

1 **Reconstruction of an early Permian, sub-lacustrine magmatic-hydrothermal system: Mt**
2 **Carlton epithermal Au-Ag-Cu deposit, northeastern Australia**

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Abstract

22

23 The Mt Carlton Au-Ag-Cu deposit, northern Bowen Basin, northeastern Australia, is an
24 uncommon example of a sub-lacustrine hydrothermal system containing economic high-
25 sulfidation epithermal mineralization. The deposit formed in the early Permian and comprises
26 vein- and hydrothermal breccia-hosted Au-Cu mineralization within a massive rhyodacite
27 porphyry (V2 open pit) and stratabound Ag – barite mineralization within volcano-lacustrine
28 sedimentary rocks (A39 open pit). These ore bodies are all associated with extensive advanced
29 argillic alteration of the volcanic host rocks. Stable isotope data for disseminated alunite ($\delta^{34}\text{S}$
30 = +6.3 to +29.2 ‰; $\delta^{18}\text{O}_{\text{SO}_4}$ = -0.1 to +9.8 ‰; $\delta^{18}\text{O}_{\text{OH}}$ = -15.3 to -3.4 ‰; δD = -102 to -79 ‰)
31 and pyrite ($\delta^{34}\text{S}$ = -8.8 to -2.7 ‰), and void-filling anhydrite ($\delta^{34}\text{S}$ = +17.2 to +19.2; $\delta^{18}\text{O}_{\text{SO}_4}$ =
32 +1.8 to +5.7 ‰), suggest that early advanced argillic alteration formed within a magmatic-
33 hydrothermal system. The ascending magmatic vapor ($\delta^{34}\text{S}_{\Sigma\text{S}} \approx -1.3$ ‰) was absorbed by
34 meteoric water (~50 to 60 % meteoric component), producing an acidic (pH \approx 1) condensate
35 that formed a silicic \rightarrow quartz – alunite \rightarrow quartz – dickite – kaolinite zoned alteration halo
36 with increasing distance from feeder structures. The oxygen and hydrogen isotopic
37 compositions of alunite-forming fluids at Mt Carlton are lighter than those documented at
38 similar deposits elsewhere, probably due to the high paleo-latitude (\sim S60°) of northeastern
39 Australia in the early Permian. Veins of coarse-grained, banded plumose alunite ($\delta^{34}\text{S}$ = +0.4
40 to +7.0 ‰; $\delta^{18}\text{O}_{\text{SO}_4}$ = +2.3 to +6.0 ‰; $\delta^{18}\text{O}_{\text{OH}}$ = -10.3 to -2.9 ‰; δD = -106 to -93 ‰) formed
41 within feeder structures during the final stages of advanced argillic alteration. Epithermal
42 mineralization was deposited subsequently, initially as fracture- and fissure-filling, Au-Cu-rich
43 assemblages within feeder structures at depth. As the mineralizing fluids discharged into lakes,
44 they produced syngenetic Ag – barite ore. Isotope data for ore-related sulfides and sulfosalts
45 ($\delta^{34}\text{S}$ = -15.0 to -3.0 ‰) and barite ($\delta^{34}\text{S}$ = +22.3 to +23.8 ‰; $\delta^{18}\text{O}_{\text{SO}_4}$ = -0.2 to +1.3 ‰), and
46 microthermometric data for primary fluid inclusions in barite (Th = 116 to 233°C; 0.0 to 1.7

47 wt.% NaCl), are consistent with metal deposition at temperatures of $\sim 200 \pm 40^\circ\text{C}$ (for Au-Cu
48 mineralization in V2 pit) and $\sim 150 \pm 30^\circ\text{C}$ (Ag mineralization in A39 pit) from a low-salinity,
49 sulfur- and metal-rich magmatic-hydrothermal liquid that mixed with vapor-heated meteoric
50 water. The mineralizing fluids initially had a high-sulfidation state, producing enargite-
51 dominated ore with associated silicification of the early-altered wall rock. With time, the fluids
52 evolved to an intermediate-sulfidation state, depositing sphalerite- and tennantite-dominated
53 ore mineral assemblages. Void-filling massive dickite ($\delta^{18}\text{O} = -1.1$ to $+2.1$ ‰; $\delta\text{D} = -121$ to -
54 103 ‰) with pyrite was deposited from an increasingly diluted magmatic-hydrothermal fluid
55 (≥ 70 % meteoric component) exsolved from a progressively degassed magma. Gypsum ($\delta^{34}\text{S}$
56 $= +11.4$ to $+19.2$ ‰; $\delta^{18}\text{O}_{\text{SO}_4} = +0.5$ to $+3.4$ ‰) occurs in veins within post-mineralization
57 faults and fracture networks, likely derived from early anhydrite that was dissolved by
58 circulating meteoric water during extensional deformation. This process may explain the
59 apparent scarcity of hypogene anhydrite in lithocaps elsewhere. While the Mt Carlton system
60 is similar to those that form subaerial high-sulfidation epithermal deposits, it also shares several
61 key characteristics with magmatic-hydrothermal systems that form base- and precious-metal
62 mineralization in shallow-submarine volcanic arc and backarc settings. The lacustrine
63 paleosurface features documented at Mt Carlton may be useful as exploration indicators for
64 concealed epithermal mineralization in similar extensional terranes elsewhere.

65

66

Introduction

67 The Mt Carlton Au-Ag-Cu deposit is located within an extensively mineralized, lower Permian
68 volcano-sedimentary succession in the northern Bowen Basin, northeastern Queensland,
69 Australia (Fig. 1). The deposit was discovered in 2006 and contains a total resource of 11.9 Mt
70 averaging 2.8 g/t Au, 28 g/t Ag and 0.44 % Cu (as of December 2017; M. Obiri-Yeboah, pers.

71 commun., 2018; Evolution Mining, 2019). The mining operation at Mt Carlton includes the
72 larger V2 open pit in the NE – producing 112 koz Au with a processed grade of 5.6 g/t Au in
73 2018 (Evolution Mining, 2019) – and the smaller A39 open pit in the SW (Fig. 2).

74 Previous studies (Dugdale and Howard, 2017; Sahlström et al., 2017, 2018) indicated
75 that Mt Carlton has many characteristics typical of high-sulfidation (HS) epithermal deposits
76 (Arribas, 1995a; Hedenquist et al., 2000). Key among these are its tectonic setting within a
77 volcanic arc (in this case, a backarc rift); intermediate-composition, calc-alkaline volcanic host
78 rocks (rhyodacite porphyry and tuff) with an age similar to that of mineralization; laterally
79 extensive, zoned advanced argillic alteration of magmatic-hydrothermal origin (Rye et al.,
80 1992); and a HS ore mineral assemblage dominated by enargite (Einaudi et al., 2003).
81 However, Mt Carlton differs from other HS epithermal deposits in that a lacustrine, shallow-
82 water vent facies contains economic mineralization. Other notable features of Mt Carlton are
83 well-defined mineral and metal zonation patterns in both time and space and its early Permian
84 age (~280 Ma; Sahlström et al., 2018), which places Mt Carlton among the world's oldest non-
85 metamorphosed HS epithermal deposits.

86 This study examines the genetic processes at Mt Carlton, building on the geological
87 framework and mineral paragenesis studies presented by Sahlström et al. (2018). Emphasis is
88 placed on the stable isotope and fluid inclusion systematics of the alteration and ore mineral
89 assemblages, and the implications for processes within the magmatic-hydrothermal, magmatic-
90 steam, and supergene environments (Rye et al., 1992; Rye, 2005). Our main objective is to
91 reconstruct the Mt Carlton magmatic-hydrothermal system and to make comparisons to the
92 generally much younger magmatic-hydrothermal systems that form comparable subaerial or
93 shallow-water mineral deposits elsewhere. Some implications for mineral exploration are also
94 discussed.

Regional Setting

95

96 The New England Orogen extends from eastern New South Wales to northeastern Queensland
97 along the eastern coast of Australia (Fig. 1A), and may extend further offshore (Mortimer et
98 al., 2008). It was a dynamic continental margin active mainly during the Carboniferous and
99 Permian (Donchak et al., 2013). The Bowen Basin is an elongate, early Permian-Middle
100 Triassic sedimentary basin that formed inland of the New England Orogen (Fig. 1). This basin
101 is the northern continuation of a larger basin system that also includes the Sydney and
102 Gunnedah basins (Korsch and Totterdell, 2009; Korsch et al., 2009; Donchak et al., 2013). The
103 Bowen Basin opened in the early Permian, as a result of backarc extension within the lower
104 Paleozoic metamorphic and sedimentary basement. Backarc rifting produced a series of
105 isolated grabens and half-grabens that were filled with volcanic and sedimentary rocks (Esterle
106 et al., 2002). The Lizzie Creek Volcanic Group and its associated hydrothermal mineral
107 systems formed in the northern Bowen Basin during this stage. Hydrothermal activity mostly
108 produced sub-economic porphyry and epithermal prospects (Fig. 1B), but the recent discovery
109 of Mt Carlton has confirmed the potential for economic mineralization in the region.

110 Backarc rifting was followed in the middle Permian by thermal relaxation, widespread
111 marine transgression, and deposition of shallow-water sediments (Malone et al., 1969; Allen
112 and Fielding, 2007; Korsch and Totterdell, 2009; Korsch et al., 2009). In the late Permian, a
113 series of westward-stepping orogenic fronts formed as a consequence of the ~265 – 235 Ma
114 Hunter-Bowen Orogeny, causing tectonic inversion and development of a foreland basin
115 sequence on top of the backarc basin (Donchak et al., 2013). Infilling by terrestrial sediments
116 and thick coal measures in the foreland basin proceeded until the Bowen Basin closed in the
117 Middle Triassic (Fielding et al., 1990; Fergusson, 1991; Holcombe et al., 1997; Esterle et al.,
118 2002).

119

Mt Carlton HS Epithermal Deposit

120 *Tectono-stratigraphic evolution*

121 The crystalline basement in the Mt Carlton area is formed of fine- to medium-grained
122 monzogranites of the Urannah Batholith (Unit 1, ~302-296 Ma; Sahlström et al., 2018; I.
123 Corral, unpub. data, 2018). The basement is overlain by a sequence of volcanic and
124 sedimentary rocks belonging to the lower Permian Lizzie Creek Volcanic Group (~288-275
125 Ma; Sahlström et al., 2018; I. Corral, unpub. data, 2018). From bottom to top, the Lizzie Creek
126 sequence exposed near the open pits at Mt Carlton consists of (Fig. 3): fine-grained andesite
127 lava and minor volcanoclastic rocks (Unit 2); massive, locally flow-banded, porphyritic
128 rhyodacite and minor rhyolite lava (Unit 3); well-bedded, fragmental, rhyodacitic lapilli tuff
129 with interbedded carbonaceous lacustrine sediments (Unit 4A); massive, bedded dacitic tuffs
130 (Unit 4B); volcanoclastic dacitic rocks including ignimbrites (Units 5A-C); fragmental andesite
131 with local rounded boulders (Unit 6); coal-bearing volcano-sedimentary rocks of andesitic to
132 dacitic composition (Unit 7); and strongly flow-banded porphyritic rhyolite (Unit 8). Whereas
133 Units 2 and 8 are regionally continuous, Units 3-7 are laterally discontinuous due to deposition
134 during active rifting and exhibit facies variability typical of volcanic-dome complex
135 environments, the most common setting for HS epithermal deposits (Sillitoe, 1999; Hedenquist
136 et al., 2000). Mineralization is hosted partly in massive and flow-banded rhyodacite porphyry
137 (Unit 3), and partly in overlying volcano-lacustrine sedimentary rocks (Unit 4A; Fig. 3;
138 Sahlström et al., 2018).

139 Hydrothermal alteration and mineralization at Mt Carlton occurred at ~280 Ma, during
140 high-angle normal faulting in response to both E-W and N-S extension, partly
141 contemporaneously with deposition of volcanic sediments into localized half-graben and
142 graben basins (D₁; Sahlström et al., 2018). Continued E-W extension in the northern Bowen

143 Basin caused Mt Carlton to be buried beneath volcano-sedimentary cover, which contributed
144 to the preservation of the deposit. Extension also produced a series of low-angle, locally layer-
145 parallel normal faults that cut across the mineralized rock pile (D₂). In the northern wall of the
146 V2 pit, D₂ faults are crosscut by a younger, rhyodacitic volcanic vent that was emplaced at
147 ~266 Ma (Sahlström et al., 2018; I. Corral, unpub. data, 2018). High-angle normal faulting in
148 response to N-S extension with an overall S-down movement (D₃) was followed by block
149 rotation of kilometer-scale lithological domains across steep NNW-trending normal faults and
150 ENE-trending cross faults (D₄). A NNW-trending D₄ normal fault separates the V2 and A39
151 pits, and has caused segmentation of the stratigraphy, the ore zones and the alteration halo of
152 Mt Carlton. Primary sedimentary layering and mineralization have been reoriented locally. Due
153 to D₄ block rotation, mineralization in the A39 pit now plunges ~32° to the WSW, whereas
154 mineralization in the V2 pit retains its originally steep orientation (Figs. 2, 3). Emplacement of
155 plagioclase – pyroxene ± hornblende phyric basaltic dikes into pre-existing high-angle faults
156 (D₅) was followed by dextral-dominant strike-slip faulting along the dike margins (D₆). The
157 final deformation stage at Mt Carlton involved emplacement of WNW-trending, feldspar
158 phyric basaltic dikes containing quartz amygdules (D₇), which may be linked to basaltic flows
159 that occur west of the pit area (Sahlstrom et al., 2018).

160 As a result of post-mineralization deformation, the deeper parts of the mineralized
161 feeder system and the causative intrusion (including any potentially associated porphyry
162 mineralization) are expected to be displaced from the current location of the Mt Carlton HS
163 epithermal deposit (Sahlström et al., 2018). The rhyodacite package that hosts Mt Carlton is
164 divided into two tectonic blocks that are internally intact (V2 and A39 blocks; Fig. 3A), such
165 that the magmatic-hydrothermal system is undisturbed over several hundred meters.

166 *Paragenetic sequence*

167 A paragenetic sequence of alteration and ore minerals at Mt Carlton was established by
168 Sahlström et al. (2018) based on overprinting and crosscutting relationships (Fig. 4). Stage 1A
169 is represented within the host volcanic units by alteration assemblages that are zoned outward,
170 away from the paleo-fluid conduits as follows: silicic (locally with vuggy texture) → quartz –
171 alunite → quartz – dickite – kaolinite (Figs. 3, 4). This alteration halo envelops mineralization
172 in Units 3 and 4A and has a lateral extent of several hundred square meters. The silicic zones
173 have a gradational contact with the surrounding quartz – alunite zones. Stage 1A alunite is
174 typified by up to ~300 µm long, pink to red, euhedral and platy crystals that are either
175 disseminated in microcrystalline quartz, or aggregated, replacing feldspar phenocrysts or filling
176 vugs (Figs. 5A-B). Alunite has compositions within the alunite – natroalunite solid solution
177 series [trigonal $\text{KAl}_3(\text{SO}_4)_2(\text{OH})_6$ – $\text{NaAl}_3(\text{SO}_4)_2(\text{OH})_6$]. The crystals are commonly zoned
178 with respect to Na and K contents, locally with cores of aluminium-phosphate-sulphate (APS)
179 minerals of the woodhouseite – svanbergite solid solution series [trigonal
180 $\text{CaAl}_3(\text{SO}_4)(\text{PO}_4)(\text{OH})_6$ – $\text{SrAl}_3(\text{SO}_4)(\text{PO}_4)(\text{OH})_6$], typical of magmatic-hydrothermal alunite
181 (c.f., Rye et al., 1992; Rye, 2005). Finely disseminated pyrite generally co-exists with alunite
182 (Fig. 5A). Anhydrite, which is rarely reported in HS epithermal deposits and lithocaps, is
183 present locally within the silicic cores (Fig. 4). It occurs as euhedral, tabular crystals, locally in
184 massive aggregates, that fill open spaces (Fig. 5E). However, most of the Stage 1A anhydrite
185 was dissolved, leaving behind distinctive casts (Fig. 5F).

186 Alunite also occurs in monomineralic veins within the silicic alteration zones (Stage
187 1B; Figs. 4, 5A, 5C-D, 6C; Sahlström et al., 2018). The Stage 1B veins generally post-date
188 silicic alteration and exhibit a banded arrangement of coarse, plumose alunite crystals. They
189 are texturally similar to the type occurrence of magmatic-steam alunite at Marysvale, Utah
190 (Cunningham et al., 1984; Rye et al., 1992; Rye, 2005).

191 Epithermal mineralization at Mt Carlton evolved temporally with respect to metal
192 contents and mineral assemblages. Stage 2 of the paragenetic sequence comprises three sub-
193 stages (Stages 2A-C; Fig. 4; Sahlström et al., 2018), all hosted mainly within the silicic cores.
194 The initial stage of Cu-Au-Ag mineralization is dominated by enargite, a HS state sulfosalt,
195 and is associated with barite gangue as well as silicification of the early-altered wall rocks
196 (Stage 2A; Figs. 4, 5G, 6A-D). The Stage 2A assemblage is crosscut by a younger, well-
197 developed stage of Zn-Pb-Au-Ag mineralization dominated by Fe-poor (<350 ppm Fe)
198 sphalerite, with tetrahedrite-group minerals and chalcopyrite present, indicating an
199 intermediate-sulfidation (IS) state (Stage 2B; Figs. 4, 6C-D; Sahlström et al., 2017). Stage 2B
200 mineralization is overprinted by another IS stage that is rich in Cu, Au and Ag (Stage 2C; Fig.
201 4). The Stage 2C mineral assemblage is dominated by tennantite, visible at a microscopic scale.
202 Late-stage, voluminous hydrothermal void fill of massive, microcrystalline dickite with pyrite
203 overprinted Stage 2 mineralization throughout Mt Carlton (Stage 3; Figs. 4, 5G, 6D-E).

204 Pervasive illite – montmorillonite ± red hematite alteration developed during the D₂
205 deformation event, affecting Units 2, 4, 5 and 6 in the area near the open pits (Stage 4; Figs. 3,
206 4). Veins of fibrous gypsum with a well-developed shear fabric occur near the major D₂ and
207 D₃ faults (Fig. 5H). Finally, a suite of secondary Cu-bearing ore minerals, including covellite,
208 chalcocite and malachite, have locally overprinted the Stage 2 ore assemblages, caused by
209 recent supergene oxidation in the upper ~50 meters of the present-day deposit (Stage 5; Fig. 4;
210 Sahlström et al., 2018).

211 *Alteration and mineralization zonation*

212 The alteration and mineralization at Mt Carlton produced distinct spatial zonation patterns
213 across the deposit. In the V2 pit, Au-Cu mineralization occurred in paleo-fluid conduits in the
214 rhyodacite porphyry (Unit 3), producing steeply dipping veins and hydrothermal breccia (Fig.

215 6A). The mineralized fracture systems have predominantly NE to NNE trends, and three
216 distinct ore zones (Eastern, Western and Link; Fig. 2) are aligned in an en echelon fashion
217 along a broadly E-trending corridor. The Western ore zone extends ~600 m to the SW, into the
218 A39 pit, and it exhibits a well-defined metal zonation along strike from NE to SW of: Au-Cu
219 → Cu-Zn-Pb-Ag → Ag-Pb-(Cu) → Ag (Figs. 2, 3). Along the same spatial trend, the Na-
220 content of Stage 1A disseminated alunite in the alteration halo decreases (Sahlström et al.,
221 2018). Furthermore, the abundance of both APS minerals and pyrite associated with
222 disseminated alunite decreases to the SW, and both minerals are largely absent in the SW parts
223 of the deposit. Stage 1B plumose alunite veins are predominantly concentrated in the high-
224 grade feeder structures in the V2 pit. However, they have also been observed locally in small
225 veins occurring distal to the main feeders, including in the A39 pit.

226 Silver mineralization is concentrated in the A39 pit within the volcano-lacustrine
227 sedimentary rocks (Unit 4A) that overlie the rhyodacite porphyry (Fig. 3). In A39 pit,
228 mineralization is stratabound and oriented parallel to primary sedimentary bedding.
229 Hydrothermal features observed in rocks from A39 pit include: 1) voluminous deposits of
230 finely laminated quartz sediment (interpreted as recrystallized from amorphous silica; Sillitoe,
231 2015); 2) massive layers of spherulitic pyrite and dickite; 3) skeletal aggregates of enargite and
232 luzonite; and 4) a variety of synsedimentary deformation textures, such as growth faults and
233 soft-sediment slumping and fluidization structures (Figs. 6B, 6D-E; Sahlström et al., 2018,
234 their Figure 13). Barite is present in minor amounts in the V2 pit, occurring as up to ~0.5 cm
235 diameter crystals intergrown with massive enargite or sphalerite. Barite is primarily
236 concentrated in the A39 pit, where it occurs as disseminated to massive accumulations closely
237 associated with, and capping, the Ag mineralization. In A39 pit, barite locally also occurs as
238 coarse-grained halos to enargite veinlets (Fig. 7B). Barite at Mt Carlton is highly enriched in
239 Pb (average 1.3 wt.%), and to a lesser degree, in Sr (solid solution toward hokutolite:

240 [Ba,Pb,Sr]SO₄; Sahlström et al., 2017). Because galena and other Pb-rich minerals are
241 volumetrically minor in the A39 pit, the distal geochemical anomaly of Pb observed there (Figs.
242 2, 3) essentially outlines the spatial distribution of barite.

243 Overall, these observations consistently indicate that, at the time of ore formation, the
244 position of the stratigraphic sequence in the V2 pit was deep and intrusion-proximal relative to
245 the sequence in the A39 pit. Mineralization in A39 pit has been interpreted to have formed
246 essentially at the paleosurface in a sub-lacustrine environment (Sahlström et al., 2018).

247 **Fluid Inclusions**

248 *Petrography*

249 Suitable host minerals for fluid inclusion study are typically uncommon in HS epithermal
250 deposits (Arribas, 1995a), which also limited this study. Aqueous fluid inclusions occur in
251 igneous quartz phenocrysts within the silicic and quartz – alunite alteration zones at Mt Carlton.
252 Although lacking temporal information, such inclusions can provide insights into the fluid
253 regime in HS epithermal deposits (e.g., Bruha and Noble, 1993; Arribas et al., 1995, Ruggieri
254 et al., 1997; Corral et al., 2017). There is evidence for hydrothermal fluids being trapped along
255 fractures in quartz phenocrysts during alteration (e.g., replacement of primary melt inclusions
256 by hydrothermal illite in quartz phenocrysts at the Rodalquilar HS epithermal deposit, Spain;
257 Arribas et al., 1995, their Figure 8G). In the studied sections from Mt Carlton, quartz
258 phenocrysts exhibit fractures and corrosion textures, which are infilled with Stage 1A
259 hydrothermal quartz and alunite (Fig. 7A). Fluid inclusions locally occur in trails that propagate
260 from such fractures or from the quartz crystal boundaries (Fig. 7C), and are interpreted to be
261 secondary in nature (Roedder, 1984). The inclusion trails cannot be observed to crosscut quartz
262 crystal boundaries, due to the microcrystalline texture of the surrounding minerals (Fig. 7A).
263 The inclusions range from <1 to ~35 μm in diameter, and have irregular to sub-rounded shapes.

264 They are two-phase and liquid-rich, with a dark vapor bubble typically occupying up to ~30
265 vol.% of the inclusion (Fig. 7C).

266 Aqueous fluid inclusions are also present in barite associated with Stage 2A
267 mineralization, as isolated, two-phase liquid-vapor inclusions with up to ~30 vol.% vapor (Fig.
268 7D). The inclusions are irregular to sub-rounded, up to ~15 μm in diameter, and are interpreted
269 as primary based on the criteria of Roedder (1984). The primary inclusion fields are crosscut
270 by trails of small ($<2 \mu\text{m}$) monophasic liquid inclusions that also crosscut cleavage planes and
271 barite crystal boundaries. These secondary fluid inclusions (Roedder, 1984) were not studied
272 as they represent late-stage fluids not related to mineralization.

273 *Methods*

274 Microthermometric measurements of fluid inclusions were conducted at the Department of
275 Geosciences, James Cook University, Australia, on doubly polished sections (~200 μm thick)
276 using a Linkam MDS600 heating-freezing stage. The instrument was calibrated by measuring
277 the melting temperature of CO_2 (-56.6°C) in synthetic $\text{H}_2\text{O}-\text{CO}_2$ fluid inclusions, and the
278 melting point of pure H_2O (0°C) in synthetic H_2O fluid inclusions. The two-phase fluid
279 inclusions in quartz and barite were observed for final melting of ice (T_m ; ice + liq_{aq} + vap \rightarrow
280 liq_{aq} + vap) and homogenization into the liquid phase (T_h ; liq_{aq} + vap \rightarrow liq_{aq}). The instrument
281 reproducibility is generally $\pm 0.5^\circ\text{C}$ during T_m measurements and $\pm 5^\circ\text{C}$ during T_h
282 measurements. Salinities and densities were calculated from final ice melting and
283 homogenization temperatures using the HOKIEFLINCS_H2O-NACL spreadsheet (Steele-
284 MacInnis et al., 2012).

285 *Results*

286 *Secondary fluid inclusions in quartz phenocrysts* – Microthermometric measurements of
287 secondary fluid inclusions in quartz phenocrysts were done on two samples collected from the

288 proximal (V2 pit) and distal (A39 pit) parts of Mt Carlton, respectively (Fig. 8; Table 1).
289 Measurements of 57 inclusions in the proximal sample (MCR024D-177) yielded a Th range of
290 163 to 264°C (Figs. 8, 9; Table 1). The fluid inclusions appeared to melt instantaneously at
291 their final melting point and no initial melting could be observed, indicating that NaCl is the
292 dominant solute. The T_m values range between -0.9 and 0.0°C, corresponding to salinities
293 between 0.0 and 1.6 wt.% NaCl. The fluid inclusion densities range between 0.77 and 0.91
294 g/cm³. Measurements of 72 inclusions in the distal sample (HC14DD1133-178) yielded partly
295 overlapping, but generally lower Th values, between 137 and 231°C, and T_m values between
296 -0.7 and 0.0°C (0.0 and 1.2 wt.% NaCl; Figs. 8, 9; Table 1). The fluid inclusion densities range
297 between 0.82 and 0.93 g/cm³.

298 *Primary fluid inclusions in barite* – Primary fluid inclusions in Stage 2A barite were measured
299 from two samples – one from the V2 pit (proximal) and another from the A39 pit (distal; Fig.
300 8; Table 1). The sample from V2 pit (V2-ORE-2) is a high-grade vein that contains barite
301 crystals intergrown with massive enargite ore, whereas the sample from A39 pit
302 (HC10RCD935-168) is an enargite veinlet that has coarse-grained barite in its halo (Fig. 7B).
303 Barite-hosted fluid inclusions are particularly susceptible to stretching during freezing and
304 heating (Ulrich and Bodnar, 1988). However, after measurements in this study, no petrographic
305 evidence for stretching (i.e., visible change in the size of the vapor bubble, or fracturing after
306 the heating-freezing experiments) was observed, and repeat measurements were within error.

307 Measurements of 38 inclusions in the proximal sample (V2-ORE-2) yielded a Th range
308 of 157 to 233°C (Figs. 8, 9; Table 1). Similar to the quartz phenocryst-hosted fluid inclusions,
309 these inclusions melted instantaneously at their final melting point and no initial melting could
310 be observed. Final melting of ice ranges between -1.0 to -0.2°C, corresponding to salinities
311 between 0.4 and 1.7 wt.% NaCl. The densities of the fluid inclusions range between 0.82 and
312 0.92 g/cm³. Measurements of 32 inclusions in the distal sample (HC10RCD935-168) yielded

313 Th values varying between 116 and 183°C and a Tm range of -0.2 to 0.0°C (Figs. 8, 9; Table
314 1). These measurements indicate salinities and densities ranging between 0.0 and 0.4 wt.%
315 NaCl and between 0.89 and 0.95 g/cm³, respectively.

316 *Interpretations*

317 Previous studies of primary fluid inclusions hosted in enargite from other HS epithermal
318 deposits have reported a wide range of salinities, from 0.2 to 20.4 wt.% NaCl (Deen et al.,
319 1994; Mancano and Campbell, 1995; Arribas, 1995a; Kouzmanov, 2001; Bailly et al., 2002;
320 Sillitoe and Hedenquist, 2003; Moritz et al., 2004). However, a study by Moritz (2006) found
321 that salinities obtained for fluid inclusions in enargite were commonly overestimated due to
322 heating of the sample by the infrared light source used during measurements (e.g., salinities of
323 ~20 wt.% NaCl were actually <2 wt.% NaCl). At the El Indio HS epithermal deposit, Chile,
324 quartz associated with enargite contains fluid inclusions with salinities of 2.9 to 4.6 wt.% NaCl,
325 whereas the salinity of the later, Au-rich stage is <1.7 wt.% NaCl (Jannas et al., 1990). The
326 primary fluid inclusions in barite measured here show relatively low salinities (<1.7 wt.%
327 NaCl), similar to those identified from the study of enargite- and quartz-hosted fluid inclusions
328 from other HS epithermal deposits. It remains unclear whether the secondary fluid inclusions
329 in quartz phenocrysts at Mt Carlton formed during early advanced argillic alteration (as
330 suggested by their intimate association with Stage 1A quartz and alunite; Fig. 7A) or from later
331 mineralizing fluids. However, the overlap in density and salinity between the quartz-hosted
332 and barite-hosted fluid inclusions, and identical Tm vs. Th trends (Fig. 9C-D), indicate that the
333 two fluid inclusion assemblages formed from similar fluids at similar temperatures. Geologic
334 evidence indicates that Mt Carlton formed at shallow depth (Sahlström et al., 2018), such that
335 the temperature of homogenization was close to the trapping temperature of the fluid inclusions
336 (Wilkinson, 2001).

361 Ag mineralization (pyrite, sphalerite and galena); Stage 2C, Cu-Au-Ag mineralization
362 (tennantite); Stage 3, late-hydrothermal void fill (dickite); and Stage 4, post-hydrothermal
363 gypsum veins. Backscatter electron (BSE) imaging revealed no more than trace to minor
364 amounts (<10 vol.%) of contaminating quartz, clays or sulfides in the separates.

365 *Sulfur isotope analyses of sulfates, sulfides and sulfosalts* – Sulfur isotope analyses were
366 conducted in laboratories of the Geology, Geophysics and Geochemistry Science Center, U.S.
367 Geological Survey, Denver, Colorado, USA. Mineral separates were analyzed for $\delta^{34}\text{S}$ using a
368 CE Elantech Inc. Flash 2000 elemental analyzer coupled to a ThermoFinnigan Delta Plus XP
369 mass spectrometer (Johnson et al., 2018). The system was calibrated with NBS-127 and IAEA-
370 SO-6 standards (for sulfate analysis) or NBS-123 and IAEA-S-3 standards [for sulfide and
371 sulfosalt analysis, the accepted compositions for which were taken from Brand et al. (2014)].
372 The $\delta^{34}\text{S}$ values are reported in per mil relative to Vienna Cañon Diablo Troilite (VCDT) with
373 a precision of 0.2 ‰ based on replicate analyses.

374 *Oxygen and hydrogen isotope analyses of sulfates* – In order to isolate the SO_4 molecule in
375 alunite, anhydrite and gypsum, mineral separates were dissolved in a sodium hydroxide
376 solution and reprecipitated as BaSO_4 , using methods described by Carmody et al. (1998).
377 Mineral samples and BaSO_4 precipitates were weighed into silver capsules and analysed for
378 $\delta^{18}\text{O}$ (alunite, barite, anhydrite and gypsum) and δD (alunite) at the U.S. Geological Survey in
379 Denver, using a ThermoFinnigan high-temperature conversion elemental analyzer coupled to
380 a ThermoFinnigan Delta Plus XL mass spectrometer. The complete set of isotope data for
381 alunite was obtained following the methodology of Wasserman et al. (1992). The mass
382 spectrometer system was calibrated by using the IAEA-SO-5 and IAEA-SO-6 standards (for
383 $\delta^{18}\text{O}$ measurements) or a series of reference materials that included IAEA-CH-7 and NBS-22
384 standards (Brand et al., 2014) as well as kaolinite and benzoic acid internal standards (for δD
385 measurements). Isotopic compositions are reported in per mil relative to Vienna Standard Mean

386 Ocean Water (VSMOW) with a reproducibility of ± 1 ‰ for $\delta^{18}\text{O}$ and ± 4 ‰ or better for δD ,
387 based on replicate analyses.

388 *Oxygen and hydrogen isotope analyses of dickite* – Samples of dickite were analyzed for
389 oxygen and hydrogen isotopes at the GNS Stable Isotope Laboratories in Wellington, New
390 Zealand. Oxygen was extracted from sample powders using a CO_2 laser and BrF_5 , and
391 subsequently analyzed for $\delta^{18}\text{O}$ in a Geo20-20 mass spectrometer (Sharp, 1990). Samples were
392 analyzed alongside, and normalized to, international standards NBS-28 and UWG-2 (Brand et
393 al., 2014). The δD analyses were done using a HEKAtech high-temperature elemental analyzer
394 coupled to a GV Instruments IsoPrime mass spectrometer, following reduction of the samples
395 in silver capsules at 1450°C . The system was calibrated with IAEA-CH-7, NBS-30 and NBS-
396 22 standards (Brand et al., 2014). The data are reported in per mil relative to VSMOW with a
397 reproducibility of ± 0.2 ‰ for $\delta^{18}\text{O}$ and ± 2 ‰ or better for δD .

398 *Results*

399 *Alunite, pyrite and anhydrite (Stage 1A)* – Disseminated alunite of Stage 1A alteration (Figs.
400 5A-B, 7A) has variable $\delta^{34}\text{S}$ (+6.3 to +29.2 ‰), $\delta^{18}\text{O}_{\text{SO}_4}$ (-0.1 to +9.8 ‰), $\delta^{18}\text{O}_{\text{OH}}$ (-15.3 to -
401 3.4 ‰) and δD (-102 to -79 ‰) values (Figs. 10, 11; Table 2). Pyrite that co-precipitated with
402 this alunite has $\delta^{34}\text{S}$ values between -8.8 and -2.7 ‰, whereas void-filling Stage 1A anhydrite
403 (Fig. 5E) has narrow ranges of $\delta^{34}\text{S}$ (+17.2 to +19.2 ‰) and $\delta^{18}\text{O}_{\text{SO}_4}$ (+1.8 to +5.7 ‰) values
404 (Fig. 10; Table 2). The sulfur isotope compositions of Stage 1A alunite, pyrite and anhydrite
405 from Mt Carlton are similar to those reported for magmatic-hydrothermal sulfates and sulfides
406 from HS epithermal deposits elsewhere (e.g., Hedenquist and Garcia, 1990; Rye et al., 1992;
407 Vennemann et al., 1993; Arribas, 1995a; Hedenquist et al., 1998; Rye, 2005; Bethke et al.,
408 2005).

409 *Plumose alunite veins (Stage 1B)* – Alunite forms from acidic hydrothermal fluids generated
410 through the condensation of magmatic vapor into groundwater (magmatic-hydrothermal
411 alunite; Rye et al., 1992; Rye, 2005). In addition, alunite is argued to also form from expanding
412 SO₂-rich magmatic vapor that rises rapidly from a degassing magma. Such magmatic-steam
413 alunite (Rye et al., 1992; Rye, 2005) typically occurs as monomineralic, banded and plumose
414 alunite in open tension fractures and breccias. It is predicted to have $\delta^{34}\text{S}$ values near that of
415 the total sulfur value (due to sulfur isotopic disequilibrium among the parental aqueous sulfur
416 species); the δD and $\delta^{18}\text{O}_{\text{SO}_4}$ values are near magmatic values, with $\delta^{18}\text{O}_{\text{OH}}$ values in
417 equilibrium with the fluid unless OH underwent retrograde exchange (Rye et al., 1992; Rye,
418 2005; Fifarek and Rye, 2005).

419 At Mt Carlton, plumose alunite in veins (Stage 1B; Figs. 4, 5C-D, 6C) has $\delta^{34}\text{S}$ values
420 of +0.4 to +7.0 ‰, noticeably lower than the values of Stage 1A disseminated alunite yet
421 distinctly higher than the value for total sulfur (Fig. 10; Table 2). The highest $\delta^{34}\text{S}$ values
422 correspond to samples collected outside the main feeder zones, including one sample from the
423 A39 pit (Table 2). The $\delta^{18}\text{O}_{\text{SO}_4}$ (+2.3 to +6.0 ‰) and $\delta^{18}\text{O}_{\text{OH}}$ (-10.3 to -2.9 ‰) values overlap
424 those of Stage 1A alunite, whereas the δD values (-106 to -93 ‰) overlap or are slightly lower
425 than Stage 1A values (Fig. 11; Table 2).

426 *Ore minerals and barite (Stage 2)* – Sulfides and sulfosalts formed during Stage 2 have
427 negative $\delta^{34}\text{S}$ values: -8.0 to -5.0 ‰ for Stage 2A pyrite, -9.9 to -3.0 ‰ for Stage 2A enargite,
428 -6.6 to -4.4 ‰ for Stage 2B pyrite, -15.0 to -6.8 ‰ for Stage 2B galena, -6.8 to -4.5 ‰ for
429 Stage 2B sphalerite, and -9.1 ‰ for Stage 2C tennantite (Fig. 10; Table 2). Barite that formed
430 with Stage 2A enargite has $\delta^{34}\text{S}$ values of +22.3 to +23.8 ‰ and $\delta^{18}\text{O}_{\text{SO}_4}$ values of -0.2 to +1.3
431 ‰ (Fig. 10; Table 2).

432 *Void-filling dickite (Stage 3)* – Stage 3 massive dickite (Fig. 5G) has $\delta^{18}\text{O}$ values of -1.1 to +2.1
433 ‰ and δD values of -121 to -103 ‰ (Fig. 11; Table 2).

434 *Gypsum veins (Stage 4)* – Gypsum in Stage 4 veins (Figs. 4, 5H) has $\delta^{34}\text{S}$ values between +11.4
435 and +19.2 ‰ and $\delta^{18}\text{O}_{\text{SO}_4}$ values between +0.5 and +3.4 ‰ (Fig. 10; Table 2). These isotopic
436 compositions overlap those of Stage 1A and Stage 2A sulfates (Fig. 10; Table 2).

437 *Interpretation*

438 *Total sulfur composition and redox state of the magmatic-hydrothermal system* – Co-
439 precipitated, magmatic-hydrothermal alunite and pyrite in HS epithermal deposits typically
440 retain sulfur isotope equilibrium (Rye et al., 1992; Rye, 2005). Assuming sulfate-sulfide
441 equilibrium, the approximate redox state ($\text{H}_2\text{S}/\text{SO}_4$) of the fluid and the $\delta^{34}\text{S}$ value of total
442 sulfur ($\delta^{34}\text{S}_{\Sigma\text{S}}$) were estimated using a $\delta^{34}\text{S}_{\text{alunite}}$ and $\delta^{34}\text{S}_{\text{pyrite}}$ vs. $\Delta\delta^{34}\text{S}_{\text{alunite-pyrite}}$ plot (Field and
443 Gustafson, 1976; Kusakabe et al., 1984; Field et al., 2005). The resulting $\delta^{34}\text{S}_{\Sigma\text{S}}$ value during
444 early Stage 1A advanced argillic alteration at Mt Carlton is approximately -1.3 ‰ (Fig. 10).
445 This value is similar to total sulfur values reported from HS epithermal and porphyry Cu
446 systems related to I-type granites, and is consistent with a magmatic source for the sulfur
447 (Ohmoto and Rye, 1979; Hedenquist and Lowenstern, 1994; Arribas, 1995a). The approximate
448 values of $X_{\text{H}_2\text{S}}$ and X_{SO_4} are 0.85 and 0.15, respectively, which correspond to a $\text{H}_2\text{S}/\text{SO}_4$ of 5.7.
449 This value is within the range of $\text{H}_2\text{S}/\text{SO}_4$ values reported for other HS epithermal deposits (4
450 ± 2 ; Rye et al., 1992; Hedenquist et al., 1994; Arribas, 1995a).

451 *Geothermometry* – Sulfur isotope fractionation between alunite and pyrite in the magmatic-
452 hydrothermal environment is argued to be a robust geothermometer (Rye et al., 1992; Rye,
453 2005). Temperatures were calculated for 10 Stage 1A alunite – pyrite pairs from Mt Carlton,
454 using the fractionation factor of Ohmoto and Rye (1979), yielding an average of 207°C. Alunite
455 – pyrite pairs from the V2 pit show a temperature range of 152 to 251°C (n = 8; Fig. 8; Table

456 2), which compares well with homogenization temperatures measured in secondary fluid
457 inclusions in quartz phenocrysts (163-264°C, average 222°C; Figs. 8, 9; Table 1).

458 Geothermometry based on oxygen isotope fractionation between SO₄ and OH in alunite
459 was calculated using the fractionation factor of Stoffregen et al. (1994), yielding unrealistically
460 low equilibrium temperatures of -91 to 81°C (average 22°C) for Stage 1A disseminated alunite
461 and -31 to 95°C (average 18°C) for Stage 1B plumose alunite (not shown in Table 2). We
462 interpret this to be caused by retrograde isotopic exchange between water and OH in alunite,
463 as previously suggested by Rye et al. (1992) and Rye (2005).

464 Enargite and sphalerite are expected to show similar isotopic partitioning behavior due
465 to their structural similarities (Hedenquist et al., 2017). To estimate the equilibrium
466 temperatures of Stage 2A enargite – barite pairs at Mt Carlton, an equation for sulfur isotope
467 fractionation between sphalerite and barite was used (Ohmoto and Rye, 1979), because of the
468 lack of such parameters for enargite. One enargite – barite pair from V2 pit yielded a
469 temperature of 211°C, while four pairs from A39 pit yielded temperatures between 150 and
470 178°C (Fig. 8; Table 2). These temperatures compare well with homogenization temperatures
471 obtained from primary fluid inclusions in barite (157-233°C in V2 pit, 116-183°C in A39 pit;
472 Figs. 8, 9; Table 1), which adds confidence to the results.

473 *Fluid modelling* – The oxygen and hydrogen isotope composition of fluids involved in the
474 formation of Stage 1A alunite were calculated using the fractionation factor of Stoffregen et al.
475 (1994), using temperatures of 200 to 250°C based on $\Delta\delta^{34}\text{S}_{\text{alunite-pyrite}}$ geothermometry, fluid
476 inclusion microthermometry, and published temperature constraints on alunite-bearing mineral
477 assemblages (Reyes, 1990; Hedenquist et al., 1998; Watanabe and Hedenquist, 2001;
478 Hedenquist and Taran, 2013). The calculated fluid compositions are intermediate between
479 magmatic vapor discharged from high-temperature volcanic fumaroles (Giggenbach, 1992)

480 and early Permian meteoric water at Mt Carlton ($\delta^{18}\text{O} = -17 \pm 3 \text{‰}$ and $\delta\text{D} = -130 \pm 30 \text{‰}$; Fig.
481 11). The composition of meteoric water was estimated from current $\delta^{18}\text{O}$ and δD values of
482 global precipitation at a latitude of 60° , the latter based on tectonic paleo-reconstructions
483 (Scotese, 2001) indicating that northeastern Australia had a latitude of approximately $S60^\circ$ in
484 the early Permian. This estimate does not take into account $\delta^{18}\text{O}$ and δD variations due to the
485 paleo-climate or paleo-elevation (e.g., Bradley, 1999; Ravelo and Hillaire-Marcel, 2007).
486 Based on mixing between a typical magmatic vapor and early Permian meteoric water, Stage
487 1A alunite-forming fluids at Mt Carlton had a 50 to 60 % meteoric component (Fig. 11). The
488 magmatic vapor contribution to the hydrothermal condensate resulted in sufficient reactivity to
489 cause strong alteration and acid leaching of the host rocks. In general, the Stage 1A alunite-
490 forming fluids at Mt Carlton ($\delta^{18}\text{O} = -11$ to -2‰ ; $\delta\text{D} = -96$ to -73‰ ; Fig. 11) are isotopically
491 light compared to similar acidic fluids associated with hypogene advanced argillic alteration
492 elsewhere ($\delta^{18}\text{O} = -3$ to $+11 \text{‰}$; $\delta\text{D} = -80$ to -10‰ ; Arribas, 1995a, and references therein).
493 This is consistent with Mt Carlton having formed at a high paleo-latitude ($\sim S60^\circ$; Scotese,
494 2001), whereas most young porphyry-epithermal systems studied around the world have
495 formed at lower latitudes (Arribas, 1995a; Sillitoe, 2010).

496 Hydrothermal fluids in equilibrium with Stage 1B plumose alunite were modelled using
497 the same methodology and parameters as for Stage 1A disseminated alunite. Stage 1B alunite-
498 forming fluids show $\delta^{18}\text{O}$ - δD systematics similar to those of the Stage 1A fluids, albeit slightly
499 lower δD values, the latter likely explained by progressive degassing of the causative intrusion
500 (Fig. 11; Taylor, 1988; Hedenquist and Richards, 1998). These data indicate that there was also
501 a significant involvement of meteoric water during the formation of the Stage 1B plumose
502 alunite veins. Combined with the higher-than-predicted $\delta^{34}\text{S}$ values, this suggests that plumose
503 alunite at Mt Carlton did not form in a pure magmatic-steam environment, in contrast to the
504 arguments of Rye et al. (1992).

505 The $\delta^{18}\text{O}$ compositions of hydrothermal fluids in equilibrium with Stage 2A barite were
506 calculated using the fractionation factor of Zheng (1999), using temperatures of 150 to 200°C
507 based on thermometric data from fluid inclusions and enargite – barite sulfur isotope pairs. The
508 calculated barite-forming fluids have $\delta^{18}\text{O}$ values that generally overlap those of the most dilute
509 Stage 1A alunite-forming fluids, indicating a 60 to 70 % meteoric component (Fig. 11).

510 A temperature range of 120-150°C was used to calculate the $\delta^{18}\text{O}$ and δD compositions
511 of fluids in equilibrium with Stage 3 dickite, based on the late paragenesis of dickite, mineral
512 stability data (Reyes, 1990), and previously published isotopic modelling of similar dickite-
513 forming fluids at the Lepanto HS epithermal deposit, Philippines (Hedenquist et al., 1998).
514 Calculations were done using the fractionation factors of Sheppard and Gilg (1996), and Gilg
515 and Sheppard (1996). The results indicate that fluid $\delta^{18}\text{O}$ values overlap, or are slightly lower
516 than, those of the earlier alunite-forming and barite-forming fluids, which was likely caused by
517 the involvement of ≥ 70 % meteoric water (Fig. 11). However, the δD values of the Stage 3
518 dickite-forming fluids are distinctly lower than those of the Stage 1 alunite-forming fluids (Fig.
519 11). In addition to a greater meteoric water component, we interpret this to potentially reflect
520 a combination of 1) the magmatic component during Stages 2 and 3 being a liquid, isotopically
521 distinct (lower δD) from magmatic vapor during Stage 1; and 2) continuous degassing of the
522 causative intrusion over the life of the magmatic-hydrothermal system, depleting both melt and
523 exsolved fluids in δD with time (Fig. 11; Taylor, 1988; Hedenquist and Richards, 1998).

524 **Discussion**

525 *Evolution of the magmatic-hydrothermal system at Mt Carlton*

526 *Early advanced argillic alteration* – Stage 1A alteration assemblages at Mt Carlton exhibit
527 stable isotope systematics similar to those of advanced argillic alteration documented at other
528 HS epithermal deposits, consistent with a magmatic-hydrothermal origin (Rye et al., 1992;

529 Rye, 2005). As is the case with porphyry-epithermal systems in volcanic arcs worldwide, the
530 causative intrusion at Mt Carlton most likely crystallized from an I-type, magnetite-series
531 magma (Ishihara, 1981) saturated with sulfur- and metal-rich aqueous fluids (Sillitoe, 2010).
532 Saline supercritical fluid exsolved from the intrusion and initially accumulated at deeper levels
533 (e.g., below the brittle-ductile transition; Fournier, 1999). This fluid subsequently separated
534 into a hypersaline liquid and a low-salinity vapor (Henley and McNabb, 1978). The dense
535 hypersaline liquid likely produced potassic alteration and associated mineralization in and
536 around magmatic apophyses from the intrusion (Hedenquist et al., 1998; Sillitoe, 2010),
537 whereas the buoyant low-salinity vapor ascended along steep normal faults and fracture
538 networks that formed in the Mt Carlton area during early Permian rifting (Fig. 12). Lakes
539 developed at the surface within localized rift basins, into which volcanic as well as organic
540 sediments were deposited (Sahlström et al., 2018). During extensional episodes, meteoric water
541 penetrated to depth via open veins and absorbed rising magmatic vapor plumes (Fig. 12). The
542 resulting vapor condensate became increasingly acidic upon cooling, first by
543 disproportionation of SO_2 to form H_2S and H_2SO_4 below $\sim 400^\circ\text{C}$ (Sakai and Matsubaya, 1977;
544 Arribas, 1995a), then by progressive dissociation of H_2SO_4 and HCl at lower temperatures
545 ($< 300^\circ\text{C}$; Hedenquist and Taran, 2013). Alunite would have started to form at temperatures
546 $\sim 300^\circ\text{C}$ (Arribas, 1995a). Upon cooling to temperatures below 200°C at a high condensate/rock
547 ratio ($> 10:1$), alunite started to dissolve, with intense leaching at $\text{pH} \approx 1$ forming cores of silicic
548 alteration in and around the fluid channels (Hedenquist and Taran, 2013). Away from the feeder
549 structures, fluid flow was less focused and increasingly diluted, causing alunite and then dickite
550 – kaolinite to form (Stoffregen, 1987; Hedenquist and Taran, 2013). These reactions produced
551 the observed silicic \rightarrow quartz – alunite \rightarrow quartz – dickite – kaolinite zoned alteration halo
552 around Mt Carlton, which is typical of HS epithermal deposits globally (Steven and Ratté,
553 1960; Arribas, 1995a). Progressively greater dilution of the magmatic condensate by meteoric

554 water led to lower temperatures and more oxidizing conditions toward the shallow and distal
555 parts of the Mt Carlton lithocap, as indicated by the fluid inclusion and stable isotope data
556 (Figs. 8, 9), plus the decreasing Na-content in alunite (Stoffregen and Cygan, 1990; Chang et
557 al., 2011; Sahlström et al., 2018) and the disappearance of APS minerals and pyrite in the
558 alunite assemblage (Hedenquist et al., 1998; Watanabe and Hedenquist, 2001).

559 Within the volcano-lacustrine setting in which Mt Carlton formed, the paleo-water table
560 was essentially at the paleo-surface at the time of magmatic-hydrothermal activity (Fig. 12).
561 As such, the absence of a vadose zone would have inhibited atmospheric oxidation of
562 ascending, magmatic H₂S and formation of steam-heated acidic waters. This may explain why
563 steam-heated advanced argillic alteration (Rye et al., 1992; Hedenquist et al., 2000) is absent
564 from the top of the Mt Carlton deposit, at least on the basis of the evidence within the rock
565 units in the A39 pit, which represents the shallowest part of the Mt Carlton system.

566 *Deposition of plumose alunite veins* – Coarse-grained, banded, plumose alunite was deposited
567 into open space along tension fractures during the final stages of advanced argillic alteration at
568 Mt Carlton. The genetic environment that produces such alunite is poorly understood,
569 particularly with respect to the oxidation process and the transport and precipitation
570 mechanisms of alunite (Rye et al., 1992). Although the Stage 1B plumose alunite at Mt Carlton
571 has different characteristics than Stage 1A disseminated alunite, the isotopic compositions of
572 this mineral are not consistent with a magmatic-steam origin (c.f., Marysvale, Utah;
573 Cunningham et al., 1984; Rye et al., 1992; Rye, 2005; Landis and Rye, 2005).

574 Similar atypical isotopic signatures in plumose alunite in the Pierina HS epithermal
575 deposit, Peru, were interpreted to reflect unusually slow ascent of magmatic vapor from the
576 intrusive source to the site of alunite deposition (Fifarek and Rye, 2005). Longer residence
577 times during fluid transport would allow for 1) sulfur isotope exchange between H₂S and SO₄,

578 producing higher-than-predicted $\delta^{34}\text{S}$ values in alunite, and 2) uptake of groundwater,
579 producing lower-than-predicted $\delta^{18}\text{O}_{\text{SO}_4}$, $\delta^{18}\text{O}_{\text{OH}}$ and δD values in alunite (Fifarek and Rye,
580 2005). Slow magmatic vapor transport may reflect lithological and hydrodynamic barriers (e.g.,
581 Pierina; Fifarek and Rye, 2005), or simply distance between the degassing intrusion and the
582 lithocap (Rye, 2005). The latter scenario would be consistent with the observed zonation in the
583 $\delta^{34}\text{S}$ values of plumose alunite at Mt Carlton, where the highest values are seen distal to the
584 main feeder structures (Table 2). Because the Mt Carlton deposit formed in an extensional
585 tectonic setting during subsidence and burial, it is unlikely that the magmatic-hydrothermal
586 system underwent telescoping during its lifetime (c.f., Sillitoe, 1994). Therefore, the causative
587 intrusion was likely separated from the near-surface Mt Carlton HS epithermal deposit by a
588 significant distance at the time of mineralization (>1-2 km; Sillitoe, 1999).

589 An alternative, and perhaps simpler interpretation of the observed isotopic
590 compositions of plumose alunite at Mt Carlton is that they reflect a formation environment
591 unrelated to a purely magmatic-steam origin. This might have involved alunite formation
592 within a magmatic-hydrothermal environment similar to that of Stage 1A, but under more
593 oxidized conditions ($\delta^{34}\text{S}_{\text{alunite}}$ closer to $\delta^{34}\text{S}_{\Sigma\text{S}}$, no sulfide). In this case, the difference between
594 the two alunite types at Mt Carlton would largely be textural, reflecting alunite formed as wall
595 rock replacement (Stage 1A disseminated alunite) or as infill in open fractures (Stage 1B
596 plumose alunite). The greater abundance of plumose alunite veins at Mt Carlton than at most
597 other HS epithermal deposits may be attributed to the extensional tectonic setting. Overall, the
598 evidence from alunite at Mt Carlton further highlights the potential for transitions between the
599 magmatic-hydrothermal (slow vapor ascent, sulfur isotope equilibrium, extensive vapor
600 condensation) and magmatic-steam (rapid vapor ascent, sulfur isotope disequilibrium, no
601 condensation) environments in these often very dynamic volcanic settings.

602 *Epithermal mineralization* – The fluids that form HS Au ± Ag ± Cu mineralization at
603 epithermal deposits are argued to be low- to moderate-salinity aqueous liquids, sulfur- and
604 metal-rich, which are generated during the waning stages of magmatism and cause white mica
605 (sericitic) alteration below the lithocap environment (e.g., Shinohara and Hedenquist, 1997;
606 Hedenquist et al., 1998; Sillitoe and Hedenquist, 2003; Heinrich et al., 2004; Heinrich, 2005;
607 Pudack et al., 2009; Sillitoe, 2010). Such a mineralizing fluid, which originally would have had
608 an IS state at deeper levels and higher temperature, entered the lithocap at Mt Carlton, where
609 the limited buffering capacity of the leached volcanic host rocks allowed the fluid to evolve to
610 a HS state upon cooling (Einaudi et al., 2003). This fluid evolution is evident in the paragenetic
611 transition from IS state minerals (tennantite – tetrahedrite – goldfieldite) to enargite in the early
612 Stage 2A ores (Fig. 4; Sahlström et al., 2018).

613 Precipitation of Au and other metals in the epithermal environment occurs due to sharp
614 decreases in metal solubility, commonly as a result of boiling and/or fluid mixing and cooling
615 (e.g., Giggenbach and Stewart, 1982; Heinrich, 2005; Sillitoe, 2010). The fluid inclusion data
616 for Stage 2 of Mt Carlton are consistent with mineralization having formed primarily due to
617 the mixing of metalliferous, magmatic-hydrothermal liquid with vapor-heated (~150°C)
618 meteoric water (Fig. 9). This initially produced fracture- and fissure-filling Au-Cu ore within
619 the structural feeders of the deposit (Fig. 12). Subsequently, the mineralizing fluids dispersed
620 laterally into the accumulating volcano-lacustrine sedimentary sequence and vented into the
621 overlying lakes, producing syngenetic Ag – barite ore and associated amorphous silica
622 sediment, spherulitic pyrite, skeletal enargite – luzonite, and various synsedimentary
623 deformation textures (Figs. 6B, 6D-E, 12). Equivalent ore textures were produced within open
624 fractures in the structural roots of the rift basins (Sahlström et al., 2018). The interaction
625 between hydrothermal fluids and organic material within the shallow volcano-lacustrine
626 sediments has been suggested as a cause of the enrichment of some elements (Ge, Ga and In)

627 to economically significant concentrations in the ores (Fig. 12; Sahlström et al., 2017). The
628 overprinting of younger, IS state mineral assemblages (Stages 2B and 2C; Figs. 4, 6C) on the
629 initial enargite ores reflect an evolution of the mineralizing fluids, returning to lower sulfidation
630 state, less oxidized and more neutral pH conditions over time, a commonly seen pattern
631 (Arribas et al., 1995; Jannas et al., 1999), possibly due to interaction with the wall rock (Einaudi
632 et al., 2003; Sillitoe and Hedenquist, 2003).

633 *Late-stage hydrothermal void fill, and post-mineralization events* – The widespread occurrence
634 of late, massive dickite (Stage 3; Figs. 4, 5G) records the final stages of magmatic-
635 hydrothermal activity at Mt Carlton. At this point, the causative intrusion was increasingly
636 degassed, and there was a gradual transition from magmatic-dominant to meteoric-dominant
637 conditions in the epithermal environment (Fig. 11).

638 Extensive illite – montmorillonite alteration and abundant gypsum veins formed during
639 post-mineralization extensional deformation (Stage 4; Figs. 4, 5H). Gypsum sulfate likely came
640 from dissolution of magmatic-hydrothermal anhydrite formed during Stage 1A advanced
641 argillic alteration (Fig. 5E), as oxidation of pyrite to produce sulfate would tend to result in
642 negative $\delta^{34}\text{S}_{\text{sulfate}}$ values (Rye, 2005). The common occurrence of cast textures in the altered
643 rocks at Mt Carlton (Fig. 5F) indicates that original anhydrite was present and subsequently
644 dissolved due to meteoric water circulation during deformation. The dissolved sulfate was then
645 redeposited as gypsum within open space near the main fault zones, allowing the gypsum to
646 maintain the heavy $\delta^{34}\text{S}$ signature of the anhydrite. Similar processes in which soluble sulfate
647 minerals are dissolved and reprecipitated as gypsum veinlets are commonly observed in
648 modern stratovolcano environments (Zimbelman et al., 2005). In addition, magmatic-
649 hydrothermal anhydrite is predicted to be more abundant in lithocaps than what is observed in
650 nature (Hedenquist and Taran, 2013), with this discrepancy explained by dissolution of
651 anhydrite during late- to post-mineralization fluid flow (Cooke and Simmons, 2000;

652 Hedenquist and Taran, 2013). The textural and isotopic data from Mt Carlton support this
653 explanation.

654 *Conclusions*

655 The mineralogy and zonation of alteration and mineralization observed at Mt Carlton, plus the
656 stable isotope and fluid inclusion data, are generally consistent with that noted for younger,
657 subaerial HS epithermal deposits (Arribas, 1995a; Hedenquist et al., 2000). There was a strong
658 contribution of magmatic water and sulfur that generated a reactive hydrothermal system,
659 manifested in the laterally extensive, zoned advanced argillic alteration halos to leached host
660 rocks, and in monomineralic veins of banded, plumose alunite, followed by deposition of high-
661 grade, HS epithermal mineralization. However, Mt Carlton represents a marked variation on
662 this well-known ore-forming environment. Here, syn-hydrothermal extension resulted in a
663 localized sub-lacustrine setting in which syngenetic mineralization and associated paleosurface
664 features were formed and preserved since the Permian.

665 Due to the shallow-water vent component to the Mt Carlton system (Fig. 12), other
666 obvious analogs are sulfide deposits that form in shallow-submarine volcanic arc and backarc
667 settings and exhibit advanced argillic alteration, HS ore mineral assemblages, epithermal trace
668 element suites, and evidence for a magmatic fluid component in the hydrothermal system
669 (Sillitoe et al., 1996; Hannington et al., 1999; Hannington and Herzig, 2000). A particularly
670 relevant example is the epithermal mineralization at the Pliocene Lerokis and Kali Kuning
671 deposits, Wetar Island, Indonesia, where Au and Ag ore is associated with stratiform barite
672 sand, which is underlain by polymetallic massive sulfide bodies (Sewell and Wheatley, 1994;
673 Scotney et al., 2005). Other comparable examples include active submarine hydrothermal ore-
674 forming systems such as the enargite – luzonite chimneys at North Su and Kaia Natai
675 Volcanoes (Manus Backarc Basin; Dekov et al., 2016), and the barite – polymetallic sulfide

676 mineralization at Palinuro Seamount (Aeolian Arc, Tyrrhenian Sea; Minniti and Bonnavia,
677 1984; Tufar, 1991), Brothers Volcano (Kermadec Arc; de Ronde et al., 2005, 2011) and the
678 Hine Hina hydrothermal field (Lau Backarc Basin; Herzig et al., 1993, 1998; Fouquet et al.,
679 1993).

680 The Mt Carlton deposit most likely formed at a water depth of no more than a few tens
681 of meters (as supported by the presence of coal and fossilized wood fragments in the host
682 sediments; Sahlström et al., 2018), much shallower than the depths at which seafloor massive
683 sulfide deposits in submarine volcanic arc settings generally form (i.e., hundreds of meters to
684 >1 km). Yet, Mt Carlton shares several key textural, mineralogical and geochemical features
685 with such deposits. These include 1) a similar ore body morphology, with fracture- and fissure-
686 filling mineralization within structural feeders at depth and syngenetic, stratiform
687 mineralization at surface (e.g., Sato, 1977); 2) massive accumulations of Pb-Sr-enriched barite
688 that cap the stratiform ore (Bonatti et al., 1972; Sewell and Wheatley, 1994; Sillitoe et al., 1996;
689 Momoshima et al., 1997; Scotney et al., 2005); 3) laminations of amorphous silica, similar to
690 those that accumulate in hyperacidic crater lake settings (Sillitoe, 2015), occurring either
691 intimately associated with, or distal to, mineralization (e.g., Petersen et al., 2002; de Ronde et
692 al., 2011); 4) spherulitic and skeletal ore textures, indicative of rapid mineral crystallization
693 during venting (e.g., Tufar, 1991; Fouquet et al., 1993; Scotney et al., 2005; Xu and Scott;
694 2005; Dekov et al., 2016); 5) slumping textures and other synsedimentary deformation textures
695 (Tufar, 1991; Scotney et al., 2005); 6) no development of steam-heated advanced argillic
696 alteration due to the lack of a vadose zone (Sillitoe et al., 1996); and 7) evidence for a larger
697 proportion of surface water (in this case, meteoric water) in the magmatic-hydrothermal system
698 than in most subaerial HS epithermal deposits (Fig. 11), likely due to the sub-lacustrine setting
699 of Mt Carlton.

700 The close relationship between paleosurface features and epithermal mineralization has
701 been recognized for over a century (Sillitoe, 2015, and references therein), and has been
702 successfully applied in mineral exploration (e.g., the discovery of the Fruta del Norte
703 epithermal Au-Ag deposit, Ecuador, immediately beneath a silica sinter horizon; Leary et al.,
704 2016). While paleosurface features associated with lakes have been documented near other HS
705 epithermal deposits (e.g., Akaiwa, Japan, Arribas, 1995b; Yanacocha, Peru, Longo et al.,
706 2010), the study of Mt Carlton highlights that they may occur in direct contact with economic
707 mineralization. The lacustrine paleosurface features documented at Mt Carlton (e.g., laminated
708 amorphous silica sediment, massive barite accumulations, and stratiform sulfides exhibiting
709 venting textures and synsedimentary deformation textures), as well as the associated
710 geochemical anomalies (e.g., Ba-Pb-Sr anomaly caused by barite), may therefore be indicators
711 of potential concealed HS epithermal mineralization in volcano-sedimentary sequences
712 elsewhere. The presence of such paleosurface features, and their application in mineral
713 exploration, will be limited to regions with high preservation potential (Sillitoe, 2015). In
714 particular, extensional settings conducive to HS epithermal formation and in which ore deposits
715 would have been rapidly buried, such as backarc rifts, are especially deserving of attention.

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1043

1044 **Figure captions**

1045 Fig. 1: **(A)** First-order geologic subdivisions of northeastern Queensland, Australia.
1046 Abbreviations: BB – Bowen Basin; GI – Georgetown Inlier; HF – Hodgkinson Formation; MO
1047 – Mossman Orogen; NEO – New England Orogen; PPC – Post-Permian Cover; TO – Thomson
1048 Orogen. Modified from McKibbin et al. (2017) and Cheng et al. (2018). **(B)** Geologic map of
1049 the northern Bowen Basin and New England Orogen, showing the locations of the known
1050 porphyry, epithermal and orogenic deposits and prospects (compiled from Geological Survey
1051 of Queensland databases). Grid – UTM zone 55; datum – GDA94. Modified from Sahlström
1052 et al. (2017, 2018).

1053

1054 Fig. 2: 3D model of the ore zones at Mt Carlton, showing the zonation of Au, Cu, Zn, Pb and
1055 Ag across the deposit. The stippled lines indicate the surface traces of the sections shown in
1056 Figure 3. Modified from Sahlström et al. (2018).

1057

1058 Fig. 3: Sections through Mt Carlton, showing the distribution of stratigraphic units, major
1059 structures, alteration assemblages and mineralization. **(A)** NE-SW long section oriented
1060 parallel to the Western ore zone. **(B)** NW-SE cross section through the SW V2 pit.
1061 Abbreviations: alu – alunite; anh – anhydrite; APS – aluminium-phosphate-sulphate minerals;
1062 brt – barite; chl – chlorite; dck – dickite; gyp – gypsum; hbx – hydrothermal breccia; hem –
1063 hematite; ill – illite; kln – kaolinite; mnt – montmorillonite; prl – pyrophyllite; py – pyrite; qtz
1064 – quartz; rt – rutile. Modified from Sahlström et al. (2017, 2018).

1065

1066 Fig. 4: Paragenetic sequence of alteration and ore minerals at Mt Carlton. Modified from
1067 Sahlström et al. (2018).

1068

1069 Fig. 5: Photographs and photomicrographs of selected alteration assemblages from Mt Carlton.
1070 **(A)** Stage 1A quartz – alunite – dickite – pyrite alteration developed in a quartz – feldspar
1071 phytic rhyodacite (Unit 3). The Stage 1A assemblage is crosscut by a younger Stage 1B
1072 plumose alunite vein. **(B)** Aggregate of Stage 1A alunite crystals that have replaced a pre-
1073 existing feldspar crystal (transmitted light, cross polarized). **(C)** Stage 1B banded, plumose
1074 alunite vein. The alunite vein has locally been overprinted by Stage 3 dickite – pyrite veinlets.
1075 **(D)** Plumose texture in a Stage 1B alunite vein (transmitted light, cross polarized). **(E)**
1076 Aggregate of tabular Stage 1A anhydrite crystals developed within open space in a silicic
1077 altered rhyodacite porphyry. **(F)** Similar alteration as in E), but with the anhydrite having been
1078 completely dissolved, leaving behind distinctive casts in the rock. **(G)** Stage 2A enargite –
1079 pyrite mineralization overprinted by void fill made up of Stage 3 massive dickite with pyrite.
1080 **(H)** Stage 4 illite – montmorillonite alteration developed in a massive dacite tuff (Unit 4B) in
1081 the northern wall of V2 pit. Veins made up of fibrous gypsum with a well-developed shear
1082 fabric occur near a D₂ fault zone. Abbreviations from Figure 3, and Eng – enargite.

1083

1084 Fig. 6: Photographs and photomicrographs of selected ore textures from Mt Carlton. **(A)** Silicic
1085 hydrothermal breccia cemented by Stage 2A enargite – pyrite mineralization (from V2 pit). **(B)**
1086 Skeletal aggregates of Stage 2A enargite and luzonite from A39 pit (transmitted light, plane
1087 polarized; c.f., Dekov et al., 2016, their Figure 3). **(C)** Composite vein from V2 pit, illustrating
1088 part of the temporal sequence of alteration and mineralization at Mt Carlton. A Stage 1A quartz
1089 – alunite – pyrite altered rhyodacite porphyry is crosscut by a Stage 1B plumose alunite vein.

1090 The alunite vein served as a pathway for later mineralizing liquid, which deposited Stage 2A
1091 enargite ore and, subsequently, Stage 2B sphalerite – pyrite ore. **(D)** Sedimentary ore from A39
1092 pit. The sediments are predominantly made up of finely laminated quartz. The siliceous
1093 sediments incorporated mineralized material from Stage 2A (fine-grained in photo), Stage 2B
1094 and Stage 3, as well as volcanic sediment in the form of quartz phenocrysts. **(E)** Layered
1095 massive sulfide mineralization consisting of spherulitic Stage 3 pyrite and dickite, within
1096 volcano-lacustrine sediment (Unit 4A; from SW corner of V2 pit). Abbreviations from Figure
1097 3, and Eng – enargite; Luz – Luzonite; Sp – sphalerite.

1098

1099 Fig. 7: Photomicrographs of the studied fluid inclusion assemblages from Mt Carlton. **(A)**
1100 Igneous quartz phenocryst in a hydrothermally altered quartz – feldspar phyric rhyodacite. The
1101 quartz crystal exhibits fractures and corrosion textures, which have been infilled with Stage 1A
1102 quartz and alunite. Secondary fluid inclusions are locally developed in trails propagating from
1103 such fractures or from quartz crystal boundaries (transmitted light, cross polarized). **(B)**
1104 Coarse-grained, euhedral barite crystals occurring in the halo of a Stage 2A enargite veinlet.
1105 The barite crystals grew perpendicular to the enargite veinlet, and contain both primary and
1106 secondary fluid inclusions (transmitted light, plane polarized; see Sahlström et al., 2017, their
1107 Figure 4E). **(C)** Trail of secondary fluid inclusions developed within an igneous quartz
1108 phenocryst (transmitted light, plane polarized). **(D)** Primary fluid inclusion within a barite
1109 crystal associated with Stage 2A enargite mineralization (transmitted light, plane polarized).
1110 Abbreviations from Figure 3, and FI – fluid inclusion.

1111

1112 Fig. 8: Map of the Mt Carlton open pits (c.f., Fig. 2), showing the surface projections of samples
1113 used for geothermometry. Homogenization temperatures (Th) for fluid inclusions in quartz and

1114 barite, and temperatures calculated from $\Delta\delta^{34}\text{S}_{\text{sulfate-sulfide}}$ values, are listed. All temperatures
1115 rounded to the nearest 5°C. Sample HC09DD025-120 (alunite – pyrite; A39 pit) is shown, but
1116 not annotated, due to the unrealistically high calculated equilibrium temperature of this distal
1117 sample (330°C).

1118

1119 Fig. 9: Plots of fluid inclusion data. **(A)** Frequency histogram of homogenization temperatures
1120 (Th) of secondary fluid inclusions in quartz phenocrysts. **(B)** Frequency histogram of Th of
1121 primary fluid inclusions in barite associated with Stage 2A mineralization. **(C)**
1122 Homogenization temperature (Th) vs. final ice melting temperature (Tm) plot of secondary
1123 fluid inclusions in quartz phenocrysts. **(D)** Th vs. Tm plot of primary fluid inclusions in barite.
1124 The observed Th vs. Tm trends are interpreted to reflect mixing between magmatic-derived
1125 hydrothermal fluids and vapor-heated groundwater (~150°C, nil salinity). See text for detailed
1126 explanation.

1127

1128 Fig. 10: Summary plot of $\delta^{34}\text{S}$ values of sulfate, sulfide and sulfosalt minerals.

1129

1130 Fig. 11: Plot of $\delta^{18}\text{O}$ vs. δD values of Stage 1A and Stage 1B alunite ($\delta^{18}\text{O}_{\text{SO}_4}$ plotted) and Stage
1131 3 dickite from Mt Carlton. The calculated compositions of hydrothermal condensates in
1132 equilibrium with Stage 1A and Stage 1B alunite are plotted together with compositions of
1133 magmatic vapor discharged from high-temperature volcanic fumaroles (Giggenbach, 1992),
1134 water dissolved in felsic magma (Taylor, 1992), estimated early Permian meteoric water
1135 (latitude ~S60°; Scotese, 2001; this study), present-day Mt Carlton meteoric water (latitude
1136 ~S20°; $\delta^{18}\text{O} = -5.8$ to -5.5 ‰, $\delta\text{D} = -39$ to -37 ‰; F. Sahlström, unpub. data), and acidic fluids
1137 related to advanced argillic alteration of worldwide HS epithermal deposits (Arribas, 1995, and

1138 references therein). The trends defined by Stage 1A and Stage 1B alunite-forming fluids
1139 indicate condensation of magmatic vapor by local meteoric water (~50 to 60 % meteoric
1140 component). The Stage 2A barite ($\delta^{18}\text{O}$ values only) and Stage 3 dickite liquid compositions
1141 are also plotted, indicating an increasing meteoric water component with time (~60 to 70 %
1142 meteoric water during barite formation, ≥ 70 % during dickite formation). Magmatic liquid
1143 during Stages 2 and 3 was isotopically distinct (lower δD) from Stage 1 magmatic vapor.
1144 Additionally, progressive degassing of the causative intrusion depleted both melt and exsolved
1145 fluids in δD with time (Taylor, 1988; Hedenquist and Richards, 1998). See text for detailed
1146 explanation.

1147

1148 Fig. 12: Conceptual genetic model for Mt Carlton, integrating geological, structural,
1149 mineralogical and geochemical data and observations (Sahlström et al., 2017, 2018; this study).
1150 Mt Carlton formed in a shallow volcano-sedimentary rift environment, in a sub-lacustrine
1151 setting, where magmatic initially vapor condensed into meteoric water, followed by magmatic
1152 liquid diluted by meteoric water. See text for detailed explanation.

1153

1154 **Table captions**

1155 Table 1: Summary of heating and freezing results for fluid inclusions from Mt Carlton.
1156 Coordinates: Grid – UTM zone 55; datum – GDA94.

1157

1158 Table 2: Stable isotope data (reported in per mil, ‰) and geothermometry for minerals from
1159 Mt Carlton. Coordinates: Grid – UTM zone 55; datum – GDA94.

1160

