Genesis of the Fuxing porphyry Cu deposit in Eastern Tianshan, China: Evidence from fluid inclusions and C–H–O–S–Pb isotope systematics

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A B S T R A C T
The Fuxing porphyry Cu deposit is a recently discovered deposit in Eastern Tianshan, Xinjiang, northwestern China. The Cu mineralization is associated with the Fuxing plagiogranite porphyry and monzogranite, mainly presenting as various types of hydrothermal veins or veinlets in altered wall rocks, with potassic, chlorite, phyllic, and propyllic alteration developed. The ore-forming process can be divided into four stages: stage I barren quartz veins, stage II quartz–chalcopyrite–pyrite veins, stage III quartz–polymineralic sulfide veins and stage IV quartz–calcite veins. Four types of fluid inclusions (FIs) can be distinguished in the Fuxing deposit, including hypersaline (H-type), vapor-rich two-phase (V-type), liquid-rich two-phase (L-type), and trace amounts of pure vapor inclusions (P-type), but only the stage I quartz contains all types of FIs. The stages II and III quartz have two types of FIs, with exception of H- and P-types. In stage IV quartz minerals, only the L-type inclusions can be observed. The FIs in quartz of stages II, III and IV are mainly homogenized at temperatures of 357–518 °C, 234–322 °C and 145–240 °C, with salinities of 1.9–11.6 wt.% NaCl equiv., 1.6–9.6 wt.% NaCl equiv., 1.4–7.7 wt.% NaCl equiv., and 0.9–3.7 wt.% NaCl equiv., respectively. The ore-forming fluids of the Fuxing deposit are characterized by high temperature, moderate salinity and relatively oxidized condition. Carbon, hydrogen and oxygen isotopic compositions of quartz indicate that the ore-forming fluids were gradually evolved from magmatic to meteoric in origin. Sulfur and lead isotopes suggest that the ore-forming materials were derived from a deep-seated magma source. The Cu mineralization in the Fuxing deposit occurred at a depth of ~1 km, and the changes of oxygen fugacity, decomposition boiling, and local mixing with meteoric water were most likely critical for the formation of the Fuxing Cu deposit.

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1. Introduction

Being one of the largest accretionary orogens worldwide, the Central Asian Orogenic Belt (CAOB) extends for more than 5000 km from the European craton to the Pacific Ocean (Fig. 1A; Jahn et al., 2000; Windley et al., 2007), and is bounded by the Tarim–North China craton to the south and by the Siberian craton to the north (Pirajno, 2010; Xiao et al., 2010, 2013; Wilhem et al., 2012; Yang et al., 2014a,b). The CAOB is a composite collage composed of microcontinental blocks, island arcs, oceanic crustal remnants and continental marginal facies rocks (e.g., Kovalenko et al., 2004; Deng et al., 2014, 2015a,b; Goldfarb et al., 2014; Wang et al., 2015e, 2016a,b,c). It was produced as a consequence of multiple accretion and arc-continent collision events from the Early Neoproterozoic to the Late Paleozoic, driven by the successive closure of the paleo-Asian Ocean (Sengor et al., 1993; Zhou et al., 2008, 2015; Xie et al., 2014; Deng and Wang, 2015; Zhang et al., 2016b). The Eastern Tianshan orogenic belt located on the southern margin of the CAOB (Fig. 1A, B), is one of the China’s important Cu metallogenic belts (Charvet et al., 2007; Pirajno et al., 2011; Wang et al., 2014a, 2015a,b). The copper mineralization in the Eastern Tianshan belt was mainly deposited in the Late Paleozoic (Wang et al., 2015a,b; Xiao et al., 2015), which is relatively earlier than the common mineralization age of the Gangdese Cu belt in south Tibet (Hou et al., 2011; Li et al., 2011). The most important Cu deposits in Eastern Tianshan are dominated by porphyry system, and principally include the Tuwu (Mao et al., 2005; Wang et al., 2014a, 2015a; Xiao et al., 2015), Yandong (Shen et al., 2014a,b; Wang et al., 2015b), Fuxing (ZMGNC, 2015; Wang et al., 2016c), Linglong (Zhang et al., 2008), Chihu (Zhang et al., 2016a), Yuhai (Zang, 2014; Wang et al., 2016d) and Sanchakou (Qin et al., 2009; Wang and Zhang, 2016) porphyry Cu deposits, which are mainly hosted in the Late Paleozoic (mainly Carboniferous) volcanic and intrusive rocks.

Fuxing deposit is a recently discovered porphyry Cu deposit in Eastern Tianshan, Xinjiang, northwestern China (Fig. 1C). It is another breakthrough for the porphyry Cu prospecting, while its size was estimated in recent years by the Zijin Mining Group Northwest Co., Ltd. To date, only minimal basic geological research has been carried
out on this deposit (ZMGNC, 2015; Wang et al., 2016c), meaning that the evolution of ore-forming fluids and the source of the Fuxing deposit have not been conducted, especially the lack of any stable and radioactive isotope studies. In this contribution, we not only provide a detailed description of the geological features, but also present new results obtained from fluid inclusions and C–H–O–S–Pb isotopic data for the Fuxing deposit. We further discuss the characteristics and evolution of the ore-forming fluids, as well as migration and metallogenic mechanism. This information will provide new clues for fundamental research and exploration in Eastern Tianshan orogenic belt.

2. Regional geology

The Eastern Tianshan orogenic belt lies between the Junggar Basin to the north and the Tarim Basin to the south (Fig. 1B), and is a typical Palaeozoic island arc system characterized by a complicated tectonic history and diverse styles of mineral system (Xiao et al., 2004; Qin et al., 2011; Wang et al., 2015c; Zhang et al., 2015). It consists of the Bogeda–Haerlike Belt, the Jueluotage Belt and the Central Tianshan Terrane from the north to the south, which has different rock associations and hosts distinct mineralization (Fig. 1C; Pirajno et al., 2011; Huang et al., 2013). The Bogeda–Haerlike Belt comprises well-developed Ordovician to Carboniferous volcanic rocks, granites and mafic–ultramafic intrusions, which contains only few porphyry Cu and Au prospects (e.g., Chen et al., 2012; Gao et al., 2015). The Jueluotage Belt is characterized by Middle Palaeozoic volcanic and sedimentary rocks, intruded by voluminous Carboniferous–Jurassic felsic and mafic–ultramafic complexes (e.g., Zhou et al., 2010; Qin et al., 2011; Wang et al., 2015c, 2016b). The Central Tianshan Terrane is comprised of Precambrian basement, and hosts some hydrothermal magnetite deposits (Bgmrxuar, Bureau of Geology and Mineral Resources of Xinjiang Uygur Autonomous Region, 1993; Shen et al., 2014a,b).

The Jueluotage Belt can be subdivided into the Dananhu–Tousuquan arc belt in the north, the Kanggur–Huangshan ductile shear zone in the central, and the Aqishan–Yamansu arc belt in the south, which are separated by the Kanggur and Yamansu faults (Fig. 1C). The Dananhu–Tousuquan arc belt, situated north of the Kanggur Fault, mainly consist of Devonian volcanic and clastic sedimentary rocks of the Dananhu Formation, Carboniferous turbidites of the Gandun Formation, Carboniferous volcaniclastic rocks, basalt, tuff, limestone, sandstone and ophiolitic slice (Mao et al., 2005, 2015), hosting most porphyry Cu–Mo deposits (Han et al., 2014; Shen et al., 2014a,b). The Kanggur–Huangshan ductile shear belt is located between the Kanggur and Yamansu faults, an area in which most rocks have undergone lower greenschist facies metamorphism and strong ductile deformation, mainly including Devonian–Carboniferous volcaniclastic rocks, basalt, tuff, limestone, sandstone and ophiolitic slice (Mao et al., 2008, 2014; Xiao et al., 2013). There are a number of porphyry Mo deposits, orogenic Au deposits, and magmatic Cu–Ni sulfide deposits formed along the margin of the Kanggur–Huangshan ductile shear belt (Zhang et al., 2008; Wang et al., 2015c, d, 2016d; Wang and Zhang, 2016; Zhang et al., 2015). The Aqishan–Yamansu arc belt lies between the Yamansu and Aqikuduke faults, and is characterized by Early Carboniferous basalt, andesite, dacite, and tuff of the Yamansu Formation and Late Carboniferous rhyolite of...
the Tugutubulake Formation (Zhou et al., 2010; Xiao et al., 2013). The arc belt hosts numerous F-e(–Cu) and Cu–Ag–Pb–Zn skarn deposits (Chen et al., 2012; Hou et al., 2014).

The Eastern Tianshan orogen has undergone a complex tectonic evolution from the Late Paleozoic to the Mesozoic, involving subduction of the paleo-Tianshan Ocean, collision-accretionary, strike-slip motion, post-collisional, and intracontinental extension between the Tarim Basin and the Junggar Basin (Zhang et al., 2008; Pirajno et al., 2011; Xiao et al., 2013). The main structures of the Eastern Tianshan are characterized by a series of approximately EW-trending faults, including the regional-scale Dacaotan, Kanggur, Yamasu, and Aqikuduke faults, and some small-scale faults (Fig. 1C; Qin et al., 2011; Huang et al., 2013). Of these faults, the Kanggur Fault is the most important one, considered as a Late Carboniferous to Early Permian suture, along which important intrusive activity and associated ore mineralization occurred within the brittle-ductile shear zones (Mao et al., 2005, 2008; Wang et al., 2015c).

The intrusive rocks in Eastern Tianshan were mainly formed in the Late Paleozoic, represented by diorite porphyry, granodiorite, syenogranite, plagiogranite porphyry, quartz porphyry, and mafic–ultramafic bodies. They have a closely temporal and spatial relationship with most of the Cu (±Ni), Au, Ag, and Fe mineralization in Eastern Tianshan (Zhang et al., 2008; Pirajno et al., 2011; Wang et al., 2015a,b). Moreover, the Mesozoic intrusions have also been reported in recent years, represented by porphyritic granite, granite porphyry, biotite granite, and monzogranite, and are considered to be spatially and genetically related to porphyry Mo mineralization, as exemplified by Baishan and Donggebi deposits in Eastern Tianshan (Han et al., 2014; Wang et al., 2015d, 2016a; Zhang et al., 2015, 2016c).

3. Ore deposit geology

The Fuxing porphyry Cu deposit is in close proximity to the EW-trending Kanggur Fault, at geographic coordinates E92°18′57″ to E92°21′21″ and N42°04′41″ to N42°05′18″ (Fig. 1C). A series of NW- and NE-trending secondary faults developed in the deposit (Fig. 2A). Ore bodies are mainly hosted in the plagiogranite porphyry and monzogranite (Fig. 2B) and the surrounding volcanic rocks of the Early Carboniferous Qi’eshan Group (ZMGNC, 2015; Wang et al., 2016c). The Qi’eshan Group exposed in the Fuxing area consists of andesite and basalt (Fig. 3A, B) intercalated with tuff (Unit 1), andesitic or basaltic breccia (Unit 2), sandstone and siltstone with tuff and basalt interlayers (Unit 3), and gray to green conglomerate (Unit 4) from base upwards (Mao et al., 2005; Shen et al., 2014b; Gao et al., 2015). Andesite of Unit 1 contains zircons with a SHRIMP U–Pb age of 336.5 ± 6.6 Ma (Hou et al., 2005). The overlying rocks of the Jurassic Xishanyao Formation consist chiefly of sandstone, siltite, mudstone, and conglomerate, and they form an angular unconformity with the strata of the Qi’eshan Group (Fig. 2A).

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**Fig. 2.** Simplified geological map (A) and geological section (B) of the No.28 exploration line in the Fuxing porphyry Cu deposit (modified from ZMGNC, 2015).
The Carboniferous igneous rocks predominantly include diabase, diorite, dacite, plagiogranite porphyry, and monzogranite (Fig. 2A, B). Among them, the plagiogranite porphyry and monzogranite intrusions are closely related to the porphyry-style Cu mineralization and associated hydrothermal alterations (ZMGNC, 2015). They generally occurred as stocks ca. 1000 m deep below the surface (Fig. 2B) and the upper part appeared as dykes and apophyses. Our previous studies (data below from Wang et al., 2016c) reveal that the plagiogranite porphyry and monzogranite emplaced at 332.1 ± 2.2 Ma and 328.4 ± 3.4 Ma by SIMS zircon U–Pb dating, respectively. These intrusions show the same geochemical affinity with modern adakites, which are characterized by high SiO₂ contents, minor negative to positive Eu anomalies, strong depletion of heavy rare earth elements and Y, positive K, Rb, Sr, and Ba but negative Nb, Ta, Th, and Ti anomalies, and high (La/Yb)N ratios and Mg# values. In addition, these adakitic felsic intrusions display relatively high positive zircon ε(t) values (+11.4 to +18.3), low ⁸⁷Sr/⁸⁶Sr (0.706080–0.711239), high ¹⁴⁴Nd/¹⁴⁴Nd (0.512692–0.512922) ratios, and consistent zircon δ¹⁸O values (4.41–5.48‰). Taken together, it is suggested that the Fuxing parental magma was most likely derived from partial melting of the subducted oceanic crust followed by mantle peridotite interaction (Wang et al., 2016c).

The poorly exposed Cu orebodies in the Fuxing deposit have a tabular surface morphology and are elongated in the E–W direction (Fig. 2A). Using a cut-off grade of 0.5% Cu for oxidized ore and of 0.2% Cu for sulfide ore to outline the orebodies, the Fuxing main orebodies are about 555–2315 m long, 2–107 m thick, and dip toward the south at an angle of 61°–89°. In the dipping direction, the explored orebodies extend down-dip for more than 600 m (Fig. 2B), and the Cu mineralization occurs predominantly in plagiogranite porphyry and monzogranite (ZMGNC, 2015; Wang et al., 2016c). Sulfide mineralization in the Fuxing deposit occurs as veins (Fig. 4A, B), stockworks (Fig. 4C), veinlets (Fig. 4D), and disseminations (Fig. 4E, F). Field and petrographic observation indicate that mineral compositions are simple and chalcopyrite is the primary commodity in the whole deposit. The ore minerals include chalcopyrite and pyrite (Fig. 5A, B), with minor molybdenite (Fig. 4G and 5C), bornite, sphalerite (Fig. 5D), and chalcocite. Chalcopyrite generally occurs as fine-grained aggregates in quartz veins along open-space fissures, or is disseminated in the ore-related felsic intrusions and intermediate to mafic volcanic rocks. The gangue minerals are composed of quartz, K-feldspar, sericite, calcite, and chlorite (Fig. 5I), with minor biotite, muscovite, and epidote.

The Fuxing deposit shows zonation of alteration and mineralization, as is the case for the majority of porphyry deposits (Sillitoe, 2010; ZMGNC, 2015). Alteration zonation associated with Cu mineralization include silicification, chlorite–sericite, potassic, and phyllic alterations (Fig. 5E–G; ZMGNC, 2015). The mineralization processes in the Fuxing porphyry Cu deposit are divided into four stages from early to late (Figs. 6), based on mineralogy, micro-textures, crosscutting relationships and associated hydrothermal alteration types. Stage I ore-barren quartz veins are characterized by abundant quartz and minor K-feldspar and biotite, which are commonly cut by the later quartz–sulfide veins or veinlets (Fig. 4A). Pyrite occasionally appears as sparse dissemination in these quartz veins. This is an early ore stage and locally associated with potassic alteration. Stage II is represented by widespread quartz–chalcopyrite–pyrite veins and disseminated mineralization (Fig. 4B–F), as the products of the main ore stage, mainly containing quartz, chalcopyrite, and pyrite assemblages. This type of veins shows close association with chlorite–sericite alteration (Fig. 5E). Stage III quartz–polymetallic sulfide veins are characterized by an assemblage of quartz, pyrite, molybdenite, chalcopyrite, sphalerite (Fig. 5D), and galena, related to phyllic alteration and chlorite–sericite alteration (Fig. 5F, G). These veins account for less amount of chalcopyrite within the Fuxing deposit. Stage IV barren quartz ± calcite veins or veinlets are usually hosted in the propylitic alteration zone (Fig. 4H, I). They typically crosscut the earlier-formed veins or veinlets (Fig. 5H, I), indicating...
that they postdate the main ore stage. This last ore stage represents
either the waning of hydrothermal activity or a later supergene episode.
It is characterized by the deposition of calcite and alteration products
such as malachite (Fig. 3C, D), atacamite, and azurite. Together with
limonite and hematite, these supergene minerals generally occur near
the surface (averaging 40 m underground) in the Fuxing ore zone
(ZMGNC, 2015).

4. Sampling and analytical methods

Fluid inclusion microthermometric analyses were studied in quartz
from various kinds of veinlets or veins formed in different hydrothermal
stages (I, II, III, and IV). Polished thin sections were examined under
microscope to identify the phase, shape, size, distribution, and thus possi-
bile origin of the fluid inclusions. Fluid inclusion microthermometric
measurements and Laser Raman spectroscopic analyses were carried out at Resources Exploration Laboratory, China University of Geosciences,
Beijing. Fluid inclusion microthermometry data were obtained using a
Linkam MDS 600 heating-freezing stage equipped with a Zeiss micro-
scope. The stage enables measurements within the range of −196 to
600 °C. Freezing and heating runs were undertaken using liquid nitrogen
and a thermal resistor, respectively. The measurements consist of ice-
melting temperature (T_{m, ice}), halite dissolution temperature (T_{s, halite}),
and homogenization temperatures of fluid phases in FIs (T_{h, total}). Heating
rate was generally 1–5 °C/min during the initial stages of each heating
run and reduced to 0.3–1 °C/min close to the phase change points. Phases
that are present in individual fluid inclusion, including vapor, liquid and
daughter minerals, were carried out using Laser Raman spectroscopic
method with a Reinishaw in Via Raman scanning spectrometer, operat-
ing with an excitation wave length of λ = 514.5 nm. Salinities of aqueous
(NaCl–H_{2}O) inclusions, expressed as wt.% NaCl equiv., were calculated
using the final melting temperatures of ice points (Bodnar, 1993), while
the salinities of halite daughter mineral-bearing inclusions were calculat-
ed using the methodology outlined by Lecumberri-Sanchez et al. (2012),
which is based on the temperature of bubble dissolution and homogeni-
ization temperature. Densities of FIs in four stages were calculated using

Fig. 4. Photographs showing mineral assemblages and ore fabrics of the Fuxing porphyry Cu deposit. (A) Stage I barren quartz veinlet cut by stage II quartz–chalcopyrite vein. (B) Stage II quartz–pyrite–chalcopyrite vein cut by stage IV calcite–quartz vein. (C) Stage II quartz–pyrite–chalcopyrite stockwork. (D) Stage II quartz–pyrite–chalcopyrite veinlet associated with chlorite alteration. (E) Stage II pyrite and chalcopyrite assemblages occurred as disseminations within the wall rock. (F) Stage II quartz–pyrite–chalcopyrite veinlet, showing that medium-coarse grained pyrite occurred as disseminations within quartz. (G) Stage III quartz–pyrite–molybdenite veinlet. (H) Stage IV barren calcite–quartz vein cutting the earlier vein. (I) Stage IV barren calcite–quartz vein associated with propylitic alteration. Abbreviations: Qz—quartz; Cc—calcite; Chl—chlorite; Ep—epidote; Cp—chalcopyrite; Py—pyrite; Mo—molybdenite.
the Flincor procedure (Brown, 1989), and the details of the calculation method were given by Brown and Lamb (1989).

Nine quartz samples from various types of quartz–sulfide veins, and one quartz sample from the quartz–calcite vein were selected for H–O isotopic analyses. Hydrogen and oxygen isotopic compositions of the quartz were accomplished using a MAT-253EM mass spectrometer, at the Institute of Mineral Resources, Chinese Academy of Geological Sciences, Beijing. Oxygen was extracted from quartz by reaction with BrF5, and converted to CO2 on a platinum-coated carbon rod for oxygen isotope analysis (Clayton and Mayeda, 1963). Hydrogen were measured on water in fluid inclusions decrepitated from quartz and the water was released by heating the quartz to above 500 °C by means of an induction furnace, and then reacted with chromium powder at 800 °C to obtain hydrogen for isotopic analysis (Wan et al., 2005). The values of δ18Oquartz and δDH2O were reported relative to Vienna standard mean ocean water (V-SMOW); the analytic precision was better than ±0.2‰ for oxygen isotopes, respectively. Carbon and oxygen isotopic data were reported in per mil relative to the Peedee Belemnite limestone (PDB) standard. The value of δ18OSMOW is calculated by δ18OSMOW = 1.03086 × δ18OPDB + 30.86 (Friedman and O’Neil, 1977).

Thirteen sulfide samples from quartz–sulfide veinlets of different mineralization stages (I, II, and III) were chosen for sulfur isotopic analyses. Sulfur isotopic compositions were determined using a Finnigan MAT-251 mass spectrometer, at the Institute of Mineral Resources, Chinese Academy of Geological Sciences, Beijing. The δ34S value of sulfide were measured on SO2 obtained by placing the sulfide–CuO composite (at weight ratio of 1/7) into a vacuum system heated to 980 °C (Robinson and Kusakabe, 1975). Sulfur isotope ratios were reported as δ34S relative to the Canyon Diablo Troilite (CDT); the analytical reproducibility was ±0.2‰.

Six samples were chosen for lead isotopic analyses. Lead isotopic compositions were measured by an ISOPROBE-T Thermal Ionization Mass Spectrometer instrument (Zhou et al., 2013), at the Analytical Laboratory of the Beijing Research Institute of Uranium Geology. All Pb ratios were corrected according to the values of NBS SRM 981. The lead isotope analytic errors were reported as ±2σ, and the measurement accuracy was better than 0.005% for 208Pb/206Pb and 204Pb/206Pb values.
5. Fluid inclusion studies

5.1. Fluid inclusion petrography

Abundant FIs trapped in quartz-bearing veins of different stages were observed, and only primary inclusions were analyzed in this study, according to the textural criteria of Roedder (1984). Primary inclusions generally occurred as isolated or random group in intragranular quartz crystals (Roedder, 1984). On the basis of their compositions, phase proportion at room temperature (21 °C), and phase transitions during total homogenization, several types of FIs were recognized in the studied quartz samples from Fuxing, including trace amounts of pure vapor inclusions, rare hypersline inclusions, a small number of vapor-rich two phase inclusions, and predominant liquid-rich two phase inclusions. The fluid inclusion types distinguished in this study are presented in Table 1 and shown in Fig. 7.

Liquid-rich FIs (L-type) are the most common type in various kinds of quartz veinlets or veins (Fig. 7A, F). They consist of a liquid phase (LH2O) and a vapor bubble (VH2O) with the liquid phase volumetrically dominant, and generally occur in native-crystal, rounded or highly irregular shapes (Fig. 7A, H, I). They are relatively small in size from 3 to 13 μm, and usually have a vapor volume varying from 5 to 30 vol%.

Vapor-rich FIs (V-type) are detected in stage I to stage III quartz veins, including trace amounts of pure vapor inclusions, rare hypersline inclusions, a small number of vapor-rich two phase inclusions, and predominant liquid-rich two phase inclusions. The fluid inclusion types distinguished in this study are presented in Table 1 and shown in Fig. 7.

5.2. Microthermometry and Raman spectroscopy

The microthermometric data and calculated parameters for FIs are summarized in Table 1 and graphically illustrated in Figs. 8 and 9, which clearly show the linkage between mineral assemblages, physico-chemical conditions and metallogenic stages. Raman spectra of selected fluid inclusions are shown in Fig. 10.

FIs in stage I quartz veins are dominated by L-, V-, and H-types, with minor P-type. The L-type FIs homogenized to liquid at temperatures of 357–454 °C, with densities ranging from 0.58 to 0.70 g/cm³. Their ice-melting temperatures range from −7.5 to −4.9 °C, corresponding to salinities of 7.7–11.6 wt.% NaCl equiv. (Table 1). The V-type FIs homogenized to vapor at temperatures of 420–483 °C, with densities ranging from 0.33 to 0.46 g/cm³. Their ice-melting temperatures are between −2.2 and −1.1 °C, corresponding to salinities of 1.9–3.7 wt.% NaCl equiv. (Table 1). All the H-type FIs homogenized by halite dissolution at temperatures of 436 to 518 °C, in which vapor and liquid phases homogenized at temperatures of 309 to 386 °C. The corresponding calculated salinities range from 51.6 to 62.2 wt.% NaCl equiv. (Fig. 8A, B), and the densities of these FIs range from 0.82 to 1.08 g/cm³ (Table 1).

Stage II quartz veins are comprised of L- and V-type FIs, and the V-type FIs were present with L-type FIs in some quartz samples (Fig. 7D), implying that the system formed close to or on the two-phase line. The last ice-melting temperatures for the L-type FIs range from −6.3 to −3.2 °C, corresponding to salinities of 5.2 to 9.6 wt.% NaCl equiv. (Table 1). They homogenized to liquid at temperatures of 255–393 °C (Fig. 8C), with calculated densities ranging from 0.63 to 0.84 g/cm³. The last ice-melting temperatures for the V-type FIs are between −1.8 and −0.9 °C, with corresponding salinities of 1.6–3.1 wt.% NaCl equiv. (Fig. 8D). The total homogenization temperatures range from 319 to 383 °C, with densities of 0.56–0.68 g/cm³.

FIs in stage III quartz veins are dominant by L-type, with minor V-type. The last ice-melting temperatures for L- and V-type FIs range from −4.9 to −0.8 °C, with corresponding salinities of 1.4–7.7 wt.% NaCl equiv. (Fig. 8F). On heating, inclusions in quartz showed homogenization temperatures ranging from 234 to 322 °C, with peaks in the range of 260 to 300 °C (Fig. 8E). The densities of these FIs are calculated to be 0.69 to 0.85 g/cm³ (Table 1). The coexistence of V-type and L-type FIs (Fig. 7C) indicates that the system formed close to or on the two-phase line, indicating that boiling continued during the ore stage III.
In stage IV, the quartz veins only contain L-type FIs that homogenized to the aqueous liquid phase at temperatures of 145 to 240 °C (Fig. 8G), with calculated densities ranging from 0.84 to 0.94 g/cm³. The last ice-melting temperatures vary from −2.2 to −0.5 °C, equivalent to salinities of 0.9 to 3.7 wt.% NaCl equiv. (Fig. 8H). Clear evidence of boiling was absent in this ore stage.

5.3. Pressure estimation and mineralization depth

Based on the total homogenization temperatures and salinities of the L-, V-, and H-types inclusions, trapping pressures of FIs can be estimated in the binary system NaCl–H₂O as an approximation (Hedenquist et al., 1998; Driesner and Heinrich, 2002, 2007). Estimates of trapping pressure can be obtained only when the actual fluid entrapment temperature is known, or if fluid boiling or immiscibility was occurring in the system at the time of entrapment (Roedder and Bodnar, 1980; Roedder, 1984). Pressures determined for nonboiling assemblages are derived from the homogenization temperature and represent minimum values (Rusk et al., 2008). At Fuxing, the fluid boiling assemblages were found in the barren quartz veins (stage I), quartz-chalcopyrite-pyrite veins (stage II), and quartz-polymetallic sulfide veins (stage III), which could be supported by the intimate spatial relationship between V-type and L-type inclusions (Fig. 7C, D, G), and the analogous homogenization temperatures and contrasting salinities of these inclusions (Fig. 9A–F). Thus, homogenization temperatures are interpreted to closely approximate trapping temperatures, and trapping pressures can be estimated using the isobar equations from Driesner and Heinrich (2007). However, the lack of evidence for fluid boiling in the stage IV show that pressures determined only represent the minimum value (Rusk et al., 2008).

As shown in the plot of homogenization temperature versus salinity of L-, V-, and H-types FIs (Fig. 11; Driesner and Heinrich, 2007), the trapping pressures of the FIs in the early ore stage (I) and the main ore stage (II and III) were estimated to be 200–500 bars and 30–300 bars, respectively, given their peak homogenization temperature intervals of 420–460 °C and 260–380 °C, respectively (Fig. 8). However, it is not straightforward because the pressure of fluid system frequently alternates between lithostatic and hydrostatic (Rusk et al., 2008; Sillitoe, 2010). Based on the pressure estimated for the Fuxing deposit, the early and main stages of mineralization most likely occurred at depth.
of ~1 km, if lithostatic pressures were assumed in the high-temperature early ore stage, and hydrostatic pressures were assumed in the lower-temperature main ore stage. This estimation mineralization depth is consistent with those of many magmatic-hydrothermal systems in the world (Pirajno, 2009; Chen and Li, 2009; Li et al., 2012; Yang et al., 2012; Chen and Santosh, 2014).

Fig. 8. Histograms of homogenization temperatures and salinities of fluid inclusions in different stages for the Fuxing porphyry Cu deposit.

Fig. 9. Summary plot of homogenization temperatures and salinities of individual fluid inclusions in different stages for the Fuxing porphyry Cu deposit.

Fig. 10. Representative Laser Raman analyses of aqueous inclusions from the Fuxing porphyry Cu deposit.
The oxygen and hydrogen isotopic data are listed in Table 2 and plotted in Fig. 13. One earlier calcite sample from stage III give the $\delta^{13}C$ value of $-3.4\%$, and the $\delta^{18}O$ value of $+8.6\%$. The $\delta^{13}C$ and $\delta^{18}O$ values of six calcite samples from stage IV fall within a range of $-4.1$ to $-2.5\%$ and $-0.8$ to $+5.4\%$, respectively.

The sulfur isotopic data are shown in Table 4 and plotted in Fig. 14. The $\delta^{34}S$ values of eleven pyrite samples from the Fuxing deposit range from $-1.9$ to $+0.8\%$, with an average of $-0.7\%$, and the $\delta^{34}S$ value of one chalcopyrite sample is $-1.3\%$, and the $\delta^{34}S$ value of one molybdenite sample is $2.3\%$. Therefore, sulfur isotopic compositions in the same sample.

**6. C–H–O–S–Pb isotope systematics**

The oxygen and hydrogen isotopic data are listed in Table 2 and plotted in Fig. 12. The measured $\delta^{18}O$ values of ten quartz samples in different stages range from $+8.0$ to $+8.8\%$. The $\delta^{18}O$ values of hydrothermal fluids were calculated using the equation of Clayton et al. (1972), 1000na2o-water = $3.38 \times 10^6 \times T^{-2}$ – 3.40, together with the measured $\delta^{18}O$quartz values and the corresponding average homogenization temperatures of the Fs in the same stage of the same sample (Table 1). As a result, the $\delta^{18}O$water values from stages I to IV are $+4.9\%$, $+1.1$ to $+4.0\%$, $+0.4$ to $+0.7\%$, and $-4.1\%$, respectively (Fig. 12). All minerals selected for $\delta^{18}O$ analysis were also analyzed for their hydrogen isotopic composition. The $\deltaD$water values of the quartz samples in mineralizing stages I, II, III, and IV are $-63$ to $-56\%$, $-79$ to $-72\%$, $-80$ to $-77\%$, and $-86\%$, respectively (Fig. 11).

The carbon and oxygen isotopic data are listed in Table 3 and plotted in Fig. 13. One earlier calcite sample from stage III give the $\delta^{13}C$ value of $-3.4\%$ and the $\delta^{18}O$ value of $+8.6\%$. The $\delta^{13}C$ and $\delta^{18}O$ values of six calcite samples from stage IV fall within a range of $-4.1$ to $-2.5\%$ and $-0.8$ to $+5.4\%$, respectively.

The sulfur isotopic data are shown in Table 4 and plotted in Fig. 14. The $\delta^{34}S$ values of eleven pyrite samples from the Fuxing deposit range from $-1.9$ to $+0.8\%$, with an average of $-0.7\%$, and the $\delta^{34}S$ value of one chalcopyrite sample is $-1.3\%$, and the $\delta^{34}S$ value of one molybdenite sample is $2.3\%$. Therefore, sulfur isotopic compositions in the same sample.

**Table 2 Oxygen and hydrogen isotopic data of quartz from the Fuxing porphyry Cu deposit.**

<table>
<thead>
<tr>
<th>Sample no.</th>
<th>Mineral</th>
<th>Stage</th>
<th>T_h (°C)</th>
<th>$\delta^{18}O_{quartz}$ (‰)</th>
<th>$\delta^{18}O_{water}$ (‰)</th>
<th>$\deltaD_{water}$ (‰)</th>
</tr>
</thead>
<tbody>
<tr>
<td>FX-2804-2</td>
<td>Quartz</td>
<td>IV</td>
<td>442.8</td>
<td>4.9</td>
<td>56</td>
<td></td>
</tr>
<tr>
<td>FX-2804-2</td>
<td>Quartz</td>
<td>IV</td>
<td>430.8</td>
<td>4.9</td>
<td>63</td>
<td></td>
</tr>
<tr>
<td>FX-2804-4</td>
<td>Quartz</td>
<td>IV</td>
<td>395.6</td>
<td>4.0</td>
<td>72</td>
<td></td>
</tr>
<tr>
<td>FX-2804-5</td>
<td>Quartz</td>
<td>IV</td>
<td>353.8</td>
<td>3.0</td>
<td>73</td>
<td></td>
</tr>
<tr>
<td>FX-2804-6</td>
<td>Quartz</td>
<td>IV</td>
<td>336.5</td>
<td>2.8</td>
<td>72</td>
<td></td>
</tr>
<tr>
<td>FX-2804-2</td>
<td>Quartz</td>
<td>IV</td>
<td>324.8</td>
<td>1.9</td>
<td>78</td>
<td></td>
</tr>
<tr>
<td>FX-4201-2</td>
<td>Quartz</td>
<td>II</td>
<td>295.6</td>
<td>1.1</td>
<td>79</td>
<td></td>
</tr>
<tr>
<td>FX-4201-2</td>
<td>Quartz</td>
<td>II</td>
<td>273.5</td>
<td>0.4</td>
<td>77</td>
<td></td>
</tr>
<tr>
<td>FX-4201-2</td>
<td>Quartz</td>
<td>II</td>
<td>273.6</td>
<td>0.7</td>
<td>80</td>
<td></td>
</tr>
<tr>
<td>FX-4201-2</td>
<td>Quartz</td>
<td>II</td>
<td>190.3</td>
<td>-4.1</td>
<td>86</td>
<td></td>
</tr>
</tbody>
</table>

**Table 3 Carbon and oxygen isotopic data of calcite from the Fuxing porphyry Cu deposit.**

<table>
<thead>
<tr>
<th>Sample no.</th>
<th>Mineral</th>
<th>Stage</th>
<th>$\delta^{13}C_{PDB}$ (‰)</th>
<th>$\delta^{18}O_{PDB}$ (‰)</th>
<th>$\delta^{18}O_{SMOW}$ (‰)</th>
</tr>
</thead>
<tbody>
<tr>
<td>15FX-88</td>
<td>Calcite</td>
<td>III</td>
<td>-3.4</td>
<td>-21.6</td>
<td>8.6</td>
</tr>
<tr>
<td>15FX-65</td>
<td>Calcite</td>
<td>IV</td>
<td>-4.0</td>
<td>-25.1</td>
<td>5.1</td>
</tr>
<tr>
<td>15FX-38</td>
<td>Calcite</td>
<td>IV</td>
<td>-2.8</td>
<td>-30.8</td>
<td>-0.8</td>
</tr>
<tr>
<td>15FX-30</td>
<td>Calcite</td>
<td>IV</td>
<td>-3.7</td>
<td>-29.1</td>
<td>0.9</td>
</tr>
<tr>
<td>15FX-42</td>
<td>Calcite</td>
<td>IV</td>
<td>-2.5</td>
<td>-24.8</td>
<td>5.4</td>
</tr>
<tr>
<td>15FX-02</td>
<td>Calcite</td>
<td>IV</td>
<td>-4.1</td>
<td>-25.5</td>
<td>4.6</td>
</tr>
<tr>
<td>15FX-64</td>
<td>Calcite</td>
<td>IV</td>
<td>-4.0</td>
<td>-24.8</td>
<td>5.4</td>
</tr>
</tbody>
</table>

Note: The data were reported in permil relative to the Pee Dee Belemnite limestone (PDB) standard with total uncertainties were estimated to be better than 0.2‰ for $\delta^{18}O$ and 0.1‰ for $\delta^{13}C$. $\delta^{18}O_{SMOW}$ Values are calculated according to the equation ($\delta^{18}O_{SMOW} = 1.03086 \times \delta^{18}O_{PDB} + 30.86$) provided by Friedman and O'Neil (1977).

**Fig. 12. δD–δ18O diagram of the ore-forming fluids in the Fuxing porphyry Cu deposit.**

Fuxing deposit show a relatively narrow range (−1.9 to +2.3‰), indicating a homogenous sulfur source.

The lead isotopic data of intrusive rocks are shown in Table 5 and Fig. 15. One monzogranite sample at Fuxing has 206Pb/204Pb, 207Pb/204Pb, and 208Pb/204Pb ratios of 18.658, 15.476, and 37.9, respectively. Five plagiogranite porphyry samples have 206Pb/204Pb ratios ranging from 18.367 to 19.481, 207Pb/204Pb ratios ranging from 15.474 to 15.628, and 208Pb/204Pb ratios ranging from 37.6 to 39.3, respectively. As shown in Fig. 15, all of these data at Fuxing show analogous lead isotope compositions.

**7. Discussion**

**7.1. Origin of the ore-forming fluids and magmatic water**

The δ18Owater and δDwater values of the fluids from the early barren quartz stage at Fuxing plot near the magmatic water field (Fig. 12), indicating that magmatic fluids may have played a significant role during the formation of the initial ore-forming fluids. The samples from the quartz–chalcopyrite–pyrite veins in stage II and the quartz–polymetallic sulfide veins in stage III show a relatively narrow range of δ18Owater and δDwater values (Table 2) and plot into the region between the primary magmatic water box and the meteoric water line, but still dominated by magmatic water. One sample from the late quartz + calcite stage yielded remarkably lower values, which plot in the right of the meteoric water line (Fig. 12), indicating that some meteoric water were probably incorporated in the ore-forming system in stage.

**Fig. 11. Pressure estimation for fluid inclusions of the Fuxing porphyry Cu deposit.**

Note: This is the average homogenization temperature of the Fs in the same stage of the same sample. δ18Owater values are calculated according to the quartz–water equilibrium temperature formula provided by Clayton et al. (1972).
and Junggar orogenic belt (Fig. 14B), such as Tuwu (−3.0 to +1.7‰; Wang et al., 2015a), Yuhai (−1.3 to +0.3‰; Zang, 2014), and Baogutu (−5.1 to +0.7‰; Zhang et al., 2010). Furthermore, these data are also comparable to those of typical porphyry Cu–Mo deposits elsewhere (Fig. 14B), such as Butte Cu–Mo deposit in USA (−0.1 to +4.0‰; Field et al., 2005), Erdenet-Ubin Cu deposit in Mongolia (−1.8 to +1.5‰; Sotnikov et al., 2004), and Mariquita and Lucky Cu–Mo deposits in Mexico (−0.1 to +3.8‰ and −4.7 to +3.3‰; Salas et al., 2013), which are interpreted to be magmatic in origin. Therefore, the Fuxing Cu deposit has a solely magmatic source of sulfur.

Lead isotopic data point of the intrusive rocks from the Fuxing porphyry Cu deposit are distributed between the oreogenic belt and the mantle evolution line on the 

\[ \delta^{206}Pb/204Pb \] versus \[ \delta^{207}Pb/204Pb \] diagram (Fig. 15A), but they are near the upper crust growth curve on the \[ \delta^{206}Pb/204Pb \] versus \[ \delta^{208}Pb/204Pb \] diagram (Fig. 15B; Zartman and Doe, 1981). Moreover, the lead isotopic compositions of the monzogranite and plagiogranite porphyry at Fuxing are similar to those of sulfides and plagioclase porphyry in the Tuwu–Yandong porphyry Cu deposits (Zhang et al., 2006), indicating a relatively uniform lead source. Therefore, we suggest that the lead was most likely derived from a deep-seated magma that involved some crustal materials.

7.2. Nature and evolution of fluid system

The halite-bearing FIs have been commonly identified in hydrothermal minerals of many porphyry deposits world-wide (Lu et al., 2004; Chen et al., 2007; Pirajno, 2009; Lecumberri-Sanchez et al., 2012; Wang et al., 2014b; Zarasvandi et al., 2015), such as Tuwu and Yandong Cu deposits, NW China (Li et al., 2009), the Bajo de la Alumbrera Cu–Au deposit, Argentina (Ulrich et al., 2002), and the Dalli porphyry Cu–Au deposit, Central Iran (Zarasvandi et al., 2015). The halite-bearing FIs display three different types based on their homogenization behaviours during heating (Becker et al., 2008): (1) saturated saline inclusions in which vapor-liquid homogenization simultaneous with halite dissolution; (2) unsaturated saline inclusions in which vapor-liquid homogenization temperatures are higher than the melting temperatures of daughter crystals; and (3) supersaturated saline inclusions in which daughter crystals (halite) dissolution healed by vapor-liquid homogenization (Mao and Li, 2003; Yang et al., 2015). Halite-bearing inclusions were observed in the earlier hydrothermal quartz from the Fuxing porphyry Cu deposit. They were all homogenized by halite dissolution, with a wide homogenization temperatures range (436 to 518 °C), indicating that the hydrothermal fluids were halite-supersaturated, most likely resulted from fluid boiling or post-entrapment modification (Masterman et al., 2005; Yang et al., 2015; Lecumberri-Sanchez et al., 2015). At Fuxing, the fluid boiling can be supported by the coexistence of H-type, V-type and L-type FIs that are homogenized at similar temperatures and contrasting salinities (Fig. 9A–F).

Because the oversaturated saline inclusions represent cavities in which the daughter minerals and solution were captured at the same time, the melting temperatures of the daughter minerals do not reflect the salinities of the original solutions (Mao and Li, 2003). Therefore, we chose the other types FIs in stage I quartz veins to calculate the salinities, which are ranging from 1.9 to 11.6 wt.% NaCl equiv., indicating initially moderate salinities. Furthermore, the peak homogenization temperature interval of FIs in the early quartz veins ranges from 420 to 460 °C (Table 1), showing the characteristic of a high temperature fluid. In addition, the vein mineral association, such as the common occurrence of hematite and magnetite, suggest that the initial ore-forming fluids were relatively oxidizing at Fuxing. This is comparable with the studies of the Tongkuangyu porphyry Cu deposit, China (Jiang et al., 2014), the Bajo de la Alumbrera, Argentina (Ulrich et al., 2002), and the Butte porphyry Cu–Mo deposit, USA (Rusk et al., 2008), which are interpreted to have formed from relatively oxidized ore fluids. Recently, abundant CO2-bearing FIs have been reported in porphyry deposits formed in continental settings, such as the Wungetushan.

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**Table 4**

Sulfur isotopic data of sulfides from the Fuxing porphyry Cu deposit.

<table>
<thead>
<tr>
<th>Sample no.</th>
<th>Mineral</th>
<th>Stage</th>
<th>Sample description</th>
<th>( \delta^{34}S_{SMOW} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>FX-4201-14</td>
<td>Pyrite</td>
<td>I</td>
<td>Qz–Py vein</td>
<td>0.1</td>
</tr>
<tr>
<td>15FX-99</td>
<td>Pyrite</td>
<td>I</td>
<td>Qz–Py vein</td>
<td>0.3</td>
</tr>
<tr>
<td>15FX-43</td>
<td>Pyrite</td>
<td>I</td>
<td>Qz–Py vein</td>
<td>0.1</td>
</tr>
<tr>
<td>15FX-42</td>
<td>Pyrite</td>
<td>I</td>
<td>Qz–Py vein</td>
<td>0.8</td>
</tr>
<tr>
<td>FX-2604-3-6</td>
<td>Chalcopyrite</td>
<td>II</td>
<td>Qz–Cp–Py vein</td>
<td>−1.3</td>
</tr>
<tr>
<td>FX-2804-18</td>
<td>Pyrite</td>
<td>II</td>
<td>Qz–Cp–Py vein</td>
<td>−0.6</td>
</tr>
<tr>
<td>FX-2804-24</td>
<td>Pyrite</td>
<td>II</td>
<td>Qz–Cp–Py vein</td>
<td>−0.7</td>
</tr>
<tr>
<td>FX-2604-3-6</td>
<td>Pyrite</td>
<td>II</td>
<td>Qz–Cp–Py vein</td>
<td>−1.4</td>
</tr>
<tr>
<td>FX-2804-18</td>
<td>Pyrite</td>
<td>II</td>
<td>Py–Cp–Qz vein</td>
<td>−1.3</td>
</tr>
<tr>
<td>FX-2604-16</td>
<td>Pyrite</td>
<td>II</td>
<td>Py–Cp–Qz vein</td>
<td>−1.5</td>
</tr>
<tr>
<td>15FX-43</td>
<td>Molybdenite</td>
<td>III</td>
<td>Qz–Mo–Py vein</td>
<td>2.3</td>
</tr>
<tr>
<td>FX-4204-19</td>
<td>Pyrite</td>
<td>III</td>
<td>Qz–Py–Mo vein</td>
<td>−1.8</td>
</tr>
<tr>
<td>FX-4201-17</td>
<td>Pyrite</td>
<td>III</td>
<td>Qz–Py veinlet</td>
<td>−1.9</td>
</tr>
</tbody>
</table>
Cu—Mo deposit in Great Xing’an Range (Li et al., 2007), the Yuchiling Mo deposit in eastern Qinling (Li et al., 2012), and the Qian’echung giant Mo deposit in the Dabie Shan (Yang et al., 2013). However, the occurrence of CO2-bearing fluids in the Fuxing deposit is limited in each stage of hydrothermal quartz veins, and detailed laser Raman spectroscopy analysis has not detected the CO2-bearing FIs. In summary, the ore-forming fluids of the Fuxing porphyry Cu deposit are characterized by high temperature, moderate salinity and relatively oxidized condition.

Projection of microthermometry data of all FIs on the salinity versus homogenization temperature diagram (Fig. 11) can be used to interpret the evolutionary trend of fluid. In the Fuxing porphyry Cu deposit, an initially supercritical fluid with low- to moderate-salinity (~6–8 wt.% NaCl equiv., ~600 °C) were thought to be exsolved from the parental magma chamber (Burnham, 1979; Hedengquist and Lowenstern, 1994). Fluid exsolved from this magma was characterized by high temperature, high salinity, and high oxygen fugacity. It seems the fluid boiling lead to the generation of hypersaline liquid phase (~57 wt.% NaCl equiv.; Fig. 11) and a low-density vapor, as is indicated by the coexistence of V-type and H-type inclusions in the same host mineral at early ore stage, which have various modes of homogenization, similar homogenization temperatures and great differences in salinity. Furthermore, FIs of the quartz-chalcopyrite and quartz-pyrite veins also record boiling conditions, but the liquid phase is cooler (234–393 °C) and more dilute (1.4–9.6 wt.% NaCl equiv.; Fig. 11). We therefore suggest that this fluid is also of magmatic origin, but it ascended and cooled more slowly such that it did not intersect its solvus curve until lower temperatures and pressures (~100 bars; ~1 km under hydrostatic pressure conditions; Hedengquist et al., 1998). Moreover, quartz in the late ore stage (IV) veins only contains L-type FIs, with lower homogenization temperatures (peak homogenization temperatures between 180 °C and 200 °C) and little variation in salinity (0.9 to 3.7 wt.% NaCl equiv.), which indicates that these fluids may be the cooled equivalent of the main ore stage (II and III) fluids (Fig. 11). However, the C–H–O stable isotopic data in this study suggest the involvement of external meteoric water throughout the hydrothermal evolution of the deposit.

7.3. Migration and precipitation mechanism

Most researchers have proposed that the migration of metal ions such as Cu and Mo mainly occurs as complexes in the ore solutions (Roedder, 1971; Holland, 1972; Crear and Barnes, 1976), as the complexes have high stability under condition of high temperature, high pressure, or supercritical state. Stage I fluids at Fuxing Cu deposit are characterized by high temperature, moderate salinity and high oxidized fugacity. Under such conditions, sulfur is usually present in the form of sulfates, which are favorable for porphyry magmas to incorporate chalcophile elements (e.g., S, Cu, and Zn) from the source region (Gaetani and Grove, 1997; Sillitoe, 1997; Mungall, 2002; Sun et al., 2015b), but sulfide mineralization would be limited. Because sulfur in ore minerals mainly exists in the reduced form of S2−. Accordingly, the early stage veins at Fuxing contain quartz ± K-feldspar ± magnetite ± pyrite. In contrast, the later main ore stage is characterized by various quartz-sulfide veins and associated chlorite–sericite alteration.

Ferrous ion is regarded as an effective reducing agent in magmas. It could play a crucial role in influencing the reduction related to sulfides of sulfates and the ore mineralization (Mungall, 2002; Sun et al., 2004; Liang et al., 2009). During the early stage in the Fuxing deposit, magnetite crystallization could lead to the progressive reduction of SO42− to S2−, based on the following reaction (Liang et al., 2009): 12[FeO] + H2SO4 = 4Fe3O4 + H2S, so that the ore stage II fluids would become more acidic with the increasing activities of H+ and S2−. Subsequently, the increase in PH and oversaturation of S2− in the hydrothermal fluids resulted in large-scale deposition of sulfate minerals (e.g., CuFeS2) in stage II generation veins.

Previous hydrothermal experiments have shown that fluid boiling can significantly alter the phase equilibrium of hydrothermal fluid (Zhang, 1997; Reed and Palandri, 2006; Wu et al., 2010; Gu et al., 2010; Wang et al., 2011). When the temperature decreased below ~400 °C, which is considered to be the transition temperature from lithostatic to hydrostatic conditions resulted from the change from ductile to brittle behavior of the system (Fournier, 1987, 1992), the...

### Table 5

Lead isotopic data of monzogranite and plagiogranite porphyry from the Fuxing porphyry Cu deposit.

<table>
<thead>
<tr>
<th>Sample no.</th>
<th>Rock</th>
<th>206Pb/204Pb</th>
<th>2σ</th>
<th>207Pb/204Pb</th>
<th>2σ</th>
<th>208Pb/204Pb</th>
<th>2σ</th>
</tr>
</thead>
<tbody>
<tr>
<td>15FX-89</td>
<td>Monzogranite</td>
<td>37.9</td>
<td>0.007</td>
<td>15.476</td>
<td>0.003</td>
<td>18.659</td>
<td>0.004</td>
</tr>
<tr>
<td>15FX-90</td>
<td>Plagiogranite</td>
<td>37.7</td>
<td>0.004</td>
<td>15.476</td>
<td>0.001</td>
<td>18.506</td>
<td>0.002</td>
</tr>
<tr>
<td>15FX-91</td>
<td>Plagiogranite</td>
<td>37.8</td>
<td>0.015</td>
<td>15.503</td>
<td>0.006</td>
<td>18.819</td>
<td>0.007</td>
</tr>
<tr>
<td>15FX-92</td>
<td>Plagiogranite</td>
<td>37.9</td>
<td>0.015</td>
<td>15.537</td>
<td>0.006</td>
<td>19.345</td>
<td>0.007</td>
</tr>
<tr>
<td>15FX-94</td>
<td>Plagiogranite</td>
<td>37.6</td>
<td>0.005</td>
<td>15.474</td>
<td>0.002</td>
<td>18.367</td>
<td>0.003</td>
</tr>
<tr>
<td>15FX-71</td>
<td>Plagiogranite</td>
<td>39.3</td>
<td>0.007</td>
<td>15.628</td>
<td>0.003</td>
<td>19.481</td>
<td>0.004</td>
</tr>
</tbody>
</table>

**Fig. 14.** (A) Histograms of $\delta^{34}$S values for sulfide minerals from the Fuxing porphyry Cu deposit. (B) Comparison with other typical porphyry deposits. Data sources: data of the Tuwu, Yuhai and Baogutu deposits in Chinese Tianshan and Junggar orogenic belt from Wang et al. (2015a), Zang (2014) and Zhang et al. (2010), respectively; data of the Butte deposit in USA, Erdenetuin-Obo deposit in Mongolia, and Mariquita and Lucky deposits in Mexico from Field et al. (2005), Sotnikov et al. (2004) and Salas et al. (2013), respectively.
fluid pressure would decrease rapidly, causing the supercritical fluid to intersect its two-phase surface and boiling, which would lead to the precipitation of metal sulfides. Fluid studies show that homogenization temperatures of ore-bearing veins from the quartz-sulfide stage at Fuxing concentrate on 340–380 °C and pressures are less than ~200 bars (Fig. 11), which suggests sulfides mainly precipitated below 380 °C with fluid boiling. It is also indicated that the decompression boiling of fluid is one of the significant factors for Cu deposition.

In addition, the precipitation of sulfides from ore solutions in the Fuxing deposit may be associated with the fluid mixing with meteoric water (Taylor, 1974; Simmons et al., 2005). The C–H–O isotopic signatures indicate that a larger amount of meteoric water was involved during fluid ascending, which were evidenced by an apparent reduction of ore-forming temperature and salinity of ore-forming solutions (Fig. 9). Decreases in the temperature, pressure and salinity, and increases in the PH would reduce the acid concentration and solubility of Cu in the fluid, which result in fast deposition of sulfide minerals in ore stage III, including chalcopyrite, molybdenite, sphalerite, and pyrite. Finally, with the inflow of abundant meteoric water into the ore-forming system at the shallow crust, the fluids evolved toward primarily meteoric, cooled and diluted conditions and generation of late stage (IV) veins. Therefore, the fluid mixing most probably is one of the main factors for polymetallic sulfide deposition.

The Fuxing deposit is located in the central part of the Dananhu–Tousuquan arc belt in Eastern Tianshan, adjacent to the Tuwu–Yandong Cu deposits. The formation of these porphyry Cu deposits was thought to be related to the evolution of the Dananhu–Tousuquan arc which has been widely discussed (Xiao et al., 2004; Mao et al., 2005; Shen et al., 2014a; Wang et al., 2015b; and references therein). Most geologists suggest that it had basically experienced the subduction tectonic stage of the Carboniferous or even earlier, the post-collisional extension since the Early Permian and intracontinental extension since the Triassic, while the porphyry Cu deposits in the Dananhu–Tousuquan arc were formed in the subduction tectonic setting (Su et al., 2012; Mao et al., 2005; Zhang et al., 2008; Han et al., 2014; Xiao et al., 2015). Numerous studies have shown that they were related to the Carboniferous magmatic rocks, including diorite, quartz diorite, monzogranite, quartz porphyry, and plagiogranite porphyry, which possess geochemical features of subduction-related arc setting and mantle peridotites interaction (Shen et al., 2014a,b; Sun et al., 2015a; Xiao et al., 2015; Wang et al., 2015a,b). Subduction processes could facilitate the change of ore-forming pressure from lithostatic pressure to hydrostatic pressure. An abrupt decline of pressure could have caused the boiling of ore-forming fluid and promoted the enrichment and precipitation of ore minerals along with the decline of temperature (e.g., Meinert et al., 2003). Moreover, the SIMS zircon U–Pb ages of plagiogranite porphyry (332.1 ± 2.2 Ma; Wang et al., 2016c) and monzogranite (328.4 ± 3.4 Ma; Wang et al., 2016c) further indicate close relationship between the Cu mineralization and the Carboniferous magmatic activities at Fuxing in the Dananhu–Tousuquan arc belt. In addition, the δ34S isotopic data of the Fuxing deposit ranging from ~1.9 to 2.3‰ indicates that the metatelic materials were mainly derived from the deep-seated magma, which has a certain geochemical affinity with the adakites.

8. Conclusions

(1) The Fuxing Cu deposit is a recently discovered porphyry ore system in Eastern Tianshan. Its mineralization process includes four stages, characterized by barren quartz vein stage I, quartz–chalcopyrite–pyrite vein stage II, quartz–polymeretic sulfide vein stage III, and quartz–calcite vein stage IV, respectively.

(2) The FiS of the Fuxing deposit are primarily liquid-rich two-phase, vapor-rich two-phase, and halite-bearing inclusions. The ore-forming fluids are characterized by high temperature, moderate salinity, and relatively oxidized condition.

(3) Carbon, hydrogen and oxygen isotopic characteristics indicate that the ore-forming fluids of the Fuxing deposit were derived from magmatic sources that were mixed with some meteoric water in the late ore stage. Sulfur and lead isotopes suggest that ore-forming metals were primarily derived from a deep-seated magma.

(4) The Fuxing deposit is a shallow-seated porphyry system with mineralization depth of ~1 km similar to most porphyry deposits in magmatic arc settings. The changes of oxygen fugacity, decompression boiling, and local mixing with meteoric water most likely played a significant role for Cu precipitation.

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