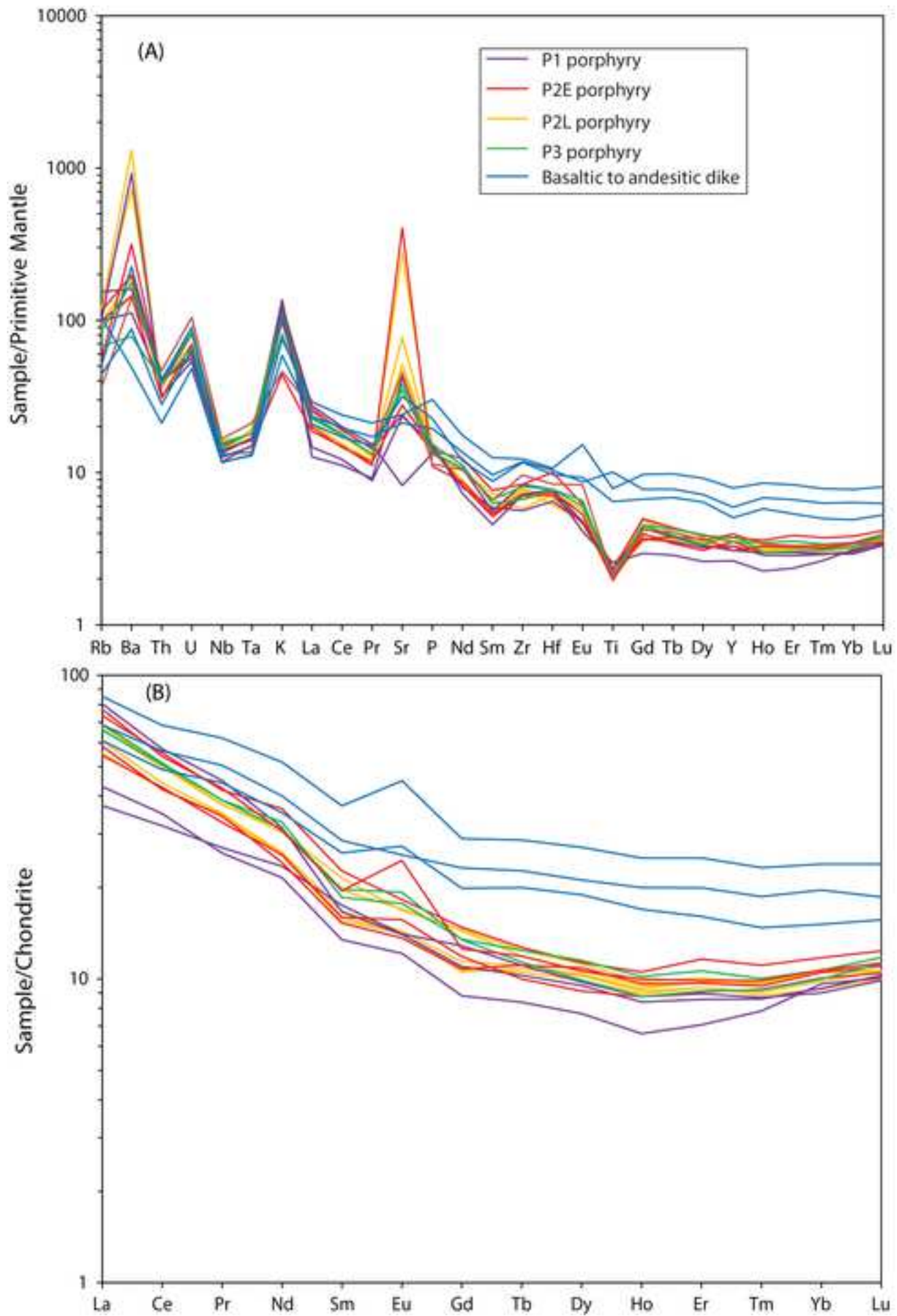
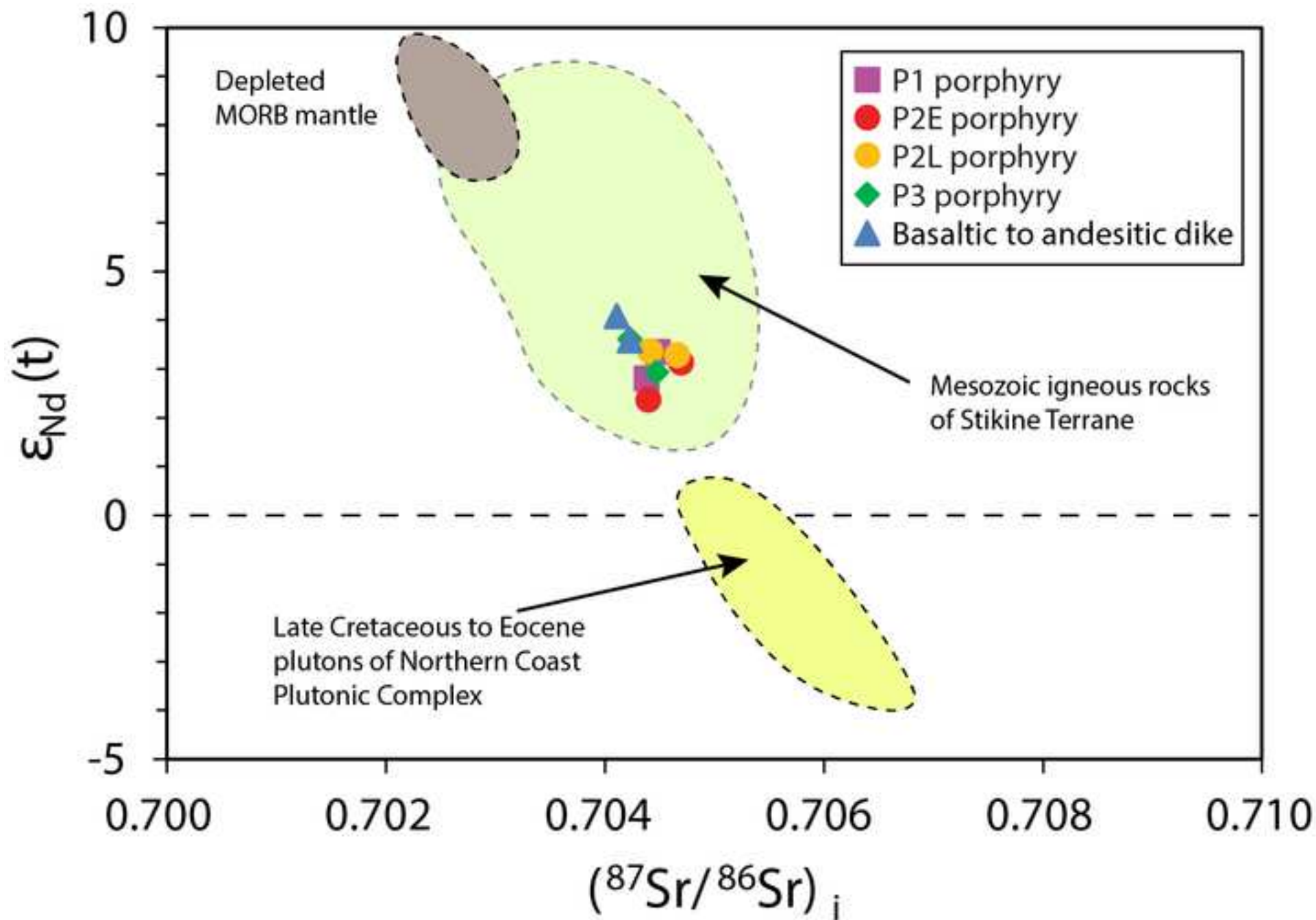


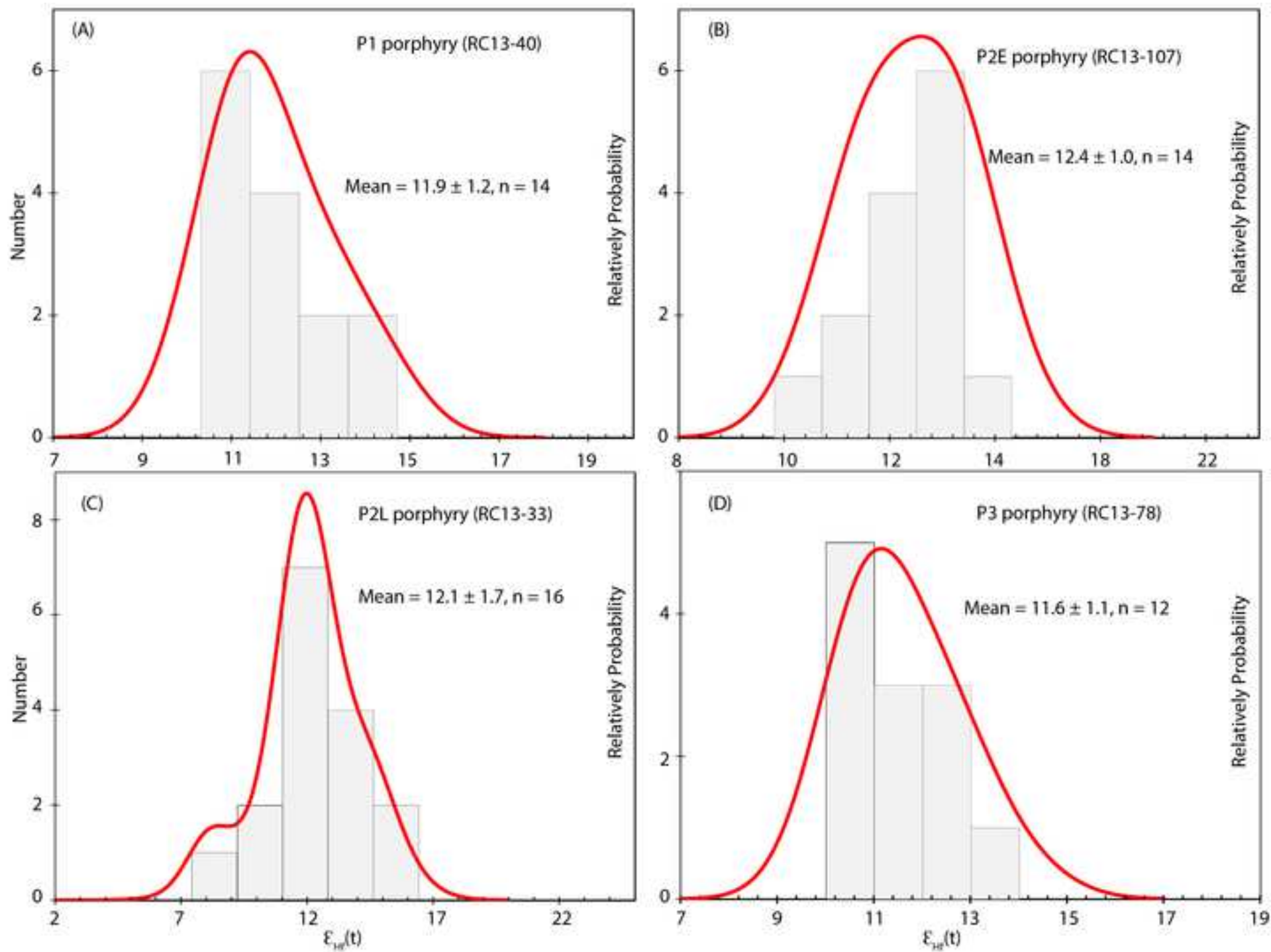
Figure 8

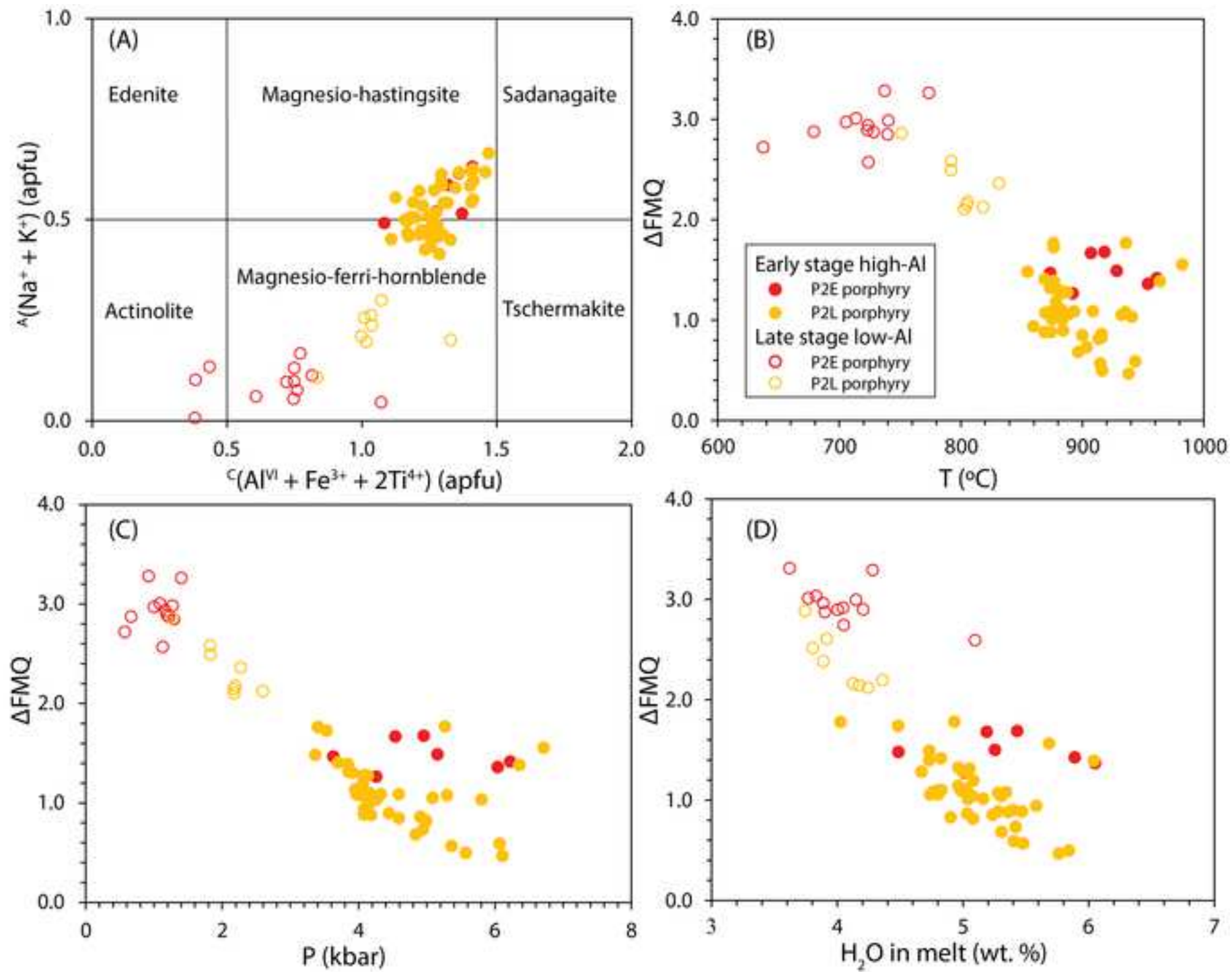


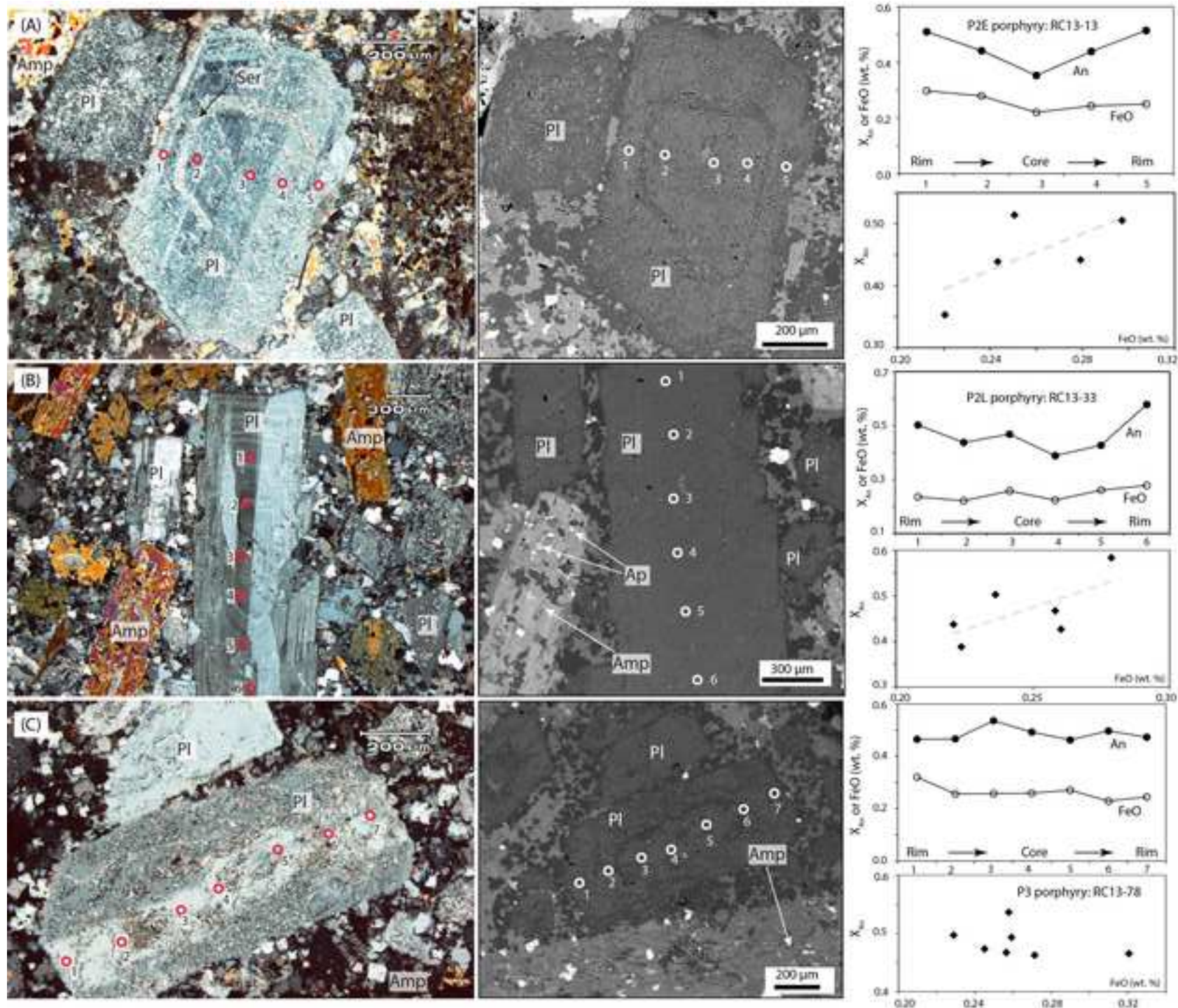


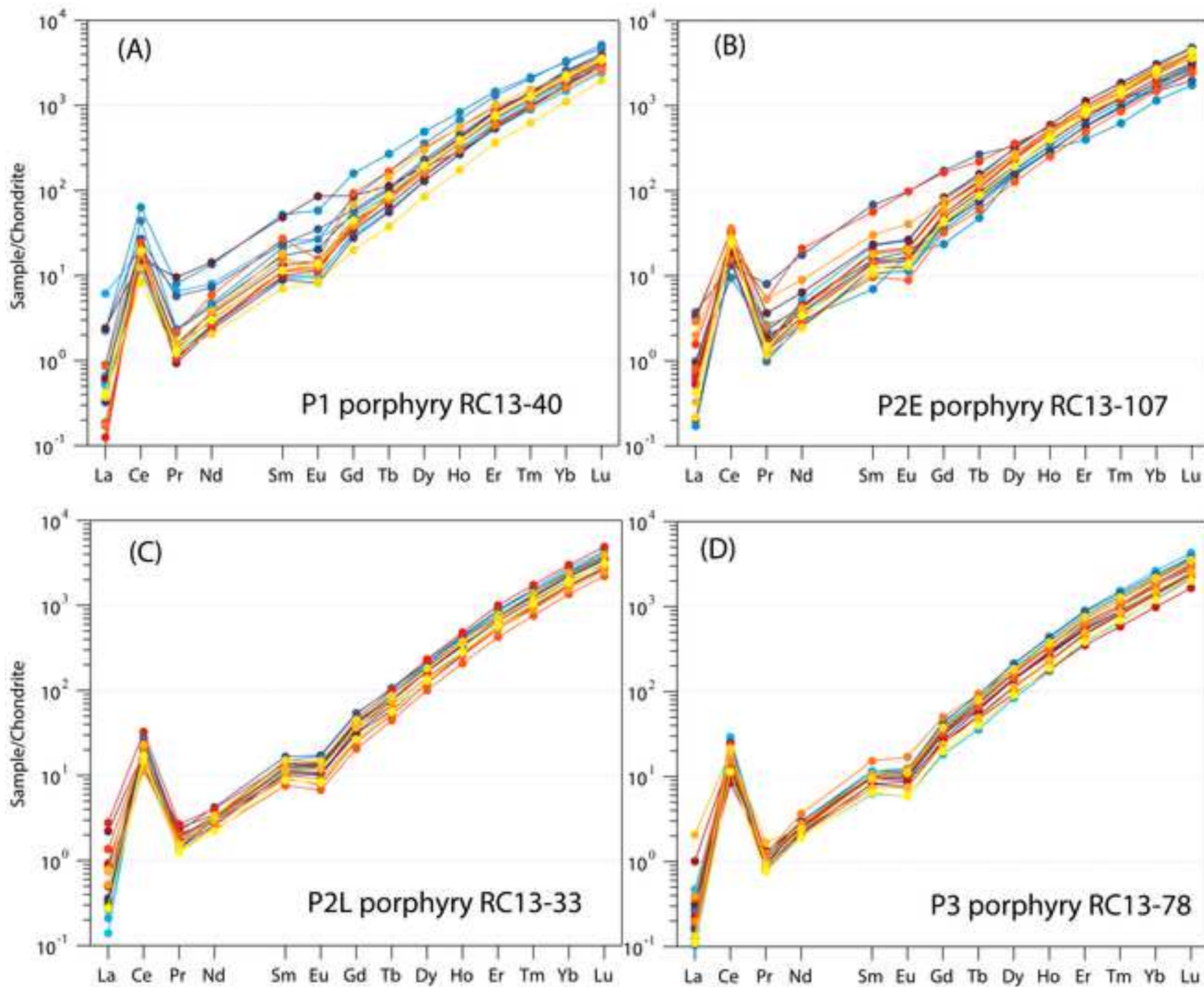


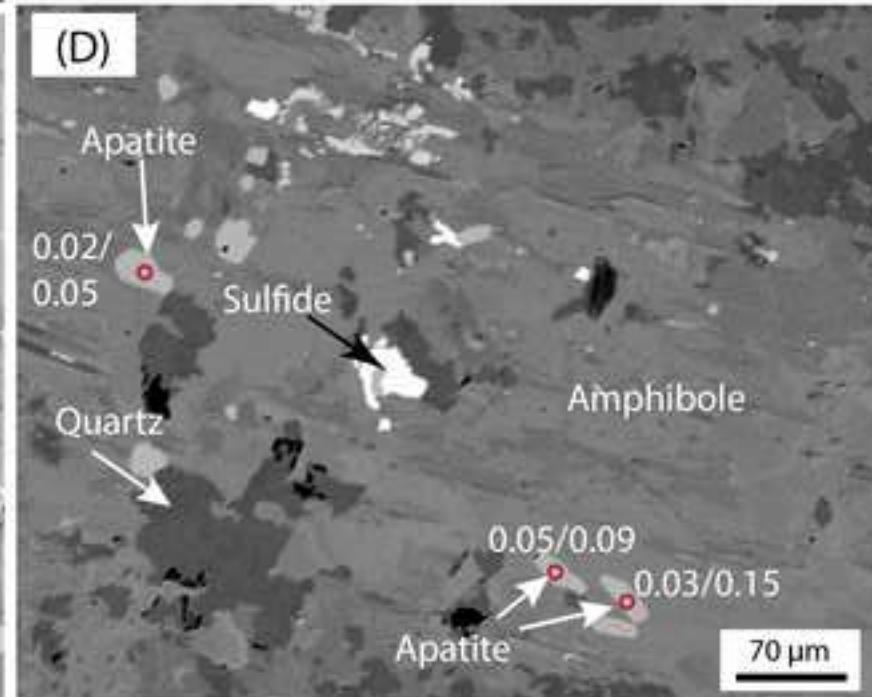
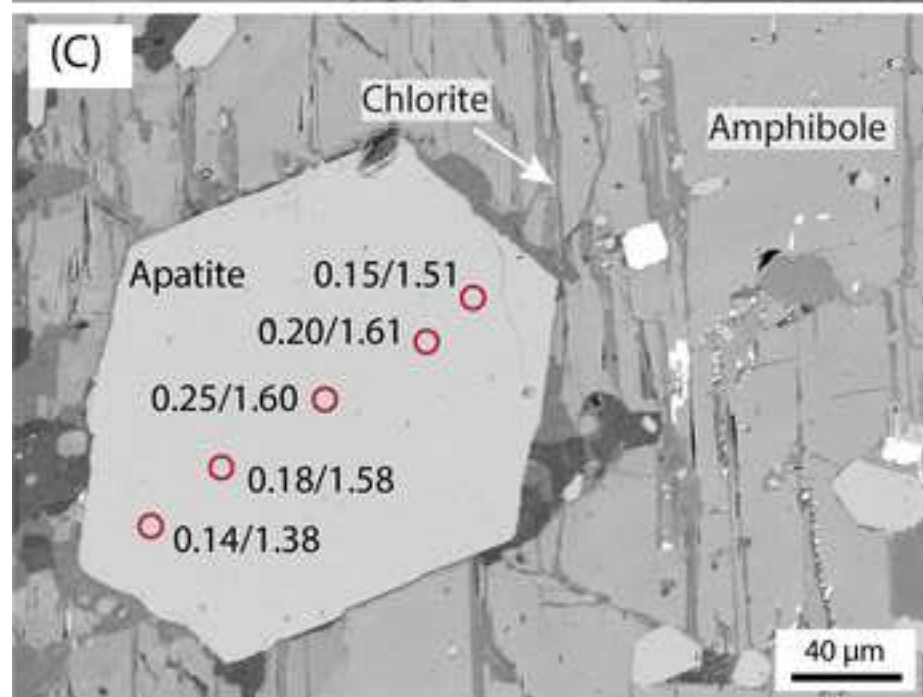
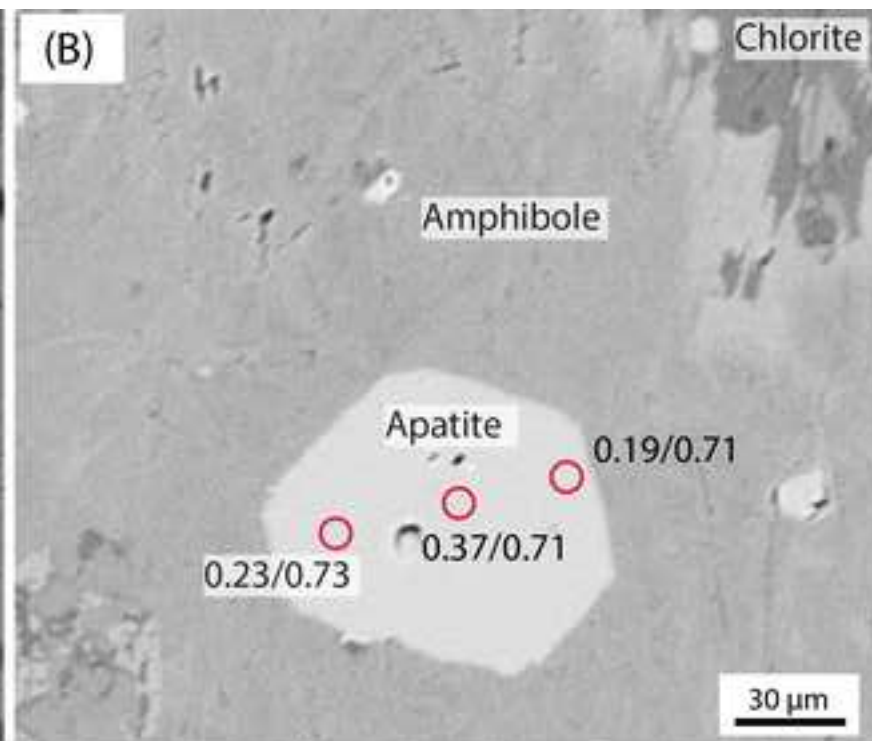
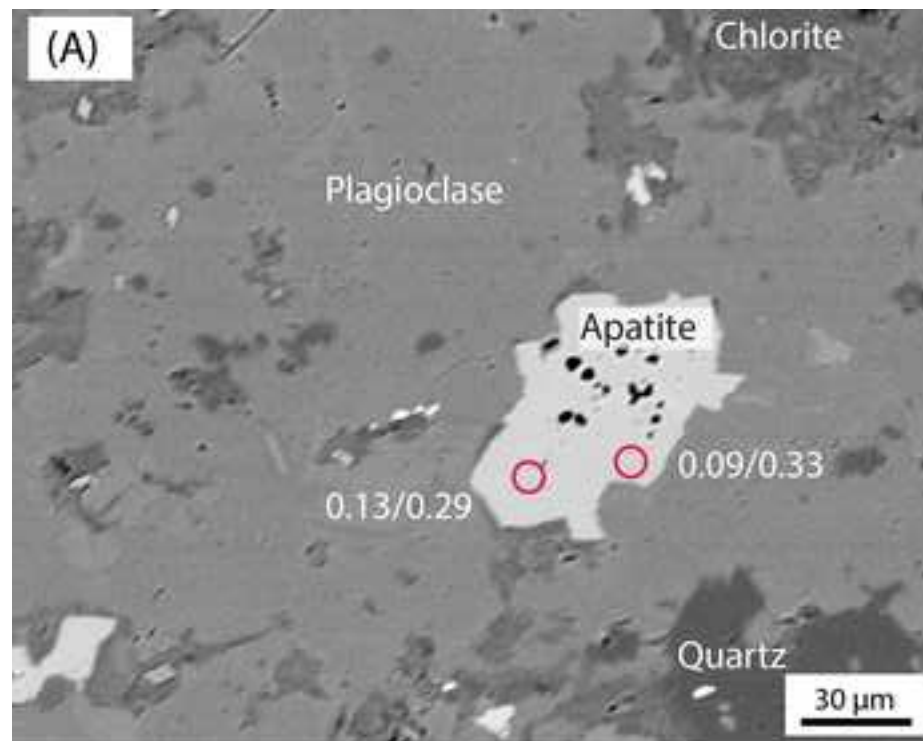












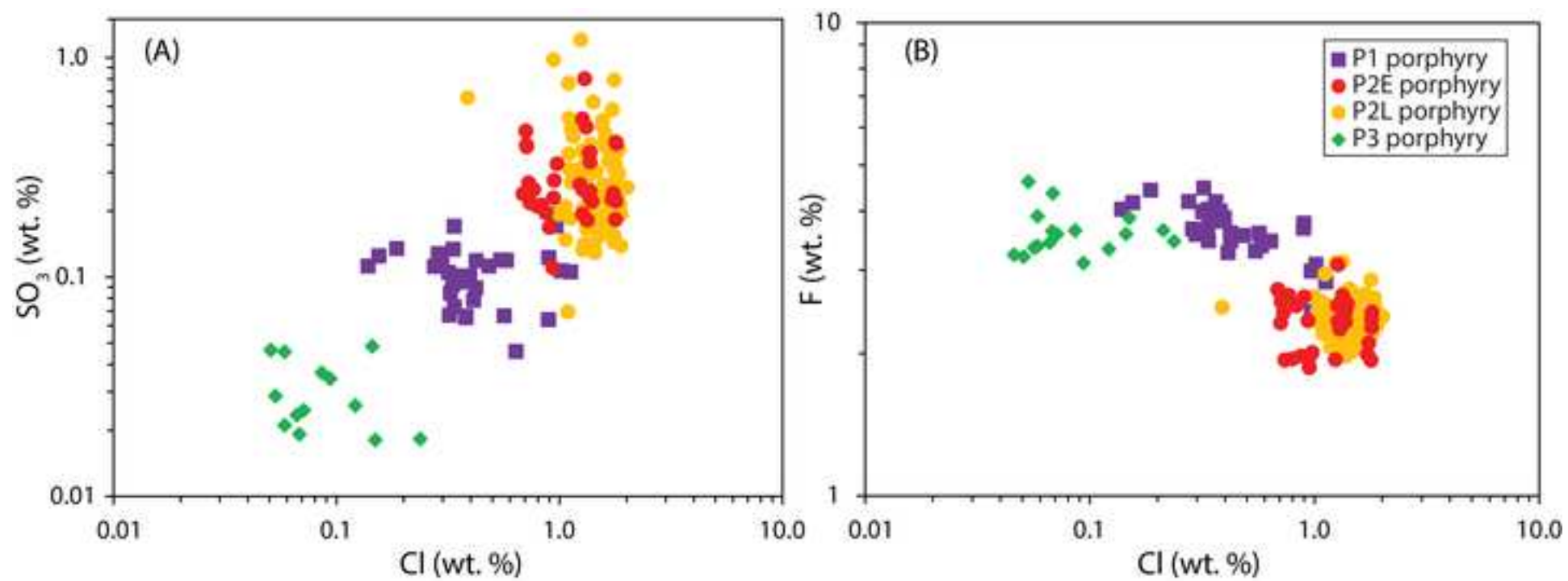


Table 1. Re-Os Isotope Analyses for molybdenite from the Red Chris Cu-Au deposit

Sample	Description	Re ($\mu\text{g/g}$)	$\pm 2\sigma$	^{187}Re ($\mu\text{g/g}$)	$\pm 2\sigma$	^{187}Os (ng/g)	$\pm 2\sigma$	Model Age (Ma)
RC13-82	Quartz-carbonate-pyrite-	497.8	1.3	312.9	0.8	1078	0.0	206.5
RC13-88	molybdenite-chalcopryrite	1771	5.0	1113	3.0	3821	2.0	205.7
RC13-103	vein in P2E porphyry	1124	3.0	706	1.8	2427	4.0	205.9

Note: See the Digital Appendix Table A1 for sample locations.

$\pm 2\sigma$
(Ma)
0.8
0.9
1.1

Table 2. Major and trace element compositions of porphyry and basaltic to andesitic dike rocks

Sample # - Detection	RC13-39	RC13-40	RC13-63	RC13-37	RC13-42	RC13-92	RC13-107	RC13-25	RC13-26	
Lithology	↓	P1	P1	P1	P2E	P2E	P2E	P2E	P2L	P2L
<i>Weight %</i>										
SiO ₂	0.01	55.43	56.57	58.25	55.54	58.61	61.29	56.4	53.86	56.05
Al ₂ O ₃	0.01	15.94	16.57	17.5	16.21	16.35	17.25	17.08	16.12	17.18
Fe ₂ O ₃ (T)	0.01	7.59	5.53	5.83	6.4	3.06	3.63	6.38	6.52	6.42
MnO	0.001	0.067	0.079	0.107	0.135	0.028	0.057	0.152	0.134	0.125
MgO	0.01	2.23	1.96	2.39	2.79	1.67	1.83	2.02	2.42	2.06
CaO	0.01	2.54	3.12	2.11	5.53	3.86	3.63	3.47	4.78	4.57
Na ₂ O	0.01	2.62	0.19	2.91	3.1	4.38	5.96	4.26	3.93	4.25
K ₂ O	0.01	3.78	3.29	3.49	2.74	1.22	1.35	3.48	3.43	3.57
TiO ₂	0.001	0.463	0.419	0.523	0.441	0.39	0.429	0.48	0.414	0.426
P ₂ O ₅	0.01	0.29	0.26	0.32	0.3	0.22	0.24	0.29	0.3	0.29
LOI		7.51	10.44	5.15	7.42	8.22	3.22	4.81	5.03	3.49
S	0.001	4.63	3.65	0.386	0.026	1.63	1.19	1.22	0.251	0.157
Total		98.45	98.43	98.57	100.6	98.02	98.9	98.81	96.94	98.43
<i>Parts per million</i>										
Ba	1	6440	783	1123	1012	2201	980	1382	9182	5082
Co	0.1	10.5	9.4	16.2	12.7	10.6	8.3	9.3	11.4	12
Cr	0.5	24.9	30.2	26.3	115	50.1	19.4	26	31.2	32.8
Cs	0.1	3.7	2.1	4.2	6.7	4.5	0.8	1.1	4.7	4.9
Ga	1	17	14	17	15	15	15	17	15	17
Ge	0.5	3.2	2.4	1.4	2	3.1	2	1.8	1.7	1.9
Hf	0.1	2	2.4	2.4	2.2	2.3	2.6	3.1	2.2	2.2
Nb	0.2	9	9.6	10	9.9	8.4	11.9	10.8	10.1	11.5
Ni	1	5	7	32	23	3	7	3	5	6
Pb	5	14	8 < 5	< 5	< 5	< 5	< 5	< 5	< 5	< 5
Rb	1	64	63	98	63	35	23	73	74	76
Sc	0.01	19.9	11.8	15	17.6	10.5	9.78	13	15.9	15.9
Sr	2	884	174	493	586	8516	944	508	5880	1639
Ta	0.01	0.61	0.69	0.67	0.68	0.62	0.87	0.74	0.67	0.73
Zr	1	63	94	80	79	81	108	94	85	65
Hf	0.1	2	2.4	2.4	2.2	2.3	2.6	3.1	2.2	2.2
U	0.01	1.13	1.72	1.19	1.46	1.26	2.2	1.75	1.31	1.69
Th	0.05	2.71	3.23	3.38	2.64	3.41	3.97	3.52	2.63	3.2
V	5	147	105	134	147	109	110	114	135	135
Y	1	14	15	12	14	16	18	17	14	16
Zn	1	724	123	58	101	29	32	56	72	79
Zr	1	63	94	80	79	81	108	94	85	65
Tl	0.05	0.59	1.02	0.6	0.51	0.17	0.12	0.12	0.28	0.06
La	0.05	8.74	18.9	10.1	13.7	12.8	18.1	17.3	13.7	15.6
Ce	0.05	19.8	35.5	21.7	26.1	26.5	33.5	34.4	26	30.9
Pr	0.01	2.54	4.23	2.44	3.23	3.08	3.99	3.93	3.24	3.52
Nd	0.05	10.8	14.5	9.9	11.1	11.8	14.1	16.7	11.7	14.4
Sm	0.01	2.62	2.49	2.02	2.29	2.39	2.93	3.4	2.33	2.94
Eu	0.005	0.801	0.799	0.691	0.778	0.893	1.4	1.04	0.805	0.968
Gd	0.01	2.57	2.2	1.76	2.16	2.37	2.51	2.96	2.11	2.88
Tb	0.01	0.41	0.38	0.31	0.41	0.37	0.44	0.47	0.4	0.44
Dy	0.01	2.44	2.38	1.92	2.72	2.28	2.65	2.83	2.56	2.67
Ho	0.01	0.49	0.47	0.37	0.54	0.49	0.56	0.59	0.51	0.53
Er	0.01	1.43	1.37	1.13	1.55	1.45	1.58	1.86	1.5	1.55
Tm	0.005	0.217	0.215	0.196	0.237	0.231	0.247	0.277	0.221	0.241
Yb	0.01	1.44	1.49	1.54	1.68	1.61	1.71	1.88	1.48	1.72

Lu	0.002	0.247	0.257	0.252	0.276	0.261	0.282	0.309	0.25	0.266
La/Yb		6.069	12.685	6.558	8.155	7.950	10.585	9.202	9.257	9.070
Sm/Y		63.143	11.600	41.083	41.857	532.250	52.444	29.882	420.000	102.438
Nb/Y		0.643	0.640	0.833	0.707	0.525	0.661	0.635	0.721	0.719
Zr/Ti		0.021	0.034	0.024	0.028	0.032	0.041	0.031	0.033	0.025
V/Sc		7.387	8.898	8.933	8.352	10.381	11.247	8.769	8.491	8.491
Eu _N /Eu _N *		0.938	1.037	1.114	1.063	1.140	1.569	0.996	1.103	1.011

Note: See the Digital Appendix Table A1 for sample locations.

ocks at Red Chris

RC13-32	RC13-33	RC13-83	RC13-77	RC13-78	RC13-62	RC13-79	RC13-97
P2L	P2L	P2L	P3	P3	Dike	Dike	Dike
57.57	56.05	57.38	58.47	57.45	46.23	49.38	54.17
17.22	16.51	17.38	17.67	17.34	15.25	14.12	14.32
6.41	6.09	6.8	5.02	6.31	9.15	11.39	8.66
0.13	0.107	0.116	0.066	0.11	0.108	0.168	0.19
2.28	1.85	1.72	1.34	1.89	4.41	3.21	2.9
5.66	6.6	6.04	4.84	6.1	7.78	7.3	5.72
3.27	3.12	3.19	4.29	3.17	1.23	2.66	3.32
4.03	3.01	3.8	2.3	3.15	2.84	1.63	2.14
0.444	0.418	0.536	0.432	0.436	1.25	2.008	1.608
0.28	0.3	0.3	0.28	0.32	0.44	0.39	0.62
2.33	4.36	3.13	5.51	4.03	10.85	7.97	5.61
0.296	0.053	0.155	0.633	0.112	0.231	0.255	0.092
99.63	98.4	100.4	100.2	100.3	99.53	100.2	99.28
1322	1323	1002	545	1232	343	614	1572
11.2	8.6	11.7	8	9.2	27.1	22.6	10.1
22.6	46.8	36	35.1	19.3	15.8	21.3	16.4
1.7	2.1	1.6	2.9	1.9	2.4	0.7	0.7
17	15	15	16	18	16	16	16
1.6	1.8	1.4	1.7	1.7	2.4	1.7	2
2.3	1.9	2.1	2.4	2.4	3	3.2	3.3
10.3	9.9	9.2	10.4	11.1	8.3	8.4	9.4
5	5	6	6	5	4	3	5
< 5	< 5	< 5	< 5	< 5	< 5	< 5	< 5
62	61	57	43	54	66	28	32
14.2	14.5	15.8	12.8	15.5	22.1	32	26.2
978	1083	985	737	800	668	447	507
0.69	0.79	0.6	0.75	0.74	0.54	0.53	0.56
74	86	88	91	75	131	132	138
2.3	1.9	2.1	2.4	2.4	3	3.2	3.3
1.46	1.71	1.27	1.74	1.38	1.01	1.34	1.89
3.48	3.01	2.7	3.45	3.26	1.8	2.39	3.46
129	130	148	122	132	221	480	129
17	14	15	18	16	23	27	36
66	52	60	36	50	84	99	111
74	86	88	91	75	131	132	138
0.06	0.24	0.2	0.14	0.08	0.61	0.14	0.17
16	14.3	13	16.2	15.5	14.3	16.1	20
31.4	27.4	26.1	31.9	31.5	30.3	34.9	42.4
3.54	3.3	3.26	3.64	3.64	4.16	4.75	5.84
14	11.7	12	14.3	15.1	16.2	18.4	23.8
3.22	2.4	2.51	2.96	2.78	3.9	4.28	5.57
0.958	0.798	0.818	1.1	1.01	1.56	1.46	2.56
2.91	2.14	2.28	2.69	2.7	3.98	4.64	5.8
0.46	0.39	0.41	0.42	0.46	0.74	0.84	1.06
2.8	2.47	2.59	2.48	2.88	4.73	5.29	6.78
0.55	0.5	0.52	0.49	0.57	0.95	1.12	1.4
1.61	1.5	1.56	1.46	1.7	2.57	3.19	4
0.243	0.226	0.241	0.23	0.251	0.369	0.467	0.581
1.71	1.57	1.68	1.6	1.71	2.42	3.14	3.82

0.282	0.265	0.264	0.281	0.294	0.391	0.465	0.597
9.357	9.108	7.738	10.125	9.064	5.909	5.127	5.236
57.529	77.357	65.667	40.944	50.000	29.043	16.556	14.083
0.606	0.707	0.613	0.578	0.694	0.361	0.311	0.261
0.027	0.033	0.027	0.033	0.028	0.016	0.010	0.014
9.085	8.966	9.367	9.531	8.516	10.000	15.000	4.924
0.951	1.070	1.039	1.185	1.120	1.203	0.996	1.369

Table 3. Whole-rock Nd-Sr isotopic results for the porphyry and basaltic to andesitic dike rocks

Sample	Rb ($\mu\text{g/g}$)	Sr ($\mu\text{g/g}$)	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	$\pm 2\sigma$	T (Ma)	$(^{87}\text{Sr}/^{86}\text{Sr})_T$	Sm ($\mu\text{g/g}$)	Nd ($\mu\text{g/g}$)
P1 porphyry									
RC13-39	44.10	640.50	0.199213	0.705022	0.000020	211.6	0.7044	2.29	9.96
RC13-40	36.38	96.01	1.096378	0.707480	0.000032	211.6	0.7042	2.94	15.77
P2E porphyry									
RC13-92	18.75	703.50	0.077095	0.704633	0.000019	206.0	0.7044	3.06	15.31
RC13-107	43.23	320.70	0.390074	0.705837	0.000037	206.0	0.7047	2.97	14.70
P2L porphyry									
RC13-32	27.07	509.50	0.153691	0.705103	0.000035	203.6	0.7047	2.65	12.77
RC13-33	47.40	846.00	0.162098	0.704873	0.000017	203.6	0.7044	2.63	12.87
P3 porphyry									
RC13-77	37.47	531.70	0.203855	0.704991	0.000016	201.7	0.7044	2.80	14.06
RC13-78	25.78	427.70	0.174403	0.704748	0.000024	201.7	0.7042	2.67	13.03
Mafic to andesitic dike									
RC13-97	23.90	385.10	0.179519	0.704792	0.000020	200.0	0.7043	6.22	25.72
RC13-79	15.73	252.40	0.180303	0.704741	0.000030	200.0	0.7042	4.81	20.18

Notes: ¹N/A = no data; the used ages for calculations are from zircon U-Pb dating results for porphyry rocks, and assuming $T_{\text{DMI}} = (1/\lambda) \times \text{Ln}[\frac{(^{143}\text{Nd}/^{144}\text{Nd})_{\text{sample}} - (^{143}\text{Nd}/^{144}\text{Nd})_{\text{DM}}}{(^{147}\text{Sm}/^{144}\text{Nd})_{\text{sample}} - (^{147}\text{Sm}/^{144}\text{Nd})_{\text{DM}} + 1}]$. See Appendix and depleted mantle. See the Digital Appendix Table A1 for sample locations.

at Red Chris¹

¹⁴⁷ Sm/ ¹⁴⁴ Nd	(¹⁴³ Nd/ ¹⁴⁴ Nd) ₀	± 2σ	ε _{Nd} (t = 0)	(¹⁴³ Nd/ ¹⁴⁴ Nd) _T	T _{DMI} (Ga)	ε _{Nd} (T)
0.139132	0.512732	0.000009	1.8	0.512540	0.88	3.4
0.112591	0.512672	0.000010	0.7	0.512516	0.74	2.9
0.120965	0.512659	0.000007	0.4	0.512496	0.83	2.4
0.122321	0.512700	0.000010	1.2	0.512535	0.77	3.2
0.125462	0.512713	0.000008	1.5	0.512546	0.78	3.3
0.123603	0.512713	0.000010	1.5	0.512548	0.76	3.4
0.120209	0.512686	0.000012	0.9	0.512527	0.78	2.9
0.123767	0.512727	0.000010	1.7	0.512564	0.74	3.6
0.146221	0.512751	0.000006	2.2	0.512559	N/A	3.5
0.144047	0.512766	0.000011	2.5	0.512578	N/A	3.9

200 Ma for basaltic to andesitic dikes based on geological relationship, see text for details.

ix 1 for the ¹⁴³Nd/¹⁴⁴Nd and ¹⁴⁷Sm/¹⁴⁴Nd values of present day chondrite

Table 4. Estimates of magmatic temperature and sulfur–chlorine content from igneous apatite and whole-rock compositions

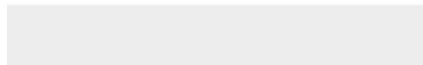
Porphyry Phase	Sample	Whole rock SiO ₂ (wt. %) ¹	Whole rock P ₂ O ₅ (wt. %) ¹	AST ² (°C)	Apatite SO ₃ (wt. %) (n) ³ [maximum, minimum]	Apatite Cl (wt. %) (n) ³	Apatite F (wt. %) (n) ³	Apatite Molar S/Cl (n) ³	Apatite Molar Cl/F (n) ³	Average magmatic S content (wt. %) ⁴ [maximum]	Average magmatic S content (wt. %) ⁵ [maximum]	Average magmatic Cl content (wt. %) ⁶
P1	RC13-39	59.93	0.31	921	0.11 ± 0.03 (34) [0.05, 0.19]	0.51 ± 0.3 (34)	3.64 ± 0.48 (34)	0.13 ± 0.08 (34)	0.08 ± 0.06 (34)	0.010 ± 0.003 [0.017]	0.002 ± 0.0004 [0.003]	0.63 ± 0.38
P2E	RC13-13 ⁷	63.33	0.25	932	0.28 ± 0.10 (19) [0.11, 0.46]	1.47 ± 0.22 (19)	2.32 ± 0.29 (19)	0.09 ± 0.03 (19)	0.35 ± 0.10 (19)	0.030 ± 0.010 [0.050]	0.006 ± 0.004 [0.015]	1.84 ± 0.27
	RC13-107	59.25	0.30	909	0.32 ± 0.17 (15) [0.18, 0.80]	0.82 ± 0.10 (15)	2.36 ± 0.30 (15)	0.17 ± 0.09 (15)	0.19 ± 0.05 (15)	0.031 ± 0.017 [0.078]	0.014 ± 0.031 [0.012]	1.02 ± 1.13
P2L	RC13-26	58.08	0.30	892	0.30 ± 0.22 (48) [0.07, 1.2]	1.57 ± 0.33 (48)	2.23 ± 0.22 (48)	0.11 ± 0.20 (48)	0.39 ± 0.10 (48)	0.017 ± 0.013 [0.070]	0.0490 ± 0.2380 [1.6168]	1.96 ± 0.41
	RC13-33	58.61	0.31	905	0.28 ± 0.13 (52) [0.14, 0.79]	1.37 ± 0.17 (52)	2.40 ± 0.21 (52)	0.09 ± 0.04 (52)	0.31 ± 0.05 (52)	0.020 ± 0.009 [0.056]	0.008 ± 0.017 [0.117]	1.71 ± 0.21
P3	RC13-78	59.86	0.33	928	0.03 ± 0.01(13) [0.02, 0.05]	0.02 ± 0.02 (17)	3.60 ± 0.40 (17)	0.16 ± 0.09 (13)	0.01 ± 0.01 (17)	0.003± 0.001 [0.005]	0.001 ± 0.0001 [0.001]	0.12 ± 0.07

Notes:

1. Normalized to 100 wt. % (Digital Appendix Table A3).
2. Apatite saturation temperature (AST) calculated from whole-rock SiO₂ and P₂O₅ concentrations using the equation of Piccoli and Candela (1994).
3. Average of all igneous apatite analyses (Digital Appendix Table A8).
4. Estimated from apatite SO₃ contents (Digital Appendix Table A8) using the temperature-dependent apatite–melt partition coefficient formula of Peng et al. (1997):
 $\ln K_D = 21130/T - 16.2$ (where T is in Kelvin).
5. Estimated from apatite SO₃ contents (Digital Appendix Table A8) using the temperature-dependent apatite–melt partition coefficient formula of Parat et al. (2011):
 $SO_3 \text{ apatite (wt. \%)} = 0.157 \times \ln SO_3 \text{ glass (melt, wt.\%)} + 0.9834$ ($r^2 = 0.62$).
6. Estimated from apatite Cl contents (Digital Appendix Table A8) using the apatite–melt partition coefficient value (mass ratios) of Mathez and Webster (2005), which is 0.8 for basaltic melt (51.1 wt. % SiO₂) and tends to be similar for rhyodacitic melt at 200 MPa (Webster et al., 2009).
7. SiO₂ and P₂O₅ compositions for this sample taken from sample RC13-92 with same lithology as RC13-13 but less altered (Digital Appendix Table A3).



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