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4   **Magnetite texture and trace-element geochemistry fingerprint of**

5   **pulsed mineralization in the Xinqiao Cu–Fe–Au deposit, Eastern**

6   **China**

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22   **ABSTRACT**

23   The origin of stratabound deposits in the Middle-Lower Yangtze River Valley Metallogenic

24   Belt (MLYRB), Eastern China, is the subject of considerable debate. The Xinqiao Cu–Fe–Au

25   deposit in the Tongling ore district is a typical stratabound ore body characterized by multi-stage

26   magnetite. A total of six generations of magnetite have been identified. Mt1 is commonly replaced

27   by porous Mt2, and both are commonly trapped in the core of Mt3, which is characterized by both

28   core-rim textures and oscillatory zoning. Porous Mt4 commonly truncates the oscillatory zoning

29   of Mt3, and Mt5 is characterized by 120° triple junction texture. Mt1 to Mt5 are commonly

30   replaced by pyrite that coexists with quartz, whereas Mt6, with a fine-grained foliated and

31   needle-like texture, commonly cuts the early pyrite as veins and is replaced by pyrite that coexists

32   with calcite. The geochemistry of the magnetite suggests that they are hydrothermal in origin. The

33   microporosity of Mt2 and Mt4 magnetite, their sharp contacts with Mt1 and Mt3, and lower trace

34   element contents (e.g., Si, Ca, Mg, and Ti) than Mt1 and Mt3 suggest that they formed via coupled

35   dissolution and reprecipitation of the precursor Mt1 and Mt3 magnetite, respectively. This was

36   likely caused by high-salinity fluids derived from intensive water-rock interaction between the

37   magmatic-hydrothermal fluids associated with the Jitou stock and Late Permian metalliferous

38   black shales. The 120° triple junction texture of Mt5 suggests it is the result of fluid-assisted

39   recrystallization, whereas Mt6 formed by replacement of hematite as a result of fracturing. The

40   geochemistry of the magnetite suggests that the temperature increased from Mt2 to Mt3, and

41   implies that there were multiple pulses of fluids from a magmatic-hydrothermal system. Therefore,

42   we propose that the Xinqiao stratiform mineralization was genetically associated with multiple

43   influxes of magmatic hydrothermal fluids derived from the Early Cretaceous Jitou stock. This

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44   study demonstrates that detailed texture examination and in-situ trace elements analysis under

45   robust geological and petrographic frameworks can effectively constrain the mineralization

46   processes and ore genesis.

47

48   **Keywords:** Magnetite; Stratabound mineralization; Xinqiao Cu–Fe–Au deposit;

49   Middle–Lower Yangtze River Valley Metallogenic Belt.

INTRODUCTION

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51   Magnetite is a ubiquitous mineral phase in different geological environments and represents a

52   dominant metallic mineral phase in wide variety of hydrothermal ore systems (Nadoll et al. 2014,

53   2015). Octahedral and tetrahedral coordinated positions in the magnetite structure provide host

54   sites for many trace elements (e.g., Al, Mn, Ti, V, Ni, Cr, Zn, Co, Sn, Ga, and Mg) via isovalent

55   and coupled substitutions (Dupuis and Beaudoin 2011; Nadoll et al. 2012; Deditius et al. 2018).

56   The trace-element compositions of hydrothermal magnetite are mainly governed by the

57   composition of the hydrothermal fluids, temperature, oxygen fugacity (fO_2), sulfur fugacity (fS_2),

58   co-crystallized mineral phases and fluid-rock interactions (McIntire 1963; Dare et al. 2012, 2014;

59   Nadoll et al. 2014; Knipping et al. 2015; Huang et al. 2019a, b; Liu et al. 2019; Salazar et al. 2019;

60   Sun et al. 2019). Therefore, a number of studies have focused on the trace-element geochemistry

61   of magnetite to constrain mineralization type and ore genesis (Beaudoin et al. 2007; Dupuis and

62   Beaudoin 2011; Dare et al. 2012, 2014; Nadoll et al. 2012, 2014, 2015), fingerprint the temporal

63   and/or spatial evolution of the ore-forming hydrothermal fluids (e.g., Li et al. 2019; Liu et al.

64   2019), and reconstruct the mineralization processes (e.g., Li et al. 2019; Hu et al., 2020). More

65   importantly, magnetite undergoes dissolution and reprecipitation (DRP), oxy-exsolution, and/or

66   recrystallization to reach chemical and textural re-equilibration, and consequently detailed studies

67   of internal textures are necessary before conducting trace-element analyses (Hu et al. 2014, 2015;

68   Salazar et al. 2019).

69   The Middle-Lower Yangtze River Valley Metallogenic Belt (MLYRB), Eastern China, is one

70   of the most important mining regions in China, and is characterized by abundant large stratabound

71   deposits (e.g., the Dongguashan Cu–Au deposit and the Xinqiao Cu–Fe–Au deposit; Mao et al.

72   2011). The origin of these stratabound deposits is still a matter of debate with a number of models

73   proposed including Late Paleozoic submarine exhalative processes (SEDEX, e.g., Gu et al. 2000;

74   Xu and Zhou 2001), Early Cretaceous magmatic-hydrothermal fluids (skarn-type, e.g., Pan and

75   Done 1999; Mao et al. 2009, 2011; Zhang et al. 2017b, c, 2018; Li et al. 2017, 2018a, 2019), or a

76   combination of the two (e.g., Tang et al. 1998; Zhou et al. 2010; Guo et al. 2011). The Xinqiao

77   Cu–Fe–Au deposit in the Tongling ore district has a proven reserve of 0.5 Mt Cu @ 0.71%, 24.9

78   Mt Fe @ 46%, 11.2 t Au @ 4.7g/t and 248.4 t Ag @ 248.4 g/t (Xu and Zhou 2001), and is

79   dominated by a large-scale stratiform orebody with extensive skarn alteration and unique

80   colloform pyrite (Zhang et al. 2017a). The fact that colloform pyrite in the Xinqiao stratiform

81   orebody is cut by magnetite veins is consistent with a syn-depositional origin of colloform pyrite

82   (Xiao et al. 2016). However, in-situ Sensitive High Resolution Ion Microprobe (SHRIMP) sulfur

83   isotope compositions ($\delta^{34}S_{CDT}$ = –0.6 to 2.7‰, Avg. 0.58‰; Zhang et al. 2017a) and Co/Ni ratios

84   (0.67 to 2.94, Avg.1.66; Zhang et al. 2017a) of the colloform pyrite, similar ages between

85   colloform pyrite (Re–Os age: 136.7 ± 4.6 Ma; Li et al. 2018a) and the quartz diorite of the Jitou

86   stock at Xinqiao (zircon U–Pb age: 139.6 ± 1.5 Ma; Zhang et al. 2017c), coupled with the fact that

87   colloform pyrite cements massive magnetite (Zhang et al. 2017a) suggests a magmatic

88   hydrothermal origin for the colloform pyrite and the occurrence of multi-stage magnetite in the

89   Xinqiao stratiform orebody. Therefore, the complex magnetite at Xinqiao offers an excellent

90   opportunity to examine the detailed paragenesis of the stratiform mineralization in MLYRB.

91   This paper reports data for a paragenetically well-constrained suite of magnetite-bearing

92   samples from the Xinqiao deposit. Using back-scattered electron (BSE) imaging of internal

93   textures combined with electron probe microanalyzer (EPMA) and laser ablation inductively

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94   coupled plasma mass spectrometry (LA-ICP-MS) analysis of magnetite, the textures and

95   geochemistry of the multi-stage magnetite at Xinqiao have been documented. These results

96   provide new constraints on the origin of the Xinqiao stratiform mineralization, and the stratabound

97   deposits of the MLYRB.

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99   **GEOLOGICAL BACKGROUND**

100   **Regional geology**

101   The Middle–Lower Yangtze River Valley Metallogenic Belt is located along the northern

102   margin of the Yangtze craton, which is separated from the North China craton to the north by the

103   Triassic Qinling-Dabieshan orogenic belt, and from the Cathaysian block to the south by a

104   Neoproterozoic suture (Fig. 1a; Li et al. 2014; Zhang et al. 2019). The belt hosts seven ore

105   districts, which are from west to east the Edong, Jiurui, Anqing-Guichi, Luzong, Tongling,

106   Ningwu and Ningzhen districts with more than 200 known polymetallic deposits (Fig. 1a; Ling et

107   al. 2009). The Tongling ore district, located in the central part of MLYRB (Fig. 1a), is the largest

108   Cu–Au–Fe–Mo ore district in the belt and hosts numerous skarn deposits (Tang et al. 1998; Pan

109   and Dong 1999; Lai and Chi 2007; Mao et al. 2011; Sun et al. 2019). More than 50 ore deposits

110   have been discovered in the Tongling district, clustered in five ore fields, namely Tongguanshan,

111   Shizishan, Xinqiao, Fenghuangshan and Shatanjiao, distributed from east to west (Wu et al. 2014;

112   Fig. 1b). The dominant lithologies in the Tongling district are marine and continental sedimentary

113   rocks. Marine sedimentary rocks, including clastic sedimentary rocks, carbonates and evaporites,

114   were deposited in the Silurian to Middle Triassic, with a break in the Early to Middle Devonian.

115   Continental rocks including clastic and volcanic-sedimentary rocks were deposited from the

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116   Middle Jurassic to Quaternary, with a break in the Middle-Late Devonian (Wang et al. 2015; Cao

117   et al. 2017). These sedimentary rocks were deposited on a stable Precambrian basement and are

118   the host rocks for later Cu, Au, Fe, and Mo mineralization (Chang et al. 1991; Zhai et al. 1992).

119   The Lower Permian Qixia Formation and the Lower Triassic Nanlinghu Formation are dominated

120   by limestone and are important ore-hosting strata for the widespread skarn mineralization in the

121   district. The unconformity between the quartz sandstone of the Upper Devonian Wutong

122   Formation and the dolomite and limestone of the Upper Carboniferous Huanglong Formation

123   constrains the economically significant stratiform mineralization in this district, including the

124   Xinqiao and Dongguashan deposits. Structurally, the region contains NE-trending folds and NNE-

125   and NW-trending faults that have influenced the emplacement of the intrusions. Igneous rocks are

126   widely distributed in the region and are dominated by Late Jurassic – Early Cretaceous (ca. 156 to

127   137 Ma) high-K calc-alkaline granitoids (e.g., granite porphyry, granodiorite porphyry and quartz

128   monzodiorite; Fig. 1b; Mao et al. 2011; Wang et al. 2015; Cao et al. 2017). Many small plutons

129   occur as composite stocks, dikes and sills (Du et al. 2015).

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131   **Ore deposit geology**

132   The outcropping sedimentary rocks in the mining area range from the Middle-Upper Silurian

133   sandstone and siltstone to the Lower Triassic Yinkeng Formation limestone, whereas the Lower

134   Carboniferous unit is absent. The NE-trending Dachengshan anticline and the NNE-trending

135   Shenchong syncline are the major structures at Xinqiao, and have influenced the emplacement of

136   the magmatic rocks (Fig. 2a). The dominant igneous rock in the region is the Jitou stock, which

137   occurs in the core of the Shenchong syncline as a multiphase intrusion with quartz diorite at the

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138   center and diorite porphyry along the margins. The quartz diorite has yielded Early Cretaceous

139   SHRIMP and LA-ICP-MS zircon U–Pb ages of 140.4 ± 2.2 Ma (Wang et al. 2004) and 139.6 ±

140   1.5 Ma (Zhang et al. 2017c), respectively.

141   There are two major mineralization types at Xinqiao, namely the economically significant

142   stratiform mineralization (No. I Orebody, accounting for 90% of the Cu, S and Fe reserves of

143   deposit) and the economically less important skarn mineralization (No. V Orebody). The

144   stratiform mineralization is confined to the unconformity between the Upper Devonian Wutong

145   Formation quartz sandstone and the Upper Carboniferous Huanglong Formation limestone,

146   whereas the skarn-type mineralization is hosted along the intrusive contact between the Jitou stock

147   and the Lower Permian Qixia Formation limestone (Fig. 2b).

148   The stratiform orebody is 2560 m long, 1810 m wide, averaging 21 m thick, striking NE and

149   dipping to the NW, parallel to the Upper Devonian Wutong and the Upper Carboniferous

150   Huanglong formations (Fig. 2b), with a quartz-pyrite stockwork in the Upper Devonian Wutong

151   Formation quartz sandstone acting as the footwall to the stratiform orebody. The stratiform ores

152   have replaced the limestones of the Middle and Upper Carboniferous Huanglong and Chuanshan

153   formations, with the Lower Permian Qixia Formation limestone and chert forming the hanging

154   wall (Fig. 2b). The ore minerals in the stratiform orebody include magnetite, chalcopyrite,

155   pyrrhotite, hematite, native gold and electrum (Dai and Liu 1984), whereas gangue minerals

156   include primarily garnet (two stages of garnet; Zhang et al. 2017b, 2018), diopside, epidote,

157   chlorite, quartz and calcite. The metallic mineral assemblages in the stratiform orebody show

158   proximal to distal zoning away from the Jitou stock: magnetite + pyrite → chalcopyrite + pyrite

159   → pyrite (Zhang et al. 2017c). The wall rock alteration includes garnet, sericite, quartz, chlorite

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160 and kaolinite, with silicic alteration mainly developed in the footwall of the orebody. The

161 paragenetic sequences of mineralization and alteration within Xinqiao was documented by Zhang

162 et al. (2017a, c). It can be divided into five stages, the early skarn (Stage I, garnet – diopside), late

163 skarn (Stage II, epidote – magnetite), metallic oxide (Stage III, hematite), colloform pyrite (Stage

164 IV) and quartz-sulfide (Stage V, quartz – chalcopyrite – pyrite).

165 Both the endoskarn and exoskarn ore-bodies are calcic skarn, and consist predominately of

166 garnet, wollastonite and subordinate pyroxene (Wang et al. 2011). Major metallic minerals include

167 massive / vein / disseminated magnetite, pyrite, chalcopyrite, pyrrhotite, sphalerite and galena.

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169 **SAMPLING AND ANALYTICAL METHODS**

170 **Sampling**

171 A total of ten magnetite-bearing samples (XQ11-3, XQ17-1, XQ17-3, XQ22-2, XQ22-7,

172 XQ33-11, XQ47-6, D006-5, D006-7 and D006-10) were collected from the Xinqiao stratiform

173 orebody. Samples XQ11-3, XQ17-1 and XQ17-3 were collected from the southwest part of the 60

174 m level in the Xinqiao open pit. Samples XQ22-2 and XQ22-7 were collected underground in the

175 E24 stope at depths of −270 m. Sample D006-5 were collected in the W16 stope underground at

176 depths of −270 m. Sample XQ47-6 was collected in the No. 6 Tunnel at a depth of −300 m, and

177 samples XQ33-11, D006-7 and D006-10 were collected in the W105 stope at a depth of −300 m.

178 Detailed sample locations are shown in Figure 2.

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180 **Analytical methods**

181 Prior to the EPMA and LA-ICP-MS in-situ major and trace elements analysis, BSE imaging

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182 of the magnetite samples was carried out using a TESCAN MIRA3 field emission scanning

183 electron microprobe (FE-SEM) at the Testing Center, Tuoyan Analytical Technology Co. Ltd.

184 (Guangzhou, China). Working conditions of the BSE imaging include 20 kV accelerating voltage

185 and 15nA beam current.

186   In-situ major element analysis using EPMA was carried out in the School of Geosciences and

187 Info-Physics of the Central South University, using a 1720 EPMA (Shimadzu Corporation, Japan).

188 Analytical parameters include 15kv (acc. voltage), 2.0×10^{-8} A (probe current) and 1μm (spot size),

189 with a detection limit of 0.01%. By using a spot size of 1μm it as possible to effectively avoid the

190 influence of micro-inclusions trapped in magnetite. Elements analyzed include TiO_2, SiO_2, FeO,

191 MnO, NiO, CoO, Cr_2O_3, CaO, V_2O_3, Al_2O_3 and MgO. The analyses were calibrated using natural

192 and synthetic mineral standards as follows: spinel for Mg and Al, diopside for Si and Ca, ilmenite

193 for Ti, chromite for Cr and Fe, manganese oxide for Mn, niccolite (NiAs) for Ni, cobalt metal for

194 Co, and vanadium metal for V.

195   Magnetite LA-ICP-MS trace element analysis was carried out at the Key Laboratory of

196 Marine Resources and Coastal Engineering, Sun Yat-sen University. The detailed method for

197 in-situ trace-element analysis for magnetite is given in Sun et al. (2017). The analysis was

198 performed using a pulsed 193 nm ArF Excimer laser (GeoLasPro), and ion-signal intensities were

199 acquired using an Agilent 7700x ICP-MS. A 32 μm spot was used with an energy density of 8

200 J/cm^2 and a repetition rate of 5 Hz. The trace element compositions of magnetite were calibrated

201 against the USGS synthetic basalt glass GSE-1G, using Fe determined by EPMA data as the

202 internal standard. Each analysis consisted of a 20 s background measurement (laser-off) followed

203 by 45 s of data acquisition. Data reduction was performed using the ICPMSDataCal software (Liu

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204   et al. 2010). Twenty-three elements were analyzed (Mg, Al, Si, Ca, Sc, Ti, V, Cr, Mn, Co, Ni, Cu,

205   Zn, Ga, Ge, As, Sn, W, Au, Pb, Bi, Th and U).

206

207   **RESULTS**

208   **Petrogenesis, internal texture and paragenesis of magnetite**

209   Based on mineral assemblages and textural relationships, six stages of magnetite were

210   identified in the Xinqiao stratiform orebody. Magnetite in the first four stages can be identified in

211   samples of XQ33-11, XQ47-6, D006-7, and D006-10. The fifth stage of magnetite (Mt5) is

212   corresponding to samples of XQ22-2, XQ22-7, and D006-5, whereas the sixth stage of magnetite

213   (Mt6) occurs in samples of XQ11-3. XQ17-1, and XQ17-3. These stages are described in greater

214   detail as follows:

215   The first stage of magnetite (Mt1) commonly occurs as anhedral grains ranging from 5 to 10

216   μm, and is dark gray under BSE (Fig. 3a and b). It is often replaced by porous anhedral second

217   stage magnetite (Mt2) which is light gray under BSE and 5 to 10μm in size (Fig. 3b). The

218   boundary between Mt1 and Mt2 is irregular and sharp (Fig. 3b), consistent with replacement

219   textures.

220   The third stage of magnetite (Mt3) commonly coexists with epidote as disseminated or

221   massive forms and replaces the early garnet (Fig. 3c). It occurs as subhedral (Fig. 3c) or anhedral

222   (Fig. 3d) grains ranging from 200 to 400 μm, and is commonly replaced by quartz, apatite, and

223   pyrite (Fig. 3d). In BSE imaging, Mt3 are characterized by well-developed oscillatory zoning and

224   core-rim textures (Fig. 3e–h). Importantly, Mt1 and Mt2 are only found in the cores of Mt3 grains

225   (Fig. 3e and f). The cores (Mt3c; ca. 40–100 μm in size) of Mt3 are typically darker gray than the

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226   rims (Mt3r; ca. 100–200 μm in size). Both Mt3c and Mt3r display chemical oscillatory zoning,

227   parallel to the contact interface between Mt3c and Mt3r (Fig. 3f and h), which is consistent with

228   growth zoning rather than replacement of multistage magnetite.

229   The local fourth stage of magnetite (Mt4) is porous and anhedral, and typically lighter gray

230   than Mt3r in BSE imaging (Fig. 3g and h). It dominantly occurs as irregular veins (ca. 2–5μm in

231   width) or patches (ca. 1–5μm in size) and truncates the oscillatory zoning of both Mt3c and Mt3r

232   (Fig. 3g and h), indicating it formed later than Mt3.

233   Mt5 occurs close to the Jitou stock with massive textures (Fig. 3i). It is commonly cut by

234   pyrite stockwork or replaced by disseminated pyrite (Fig. 3i). The magnetite primarily occurs as

235   anhedral grains between 50 and 300 μm and is characterized by a well-defined 120° triple junction

236   texture, an indication of textural equilibration and recrystallization of Mt1–Mt4 (Fig. 3j; Hu et al.

237   2015). The margins of the majority of magnetite grains are commonly oxidized into hematite (Fig.

238   3j). Mt5 shows homogeneous internal textures under BSE (Fig. 3k).

239   Mt6 commonly cuts the crystal (Fig. 3l) and colloform pyrite (Fig. 3m) as veins. The

240   colloform pyrite in the reaction front with this magnetite vein has commonly recrystallized to

241   crystal pyrite (Fig. 3m). Mt6 formed by replacement of hematite based on the metasomatic

242   pseudomorph textures, particularly the fine-grained foliaceous (Fig. 3m) or needle-like (Fig. 3n

243   and o) textures, generally 10–25μm across but up to ca. 60μm. Additionally, Mt6 is commonly

244   replaced by later pyrite that coexists with calcite (Fig. 3n). Siderite occurs as anhedral grains

245   among Mt6 grains in veins (Fig. 3m). Under BSE, Mt6 also shows homogeneous internal textures

246   (Fig. 3p). Importantly, Mt1 to Mt5 are commonly replaced by pyrite coexisting with quartz, which

247   is subsequently cut by Mt6 veins replaced by pyrite coexisting with calcite, indicating that Mt6 is

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248   the final magnetite stage at Xinqiao.

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250   **Major element geochemistry**

251   A total of 130 EPMA spot analyses were completed on the six magnetite samples, with 16, 15,

252   15, 17, 20, 37 and 10 spots on Mt1, Mt2, Mt3c, Mt3r, Mt4, Mt5 and Mt6, respectively. The major

253   element compositions of the six stages of magnetite at Xinqiao are listed in Appendix 1 and shown

254   on Figure 4a. The contents of FeO, CoO and Al_2O_3 were above detection limits in all samples. The

255   contents of TiO_2, SiO_2, Cr_2O_3, V_2O_3 and MgO in most analysis are above the detection limits,

256   whereas those of MnO, NiO and CaO were below detection limits. In general, Mt1 (86.25–90.02

257   wt. %; Avg. 88.37 wt. %) and Mt3c (87.80–89.28 wt. %; Avg. 88.49 wt. %) have lower FeO

258   contents than Mt2 (91.65–93.16 wt. %; Avg. 92.35 wt. %), Mt3r (90.17–91.85 wt. %; Avg. 90.91

259   wt. %), Mt4 (89.38–93.28 wt. %; Avg. 92.34 wt. %), Mt5 (90.12–93.20 wt. %; Avg. 91.97 wt. %)

260   and Mt6 (92.47–93.20 wt. %; Avg. 92.89 wt. %).

261

262   **Trace element geochemistry**

263   Mt1, Mt2 and Mt4 were too small for LA-ICP-MS analysis. Therefore, LA-ICP-MS spot

264   analysis were conducted on six magnetite samples, with 18, 29, 38 and 13 spots on Mt3c, Mt3r,

265   Mt5 and Mt6, respectively. All data was examined using time-resolved LA-ICP-MS signals, and

266   only flat and stable signals of trace elements were selected to calculate their contents. Trace

267   elements contents for the Xinqiao magnetite are listed in Appendix 2 and shown on Fig. 4b, with

268   the corresponding detection limits listed in Appendix 3. Most of the data ($>$ 60%) for Cu, As, Au

269   and Bi are below detection limits. In general, Mt3, Mt5 and Mt6 have higher Mg, Al, Si, Ca and

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270   Mn contents (generally higher than 100 ppm; Fig. 4) than the other stages of magnetite. Mt3 has

271   higher Al (Avg. 8131 ppm), Si (Avg. 8266 ppm), Ti (Avg. 484 ppm) contents than Mt5 (Al: Avg.

272   482 ppm, Si: Avg. 2658 ppm, Ti: Avg.18.827 ppm) and Mt6 (Al: Avg. 1447 ppm, Si: Avg. 2398

273   ppm, Ti: Avg.19.381 ppm), and Mt6 has the lowest contents of Al, Ti and Sn (Avg. 2.940 ppm),

274   and the highest contents of Sc (Avg. 8.780 ppm), Ni (Avg. 1.565 ppm), Ge (Avg. 10.632 ppm), As

275   (Avg. 51.331 ppm), W (Avg. 61.554 ppm), Pb (Avg. 4.781 ppm) and U (Avg. 1.790 ppm). Mt3c

276   have higher Mg, Al, Si, Ca, Sc, Ti, Cu and Ga, and lower Cr, Mn, Co, Ni, Zn, As, Pb and Th

277   contents than Mt3r. More importantly, Mt2 and Mt4 have lower trace-element concentrations (esp.,

278   Ti, Si, Ca and Mg; Fig. 4) than Mt1 and Mt3.

279

280   **DISCUSSION**

281   **Origin of the Xinqiao magnetite**

282   The trace-element geochemistry of magnetite can be used to investigate both the origin and

283   paragenesis of the mineralized system (Dupuis and Beaudoin 2011; Nadoll et al. 2012, 2014,

284   2015). Therefore, it is essential to document the origin of the six-stages of the Xinqiao magnetite

285   prior to discussing the ore-forming processes and the origin of the mineralization.

286   The FeO contents of Mt1, Mt3c, Mt2 and Mt4 are similar, but TiO_2, SiO_2, MnO, Cr_2O_3, CaO,

287   Al_2O_3 and MgO vary between Mt1 and Mt3c, and SiO_2, MnO and CaO between Mt2 and Mt4 (Fig.

288   4a) reveal that Mt1 is not the residue of Mt3c after being replaced by Mt4, and that porous Mt2 is

289   not equal to porous Mt4, supporting four distinct stages for the magnetite consistent with the

290   internal textural differences between Mt3c with chemical oscillatory zoning and homogenous Mt1.

291   Dare et al. (2014) proposed a Ti versus Ni/Cr discriminant diagram to distinguish magnetite from

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292   igneous and hydrothermal origins based on the different behavior of Ni and Cr in magmatic and

293   hydrothermal systems, but it is not very useful for hydrothermal magnetite formed in aqueous

294   fluids at high temperature (T) or magmatic magnetite in absence of coprecipitating Fe sulfides

295   (Knipping et al. 2015; Huang et al. 2019a). Some other diagrams, including Ti versus V (Nadoll et

296   al. 2015) and V/Ti versus Fe (Wen et al. 2017), have also been proposed to discriminate between

297   igneous and hydrothermal magnetite. At Xinqiao, most of the magnetite plots in the hydrothermal

298   magnetite zone in both Ti versus V (Fig. 5a) and V/Ti versus Fe (Fig. 5b) diagrams. Furthermore,

299   the Xinqiao magnetite contains low Ti (0.268–802 ppm) and V (0.483–625 ppm) (Appendices 1

300   and 2), distinct from igneous magnetite which is ubiquitously enriched in these lithophile elements

301   when compared to hydrothermal magnetite (Dare et al. 2014; Nadoll et al. 2015).

302   Mt1 and Mt3 have been extensively replaced by light grey Mt2 (Fig. 3b) and Mt4 (Fig. 3g

303   and h), respectively, preferentially along microfractures within or along grain margins. The

304   well-developed microporosity in Mt2 and Mt4, their sharp contacts with Mt1 and Mt3, and the

305   lower trace element contents (esp., Si, Ca, Mg, and Ti) from Mt1 to Mt2, and Mt3 to Mt4

306   (Appendix 1; Fig. 4a) suggest that they formed by coupled dissolution and reprecipitation (DRP)

307   of the precursor Mt1 and Mt3 grains (Putnis 2009; Putnis and John 2010; Dare et al. 2015; Hu et

308   al. 2015; Heidarian et al. 2016). The DRP is facilitated by a fluid-assisted process with dissolution

309   of early magnetite and precipitation of a secondary variety near the reaction front, with Si, Ca, Mg,

310   and Ti being removed from the early magnetite, and volume loss resulted in pervasive

311   micro-porosities, that were conducive for fluid infiltration that contributed to further dissolution of

312   magnetite (Hu et al. 2014, 2015).

313   Mt5 is characterized by 120° triple junction textures, which can form from either

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314 high-temperature annealing in a closed system or fluid-assisted replacement in an open system

315 (Nakamura and Watson 2001; Ciobanu and Cook 2004; Hu et al. 2015). Mt5 is characterised by

316 distinct trace-element compositions compared to the other magnetite stages (esp., Mg, Al, Ca, V,

317 Co, Ni and Cu; Fig. 4b), suggesting that Mt5 is most likely the result of fluid-assisted

318 recrystallization processes rather than high-T annealing, and consistent with a hydrothermal

319 origin.

320 Mt6 veins cut the euhedral and colloform pyrite which formed from magmatic hydrothermal

321 fluids (Zhang et al. 2017a; Li et al. 2018a), and the colloform pyrite in the reaction front with Mt6

322 has recrystallized to euhedral pyrite (Fig. 3m). This suggest a hydrothermal origin for Mt6 and

323 early hematite. The fact that hydrothermal magnetite and hematite formed later than the main

324 sulfide mineralization stage is not common in either skarn-type or SEDEX deposits, and may

325 reflect a distinct geological process. At Xinqiao, mineralization would have resulted in the

326 conversion of carbonate to skarn that would release abundant CO_2, and gradually increase the

327 pressure within the mineralization system (Meinert et al. 2005). When the fluid pressure exceeds

328 the lithostatic load, increasing shear stress creates extensive hydraulic fractures (Sibson et al.

329 1988), consistent with the magmatic hydrothermal origin of the broadly coeval quartz-pyrite

330 stockwork mineralization in the Wutong Formation quartz sandstone (Rb–Sr isotope isochron age

331 of quartz fluid inclusions from stockwork mineralization: 138.0 ± 2.3 Ma; zircon U–Pb dating of

332 Jitou stock: 139.6 ± 1.5 Ma; Zhang et al. 2017c) and initial $^{87}Sr/^{86}Sr$ values (quartz fluid inclusions

333 from stockwork mineralization: 0.71138 ± 0.00014, Zhang et al. 2017c; Jitou stock: 0.7065, Yu et

334 al. 1998) between stockwork mineralization and the Jitou stock. This fracturing caused a sudden

335 drop in fluid pressure, which would result in a short-lived FeS_2 supersaturation event, and promote

16

336   rapid growth of pyrite (Wu et al. 2018) probably causing the formation of colloform pyrite. It

337   could also have resulted in phase separation and partitioning of reduced H_2S into the vapor phase

338   to subsequently elevate the oxygen fugacity (fO_2) of the fluids (Ohmoto 1972; Drummond and

339   Ohmoto 1985). This is consistent with the negative $\delta^{34}S_{CDT}$ values in the Xinqiao colloform pyrite

340   (Zhang et al. 2017a) as fluid oxidation has been proposed to generate negative $\delta^{34}S$ values in

341   sulfides (Drummond and Ohmoto 1985; Hodkiewicz et al. 2009; LaFlamme et al. 2018). The

342   short-lived FeS_2 supersaturation and elevated fO_2 caused by the fracturing would create suitable

343   conditions for hematite formation. The fine-grained, foliaceous and needle-like textures of Mt6

344   suggest the early hematite replaced by Mt6 formed rapidly in unstable conditions, consistent with

345   the fracturing. Sealing of the fractures would allow fluid pressures to rebuild and the ore-forming

346   system to revert back to lower oxygen fugacities, which, coupled with continuous release of CO_2,

347   may explain the replacement of precursor hematite by Mt6 and the formation of siderite in the Mt6

348   veins. Therefore, Mt6 may have been genetically associated with fracturing.

349

350   **Controlling factors for magnetite precipitation**

351   Given that the trace-element composition of hydrothermal magnetite is determined by a

352   number of factors, including the physicochemical conditions (esp., T and fO_2), fluid composition

353   (intensity of fluid-rock interactions), and co-crystallizing minerals (e.g., Dare et al. 2012, 2014;

354   Nadoll et al. 2014; Li et al. 2018b; Huang et al. 2019b), the trace element geochemistry can be

355   used to constrain the precipitation conditions of the Xinqiao magnetite.

356   Temperature is thought to significantly influence the incorporation of trace elements into

357   magnetite (McIntire 1963; Nadoll et al. 2014; Knipping et al. 2015; Deditius et al. 2018). A

358   positive correlation between the formation temperature and Ti contents in magnetite has been

359   widely proven (Lindsley 1991; Rapp et al. 2010; Dare et al. 2012; Nadoll et al. 2012). At Xinqiao,

360   the decreasing TiO_2 from Mt3c (Avg. 0.11 wt.%) through Mt3r (Avg. 0.05 wt.%), Mt4 (Avg. 0.03

361   wt.%) and Mt5 (Avg. 0.01 wt.%) to Mt6 (below detection limit) (Appendix 1; Fig. 4a) is

362   consistent with gradual cooling from Mt3c to Mt6. However, the increasing Ti concentrations

363   from Mt1 (Avg. 0.05 wt.%) and Mt2 (Avg. 0.04 wt.%) to Mt3c suggest the temperature increased

364   during their formation. Plots of Al + Mn vs. Ti + V can also reflect the formation temperature of

365   magnetite (Nadoll et al. 2014; Deditius et al. 2018). The Xinqiao magnetite plots in Al + Mn vs. Ti

366   + V diagram (Fig. 5c) consistent with the gradually decreasing temperatures from Mt3c to Mt6.

367   Although the dataset for Mt1 is small, there is evidence for increasing temperatures from Mt1 to

368   Mt3 (Fig. 5c).

369   Vanadium concentrations in hydrothermal magnetite can record the fO_2 evolution of

370   hydrothermal fluids as vanadium is incompatible at high oxygen fugacities (Toplis and Corgne

371   2002; Dupuis and Beaudoin 2011; Nadoll et al. 2014; Knipping et al. 2015; Papike et al. 2005).

372   EPMA V_2O_3 data in different types of magnetite are consistent within error (Fig. 4a), but

373   LA-ICP-MS V data of Mt3, Mt5, and Mt6 decrease from Mt3c (Avg. 72.392 ppm) and Mt3r (Avg.

374   76.842 ppm) to Mt5 (Avg. 21.263 ppm), and increase from Mt5 to Mt6 (Avg. 72.000 ppm), which

375   suggests that fO_2 during the formation of Mt5 was higher than during the formation of Mt3 and

376   Mt6. This is consistent with variations in the Sn content that increase from Mt3c (Avg. 33.417

377   ppm) and Mt3r (Avg. 36.648 ppm) to Mt5 (Avg. 75.100 ppm), and then decrease from Mt5 to Mt6

378   (Avg. 2.940 ppm) because Sn is redox sensitive in magnetite (Carew 2004; Huang et al. 2019b).

379   The similar contents of V and Sn in Mt3c and Mt3r suggest that fO_2 conditions were stable during

18

380   their formation.

381   Extensive fluid-rock interactions commonly result in hydrothermal magnetite being enriched

382   in Mn, Mg, Al, and Si (Einaudi et al. 1981; Carew 2004; Meinert et al. 2005; Nadoll et al. 2014;

383   Deditius et al. 2018), but temperature can also influence the Mn content in magnetite (Nadoll et al.

384   2014; Deditius et al. 2018), so Liu et al. (2019) used Mg + Al + Si contents to investigate the

385   intensity of fluid-rock interactions. For the Xinqiao magnetite, Mt1 and Mt3c have the highest Mg

386   + Al + Si contents (Fig. 5d), suggesting that they have undergone the strongest fluid-rock

387   interactions during their formation. The lower Mg + Al + Si contents of Mt3c to Mt3r (Fig. 5d)

388   suggests a gradual decrease in the intensity of fluid-rock interactions. Both Mt5 and Mt6 have

389   lower Mg + Al + Si contents than Mt3c, suggesting they were affected by less intense fluid-rock

390   interactions, consistent with Mt5 having formed from fluid-assisted recrystallization processes and

391   Mt6 having grown rapidly under unstable conditions associated with fracturing. Moreover, Mt2

392   and Mt4 have the lowest Mg + Al + Si contents (Fig. 5d), consistent with them having formed by

393   DRP of pre-existing Mt1 and Mt3, respectively.

394   Dissolution and reprecipitation processes can be driven by variations in the properties of the

395   fluids, such as fluid compositions, T, pressure and fO_2 (Hemley and Hunt 1992; Putnis 2002;

396   Putnis and Putnis 2007). Mixing of saline fluids with Fe-rich magmatic-hydrothermal solutions, an

397   increase in T, and decreasing pressure and fO_2 are considered the most important causes for DRP

398   of primary magnetite (Hu et al. 2015) because these would enhance the solubility and

399   consequently the undersaturation of iron in the fluids (Hemley and Hunt 1992; Hu et al. 2014).

400   The lack of obvious increase in V_2O_3 and TiO_2 contents from Mt1 to Mt2 (V_2O_3: Avg. 0.14 wt.%

401   vs. 0.18 wt.%; TiO_2: Avg. 0.05 wt.% vs. 0.04 wt.%), and Mt3r to Mt4 (V_2O_3: Avg. 0.02 wt.% vs.

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402   0.03 wt.%; TiO_2: Avg. 0.05 wt.% vs. 0.03 wt.%), and the gradually increasing pressure before the

403   formation of the colloform pyrite associated with hydraulic fractures suggests variations in T, $f\mathrm{O}_2$

404   and pressure may not be the cause of DRP at Xinqiao. Previous multi-isotopic data (Re–Os, S and

405   Pb) of pyrite has shown that the Xinqiao stratiform mineralization was predominantly formed by

406   intensive water-rock interaction between the magmatic-hydrothermal fluids associated with the

407   Jitou stock and the Late Permian metalliferous black shales (Li et al. 2017, 2018a), which is

408   consistent with high Cl^- concentrations in fluid inclusions (up to 3.346 μg/g; Zhang 2015) at

409   Xinqiao. The saline fluids resulted from this water-rock interaction may be responsible for DRP of

410   the Xinqiao magnetite.

411

412   **Metallogenic implications**

413   Several diagrams, such as Ca + Al + Mn vs. Ti + V, Ni/(Cr + Mn) vs. Ti + V (Dupuis and

414   Beaudoin 2011) and Mn + Al vs. Ti + V (Nadoll et al. 2014) have been proposed to discriminate

415   various deposit types. Given that Ca and Ni concentrations in some EPMA analyses are below the

416   detection limits, the Mn + Al vs. Ti + V diagram (Fig. 6) was applied in this study, and Mt3, Mt4,

417   Mt5 and Mt6 dominantly plot in the skarn zones, suggesting that these stages of magnetite are

418   genetically associated with magmatic hydrothermal fluids, consistent with the hypothesis that the

419   Xinqiao stratiform mineralization is genetically associated with the Early Cretaceous

420   magmatic-hydrothermal fluids as a skarn deposit (e.g., Pan and Done 1999; Mao et al. 2009, 2011;

421   Zhang et al. 2017b, c, 2018; Li et al. 2017, 2018a, 2019). A large variation in Ti + V and relatively

422   stable Mn + Al in Figure 6 may be ascribed to the intensive fluid-rock interactions in the skarn

423   system. Mt1 plots in the BIF and IOCG zones close to skarn zone (Fig. 6), but a BIF source can be

20

424   excluded by the absence of Precambrian banded iron formations in the Tongling district, and the

425   IOCG mineralization is not supported by the low Ni contents of Mt1 and the absence of Na-Ca

426   alteration at Xinqiao (Huang et al. 2019a). The dark Mt1 under BSE has relatively high contents

427   of Si, Al, Ca, and Mg, which might be related to a major contribution of magmatic-derived fluids

428   due to a large fluid-rock ratio (Nadoll et al. 2012, 2014; Dare et al. 2014). Therefore, the

429   formation of Mt1 was influenced by both sedimentary and skarn-type processes, namely intensive

430   fluid-rock interactions, coinciding with the high Mg + Al + Ca of Mt1 (Fig. 5d), and the Xinqiao

431   magnetite may have been genetically associated with skarn mineralization derived from magmatic

432   hydrothermal fluids.

433   If the Xinqiao deposit formed from a single hydrothermal event, the ore-forming temperature

434   would gradually decrease over time. Although the temperatures decrease from Mt3 to Mt6, the

435   increase from Mt2 to Mt3c implies that the Xinqiao stratiform mineralization may reflect at least

436   two hydrothermal events. The combination of close spatial and temporal relationship between the

437   Jitou stock (zircon U–Pb age of 139.6 ± 1.5 Ma; Zhang et al. 2017c) and the stratiform orebody

438   (quartz fluid inclusions Rb–Sr isotope isochron age of 138.0 ± 2.3 Ma (Zhang et al. 2017c) and

439   Re–Os isotope isochron ages of 136.7 ± 4.6 Ma, 143 ± 16 Ma, and 135.5 ± 4.0 Ma for colloform

440   pyrite, pyrite from garnet-bearing skarn ore, and euhedral pyrite cemented by calcite, respectively

441   (Li et al. 2018a)), the iron (δ^{57}Fe: −1.22‰ to 0.15‰; Wang et al. 2011, 2013), S isotope

442   compositions (δ^{34}S$_{CDT}$ of −0.6 to 2.7‰, 1.8 to 2.5‰, and 1.9 to 4.4‰ for colloform pyrite,

443   fine-grained pyrite formed from recrystallization of colloform pyrite, and coarse-grained pyrite

444   coexisting with quartz and calcite, respectively; Zhang et al. 2017a) and Co/Ni ratios (0.67−2.94,

445   1.05−3.24, and 1.03−4.67 for colloform pyrite, fine-grained pyrite, and coarse-grained pyrite;

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446   Zhang et al. 2017a) of pyrite in the stratiform orebody, and the H–O isotope data (δD_{SMOW}=

447   −48.02−61.82‰, $\delta^{18}O_{SMOW}$ =12.56−22.74‰；Liu, 2002) of ore-bearing quartz in the stratiform

448   orebody suggests that the magmatic hydrothermal fluids associated with the Jitou stock are

449   dominantly responsible for the stratiform mineralization. One diorite porphyry dyke has been

450   found in the Xinqiao mining area, but it yielded a zircon U–Pb age of 130.6 ± 1.1 Ma (Li et al.

451   2017) and obviously cut through the stratiform orebody (Zhang 2015). Additionally, there is no

452   evidence to support other concealed intrusions in the region. Therefore, we conclude that the

453   increasing fluid temperatures from Mt2 to Mt3 likely resulted from episodic release of

454   hydrothermal fluids derived from magma associated with the Early Cretaceous Jitou stock, given

455   that the injection of multiple pulses of fluid is typical of magmatic-hydrothermal systems

456   (Hedenquist and Lowenstern 1994). Moreover, multiple influxes of fluid are also consistent with

457   the formation of multigenerational garnet at Xinqiao (Zhang et al. 2017b).

458   Here we propose a textural and trace-element geochemical evolution (Fig. 7) for the Xinqiao

459   magnetite. The first Si + Fe-rich magmatic-derived hydrothermal fluids released from the deep

460   magma chamber associated with the Jitou stock, entered the low-pressure unconformity between

461   the Upper Devonian Wutong Formation quartz sandstone and the Upper Carboniferous Huanglong

462   Formation limestone and interacted with limestone to form Mt1 enriched in Si, Al, Ca, and Mg.

463   Meanwhile, some fluids evolved into high-salinity fluids after intensive interaction with the Late

464   Permian metalliferous black shales, and mixed with the magmatic hydrothermal fluids to enhance

465   the solubility of iron in the fluids resulting in the formation of porous Mt2 by DRP which is

466   depleted in Ti, Si, Ca, V and Mg, that replaced Mt1. Subsequently, a second pulse of abundant Si

467   + Fe-rich magmatic hydrothermal fluid was released. Fluctuating fluid compositions and

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468   physicochemical conditions resulted in the formation of core-rim magnetite (Mt3) with oscillatory

469   zoning. Similarly, high-salinity fluids derived from interaction between magmatic hydrothermal

470   fluids and the metalliferous black shales induced the dissolution of Mt3 and precipitation of Mt4.

471   During this process, Ti, Si, Ca, Al, and Mg were removed from Mt3, and the FeO content

472   increased. During mineralization early magnetite (Mt1–Mt4) recrystallized to Mt5 under

473   conditions of decreasing temperature and fluid-rock interactions intensity and increasing fO_2.

474   Lastly, the gradual increase in ore-forming pressure due to the release of abundant CO_2 during the

475   conversion of carbonate to skarn would have exceeded the lithostatic load to create extensive

476   hydraulic fractures in the Wutong Formation quartz sandstone. This fracturing would have resulted

477   in elevated fO_2 fluids that formed hematite with fine-grained foliaceous and needle-like textures.

478   After sealing of the fractures, fluid pressures would increase and the ore-forming system revert

479   back to low oxygen fugacity, with Mt6 l replacing the precursor hematite.

480

IMPLICATIONS

482   Primary magnetite easily undergoes dissolution and reprecipitation, oxy-exsolution, and/or

483   recrystallization to reach chemical and textural reequilibration, and consequently texture

484   examination is necessary before conducting trace-element analyses. At Xinqiao, the microporosity

485   of Mt2 and Mt4 magnetite, and their sharp contacts with Mt1 and Mt3, suggest that they formed

486   via coupled dissolution and reprecipitation of the precursor Mt1 and Mt3 magnetite, respectively.

487   Based on the metasomatic pseudomorph textures, Mt6 formed by replacement of hematite as a

488   result of fracturing. Moreover, this study also demonstrates that in-situ trace elements analysis

489   based on detailed texture examination under robust geological and petrographic frameworks can

490   effectively constrain the mineralization processes and ore genesis. Increasing Ti content from Mt2

491   to Mt3 suggests the elevated temperature during their formation, which is ascribed into multiple

492   pulses of magmatic-hydrothermal fluids, and reveals that the Xinqiao stratiform mineralization

493   was genetically associated with multiple influxes of magmatic hydrothermal fluids derived from

494   the Early Cretaceous Jitou stock.

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705

FIGURE CAPTIONS

707 **FIGURE 1**. (a) Location of the Tongling ore district in the Middle-Lower Yangtze River Valley

708 metallogenic belt (after Mao et al. 2011). TLF: Tancheng–Lujiang fault; XGF: Xiangfan–Guangji

709 fault; YCF: Yangxing–Changzhou fault. (b) Geological map of the Tongling district (modified

710 from Chang et al. 1991).

711

712 **FIGURE 2.** (a) Geological map and (b) representative cross section of the Xinqiao Cu–Fe–Au

713 deposit (after Tang et al. 1998 and Zang et al. 2004, respectively).

714

715 **FIGURE 3.** Photographs showing representative mineral assemblages and textural features of

716 magnetite in the Xinqiao stratiform orebody. (a) Anhedral Mt1 grain replaced by porous Mt2

717 (BSE); (b) Irregular and abrupt contact boundary between Mt1 and Mt2, indicating replacement

718 texture (BSE); (c) Subhedral Mt3 grain coexisting with epidote occurs among garnet grains (PPL);

719 (d) Anhedral Mt3 grain coexisting with and replaced by quartz, apatite, and pyrite; (e) Mt1 and

720 Mt2 only trapped in the core of Mt3 (BSE); (f) The same chemical oscillatory zoning occurrence

721 of Mt3c and Mt3r, parallel to the contact interface between Mt3c and Mt3r (BSE); (g) Mt4 veins

722 and patches replacing Mt3c and truncating the chemical oscillatory zoning of Mt3c (BSE); (h)

723 Mt4 veins truncating the chemical oscillatory zoning of Mt3r (BSE); (i) Mt5 replaced by

724 stockwork and disseminated pyrite; (j) Mt5 grains with 120° triple junction texture and locally

725 replaced by hematite; (k) Homogeneous Mt5 replaced by pyrite and quartz (BSE); (l) Mt6 vein

726 cutting through crystal pyrite; (m) Mt6 coexisting siderite cuts through colloform pyrite as vein; (n)

34

727   Mt6 replaced hematite in the form of metasomatic pseudomorph textures; (o) Needle-like Mt6 is

728   replaced by pyrite coexisting with calcite; (p) Homogeneous internal texture of Mt6.

729   **Abbreviations**: Ep = epidote, Cal = calcite, Grt = garnet, Qtz = quartz, Ap = apatite, Hem =

730   hematite, Py = crystal pyrite, Py_c = colloform pyrite.

731

732   **FIGURE 4.** Box diagram of major (a) and trace (b) elements concentrations for the Xinqiao

733   magnetite.

734

735   **FIGURE 5.** Ti versus V (a), V/Ti versus Fe (b), (Al + Mn) vs. (Ti + V) (c; modified after Nadoll

736   et al. (2014) and Deditius et al. (2018)) and (Mg + Al + Si) vs. Ti (d) diagrams for the Xinqiao

737   magnetite.

738

739   **FIGURE 6.** Discrimination plot of Mn + Al vs. Ti + V for the Xinqiao magnetite mineralization

740   types (Nadoll et al. 2014).

741

742   **FIGURE 7.** Schematic diagram showing the textural and trace-element geochemical evolution for

743   the Xinqiao magnetite. See text for details.

744

745   **APPENDIX CAPTIONS**

746   **APPENDIX 1.** EPMA geochemical data (wt.%) of the Xinqiao magnetite. "b.d.l" means below

747   detection limit.

748

749   **APPENDIX 2.** LA-ICP-MS trace element data (ppm) of the Xinqiao magnetite. "b.d.l" means

750   below detection limit.

751

752   **APPENDIX 3.** LA-ICP-MS trace element detection limits (ppm) of the Xinqiao magnetite.
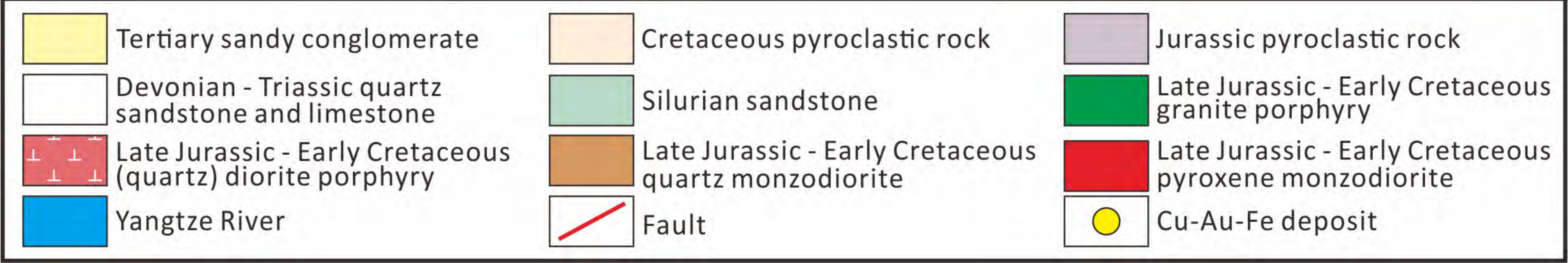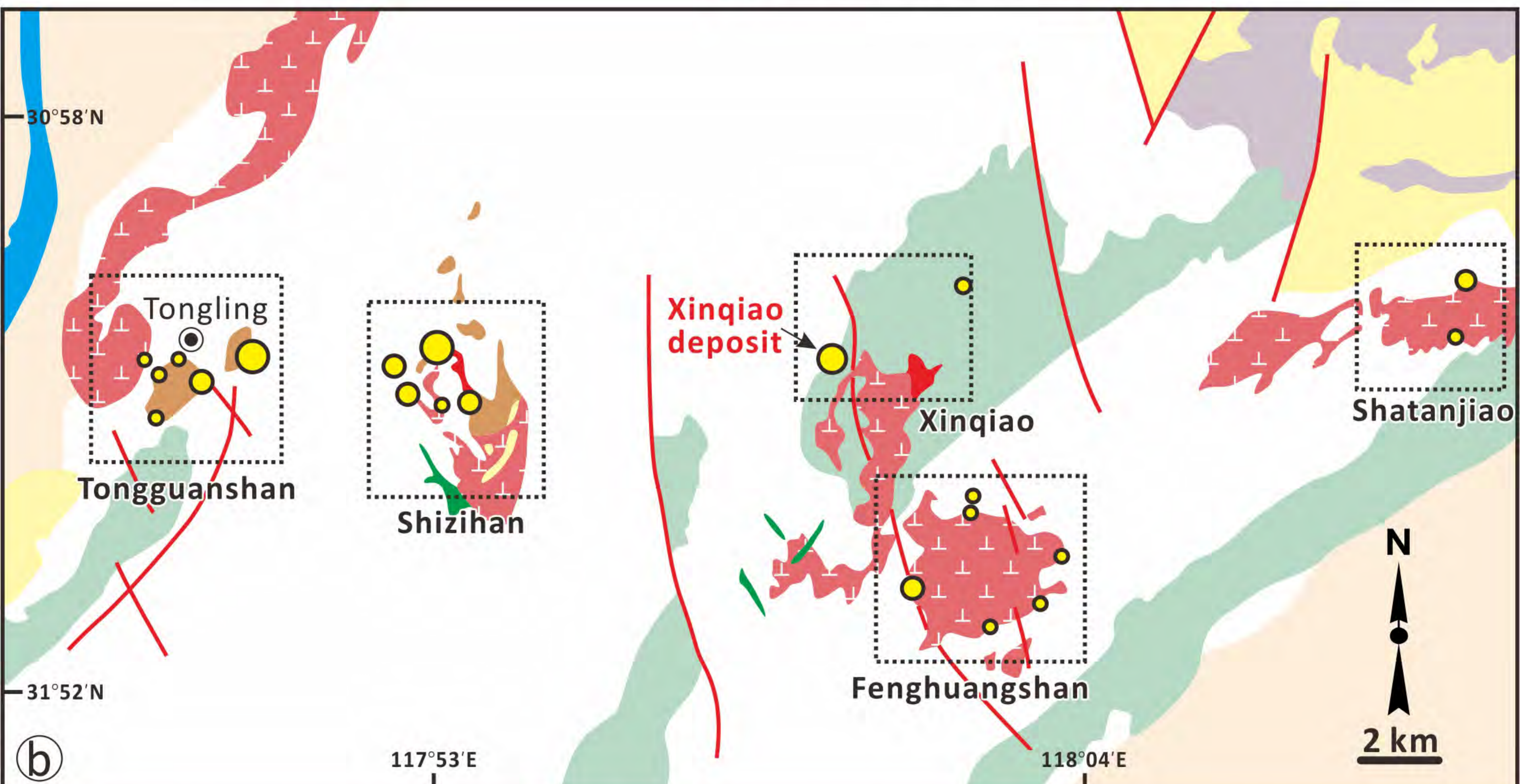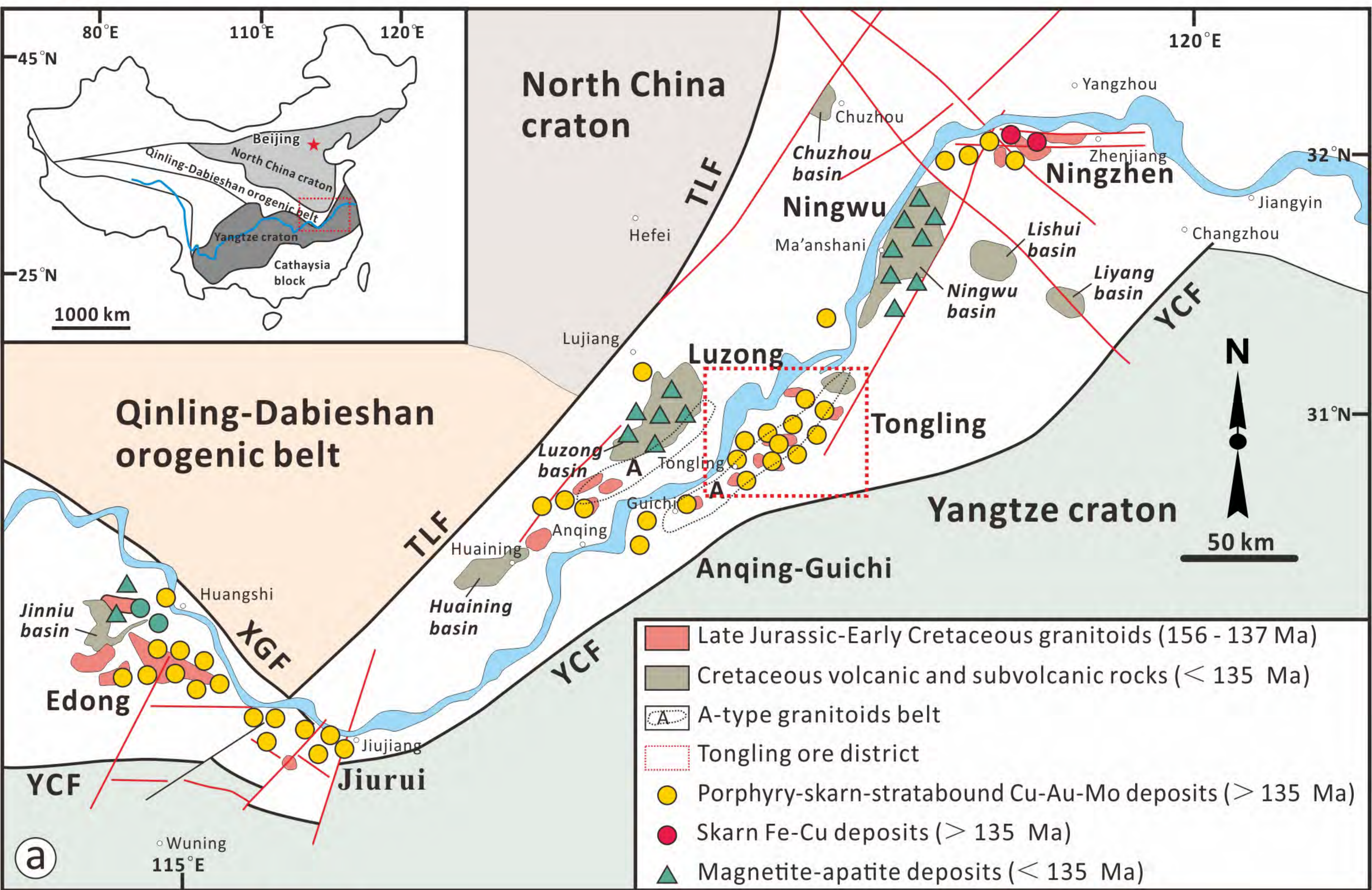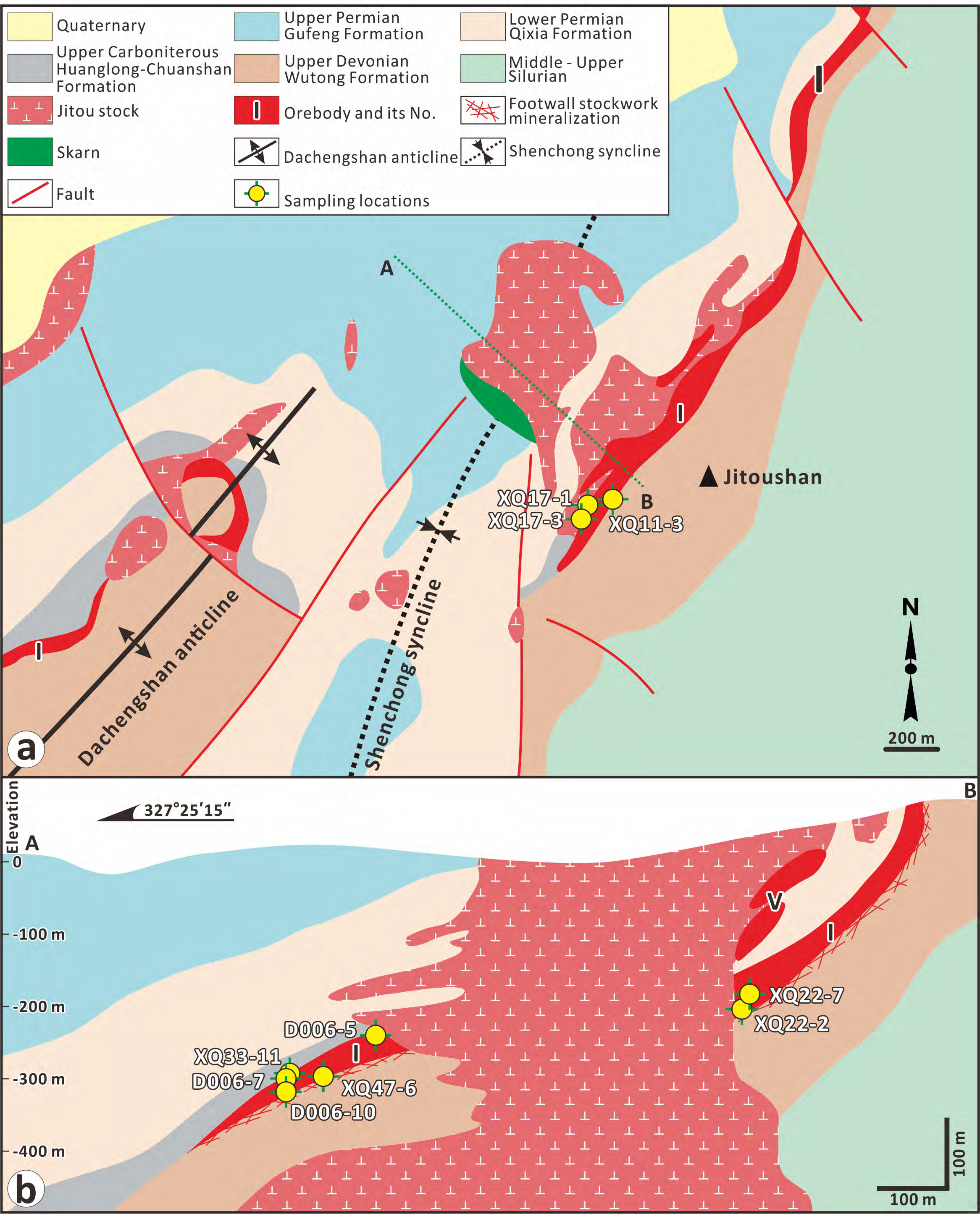
Fig. 1

(a)

Inset map labels: North China craton, Qinling-Dabieshan orogenic belt, Yangtze craton, Cathaysia block, Beijing, 1000 km

North China craton
Qinling-Dabieshan orogenic belt
Yangtze craton
Anqing-Guichi

TLF, XGF, YCF

Chuzhou, Chuzhou basin, Ningwu, Ma'anshani, Ningwu basin, Lishui basin, Liyang basin, Ningzhen, Zhenjiang, Yangzhou, Jiangyin, Changzhou
Hefei, Lujiang, Luzong, Luzong basin, Tongling, Guichi, Anqing, Huaining, Huaining basin
Huangshi, Jinniu basin, Edong, Jiurui, Jiujiang, Wuning

N, 50 km

Legend:
- Late Jurassic-Early Cretaceous granitoids (156 - 137 Ma)
- Cretaceous volcanic and subvolcanic rocks (< 135 Ma)
- A-type granitoids belt
- Tongling ore district
- Porphyry-skarn-stratabound Cu-Au-Mo deposits (> 135 Ma)
- Skarn Fe-Cu deposits (> 135 Ma)
- Magnetite-apatite deposits (< 135 Ma)

(b)

Tongling, Tongguanshan, Shizihan, Xinqiao deposit, Xinqiao, Fenghuangshan, Shatanjiao

N, 2 km

Legend:
- Tertiary sandy conglomerate
- Cretaceous pyroclastic rock
- Jurassic pyroclastic rock
- Devonian - Triassic quartz sandstone and limestone
- Silurian sandstone
- Late Jurassic - Early Cretaceous granite porphyry
- Late Jurassic - Early Cretaceous (quartz) diorite porphyry
- Late Jurassic - Early Cretaceous quartz monzodiorite
- Late Jurassic - Early Cretaceous pyroxene monzodiorite
- Yangtze River
- Fault
- Cu-Au-Fe deposit

Fig. 2

Legend:
- Quaternary
- Upper Permian Gufeng Formation
- Lower Permian Qixia Formation
- Upper Carboniterous Huanglong-Chuanshan Formation
- Upper Devonian Wutong Formation
- Middle - Upper Silurian
- Jitou stock
- Orebody and its No.
- Footwall stockwork mineralization
- Skarn
- Dachengshan anticline
- Shenchong syncline
- Fault
- Sampling locations

(a) Dachengshan anticline, Shenchong syncline, Jitoushan, XQ17-1, XQ17-3, XQ11-3, N, 200 m

(b) 327°25'15", Elevation, 0, -100 m, -200 m, -300 m, -400 m, D006-5, XQ33-11, D006-7, XQ47-6, D006-10, XQ22-7, XQ22-2, V, I, A, B, 100 m

Fig. 3

Fig. 4

Fig. 5

Fig. 6

Fig. 7

Mt1 (High Ti, Si, Ca, V, Al and Mg) → DRP → Mt2 (Low Ti, Si, Ca, V, and Mg; High Fe), Porosity, Mt1 → Second pulse → Mt3c (High Ti, Si, Ca, V, Al and Mg)

First magmatic hydrothermal pulse

Decreased fluid-rock interactions intensity and T; elevated fO_2

Mt3r (High Ti, Si, Ca, V, Al and Mg)

DRP → Porosity, Mt4 (Low Ti, Si, Ca, V, and Mg; High Fe)

Recrystallization → Mt5 (Low Ti, Si, Ca, V, and Al)

Fracturing → Mt6 (Low Ti, Si, Ca, V, and Al)