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Abstract: The Weiquan Ag-polymetallic deposit is located on the southern margin of the Central Asian Orogenic Belt and in the western segment of the Aqishan–Yamansu arc belt in East Tianshan, northwestern China. Its orebodies, controlled by faults, occur in the lower Carboniferous volcano-sedimentary rocks of the Yamansu Formation as irregular veins and lenses. Four stages of mineralization have been recognized on the basis of mineral assemblages, ore fabrics, and crosscutting relationships among the ore veins. Stage I is the skarn stage (garnet + pyroxene), Stage II is the retrograde alteration stage (epidote + chlorite + magnetite ± hematite ± actinolite ± quartz), Stage III is the sulfide stage (Ag and Bi minerals + pyrite + chalcopyrite + galena + sphalerite + quartz ± calcite ± tetrahedrite), and Stage IV is the carbonate stage (quartz + calcite + pyrite). Skarnization, silicification, carbonatization, epidotization, chloritization, sericitization, and actinolitization are the principal types of hydrothermal alteration. LA–ICP–MS U–Pb dating yielded ages of 326.5±4.5 and 298.5±1.5 Ma for zircons from the tuff and diotite porphyry, respectively. Given that the tuff is wall rock and that the orebodies are cut by a late diotite porphyry dike, the ages of the tuff and the diotite porphyry provide lower and upper time limits on the age of ore formation. The δ¹³C values of the calcite samples range from −2.5% to 3.3%, the δ¹⁸O_H₂O and δDV_SMW values of the sulfide stage (Stage III) vary from 1.1% to 5.2%, and −111.7% to −66.1%, respectively, and the δ¹³C, δ¹⁸O_H₂O and δDV_SMW values of calcite in one Stage IV sample are 1.5%, −0.3%, and −115.6%, respectively. Carbon, hydrogen, and oxygen isotopic compositions indicate that the ore-forming fluids evolved gradually from magmatic to meteoric sources. The δ³⁴S_VCDT values of the sulfides have a large range from −6.9% to 1.4%, with an average of −2.2%, indicating a magmatic source, possibly with sedimentary contributions. The ²⁰⁶Pb/²³⁸Pb, ²⁰⁷Pb/²³⁵Pb, and ²⁰⁸Pb/²³⁵Pb ratios of the sulfides are 17.9848–18.2785, 15.5188–15.6536, and 37.8125–38.4650, respectively, and one whole-rock sample at Weiquan yields ²⁰⁶Pb/²³⁸Pb, ²⁰⁷Pb/²³⁵Pb, and ²⁰⁸Pb/²³⁵Pb ratios of 18.2060, 15.5674, and 38.0511, respectively. Lead isotopic systems suggest that the ore-forming materials of the Weiquan deposit were derived from a mixed source involving mantle and crustal components. Based on geological features, zircon U–Pb dating, and C–H–O–S–Pb isotopic data, it can be concluded that the Weiquan polymetallic deposit is a skarn type that formed in a tectonic setting spanning a period from subduction to post-collapse. The ore materials were sourced from magmatic ore-forming fluids that mixed with components derived from host rocks during their ascent, and a gradual mixing with meteoric water took place in the later stages.

Key words: zircon U–Pb geochronology, C–H–O–S–Pb isotopes, skarn-type Ag–polymetallic deposit, Weiquan, East Tianshan, China

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1 Introduction

The Central Asian Orogenic Belt (CAOB) is bounded by the European, Siberia, Tarim, and North China cratons (Fig. 1), and is one of the largest and most complex of the Paleozoic accretionary orogens worldwide, showing evidence of considerable juvenile crustal growth (Sengör et al., 1993; Jahn et al., 2000, 2004; Windley et al., 2007; Xiao et al., 2010, 2013; Mao et al., 2014; Wang et al., 2015a). The CAOB resulted from the collision of the Siberia and Tarim–North China plates along the Solonker–Kumishi Suture, which progressively closed eastwards from the late Carboniferous to the early Triassic (Chen Yanjing, 1996, 2000a; Xiao et al., 2003, 2009). The East Tianshan is located between the Tarim and Junggar basins, is an important constituent of the CAOB (Mao et al., 2005; Xiao et al., 2008; Zhang et al., 2014a; Zhou Hai et al., 2016; Yang Wei et al., 2017), and is one of the important metallogenic provinces in Xinjiang, NW China (Charvet et al., 2007; Mao et al., 2008; Pirajno et al., 2011; Tang et al., 2012; Ding Jianhua et al., 2016). Many medium- to large-scale ore deposits have been discovered in the region (Qin et al., 2003; Wang Jingbin et al., 2006; Zhang et al., 2008; Pirajno, 2013; Deng et al., 2017; Fig. 1), including orogenic Au deposits (e.g., Kanggar, Jinwozi, and Hongshi; Zhang et al., 2003a; Zhang et al., 2014b; Wang et al., 2015c, 2016a), epithermal gold deposits (e.g., Mazhuangshan; Chen et al., 2012b), skarn Fe deposits (e.g., Yamansu; Hou et al., 2014), porphyry Cu–Mo deposits (e.g., Tuwu, Yandong, Linglong, Sanchakou, Fuxing, Chihu and Yuhai; Qin Kezhang et al., 2009; Wang Yinhong et al., 2014; Wang et al., 2015a, 2015b, 2016b, 2016d; Wang and Zhang, 2016c, Zhang et al., 2016a), Zn–Pb deposits (e.g., Caixishan; Li et al., 2016), Mo and W deposits (e.g., Donggebi, Baishan, and Xiaobaishitou; Xiang Nan et al., 2013; Wu et al., 2014,

![Fig. 1](image-url)
2017; Deng et al., 2017), and magmatic Cu–Ni sulfide deposits (e.g., Xiangshan, Huangshan, and Huangshandong; Mao et al., 2008; Pirajno et al., 2008, 2011; Qin et al., 2011). Therefore, the East Tianshan was recommended as one of 16 Key Exploration Areas by the China Geological Survey. However, previous studies of the East Tianshan orogenic belt have focused mainly on the Paleozoic Cu, Mo, Au, and Fe mineralization, and there are few reports on the Ag-polymetallic deposits, which in turn hinders the exploration for similar deposits in the region.

The Weiquan deposit is a medium-scale Ag deposit to have been discovered in the East Tianshan orogenic belt (Fig. 1c), and it has proven metal reserves of 532 t Ag and 15,000 t Cu. It is located ~180 km southwest of Hami City, and was discovered in 2000 by the No. 1 Geological Party of the Xinjiang Bureau of Geology and Mineral Resources. However, there has been little research on its ore genesis, and only a minimal amount of basic geological research has been carried out (Mao Jingwen et al., 2002; Pan Chengze et al., 2005; Wang Longsheng et al., 2005; Feng jing et al., 2008; Wang Xinkun et al., 2008; Liu Yan et al., 2017), which means that the sources of ore-forming fluids and the origin of the ore-forming constituents in the Weiquan deposit have not been determined. In particular, there has been a lack of stable and radioactive isotope studies. Moreover, the timing of the metallogenesis remains debated.

In this paper we provide new geological, geochronological, and isotopic data that provide insights into the genesis of the Weiquan deposit. In particular, we provide the results of LA–ICP–MS zircon U–Pb dating and analyses of C–H–O–S stable and Pb radiogenic isotopes of the Weiquan Ag-polymetallic deposit. We discuss the sources of the fluids and ore metals as well as the ore-forming mechanism. The results will aid in fundamental research and exploration in the East Tianshan orogenic belt and more generally in the CAOB.

2 Geological Setting

2.1 Regional geology

The East Tianshan lies between the Turpan–Hami (commonly abbreviated as Tuha) Basin in the north and the Tarim Basin in the south (Mao et al., 2005; Zheng Jiahao et al., 2016). It records a complex tectonic evolution from the late Paleozoic to the Mesozoic, involving subduction of the Paleo-Tianshan oceanic crust, collision–accretionary events, strike-slip faulting, post-collisional events, and intracontinental extension between the Tarim and the Junggar basins (Xiao et al., 2003, 2013; Zhang et al., 2008; Pirajno et al., 2011). Three major tectonic units can be recognized in the East Tianshan (Fig. 1c): (1) the Bogeda–Harlik Belt in the north, (2) the Jueluotage Belt in the middle, and (3) the Central Tianshan Massif in the south (Su et al., 2011).

The Bogeda–Harlik Belt comprises well-developed Ordovician–Carboniferous volcanic rocks, sedimentary rocks, granites, and mafic–ultramafic complexes (Li Jinyi et al., 2006; Ma Xinghua et al., 2015). One of the representative mafic–ultramafic complexes in the Bogeda–Harlik Belt is the Heishanxia Complex, which is composed of peridotite, olivine/plagioclase pyroxenite, gabbro, and diorite (Su et al., 2011).

The Jueluotage Belt can be further subdivided into three parts divided by the Kanggu and Yamasun faults: (1) the Dananhu–Tousuquan arc belt (to the north), (2) the Kanggu–Huangshan ductile shear zone (central), and (3) the Aqishan–Yamasun arc belt (south) (Fig. 1c; Qin Kezhang et al., 2002; Mao et al., 2005; Huang et al., 2013; Gao et al., 2015). The Dananhu–Tousuquan arc belt consists mainly of lower Devonian volcanic and clastic sedimentary rocks of the Dananhu Formation, lower Carboniferous turbidites of the Gandun Formation, lower-middle Carboniferous basaltic to andesitic volcanic rocks and sedimentary rocks of the Qi’eshan Group, Permian calc-alkaline volcanic, pyroclastic, and clastic rocks, Jurassic sandstones, and Cenozoic cover (Mao et al., 2005; Gao et al., 2015). It hosts several porphyry copper deposits (e.g., Tuwu and Yandong; Zhang et al., 2006; Han et al., 2014; Shen et al., 2014a, b) and VMS copper–zinc deposits (e.g., Kalatag; Deng et al., 2016). The Kanggu–Huangshan ductile shear zone is located between the Kanggu and Yamasun faults, and it records a widespread lower-greenschist facies metamorphism and strong ductile deformation. The rocks of the shear zone include Devonian–Carboniferous volcanioclastic rocks, basalts, tuffs, limestones, sandstones, and an ophiolitic slice (Mao et al., 2008, 2014; Xiao et al., 2013), and the shear zone hosts gold deposits (e.g., Kanggu; Wang et al., 2004), porphyry deposits (e.g., Baishan; Zhang et al., 2005), and Cu–Ni deposits (e.g., Huangshan; Zhou et al., 2004). The Aqishan–Yamasun arc belt lies between the Yamasun and Aqikuduke faults, and the main outcropping strata are early Carboniferous basalts, andesites, dacites, and tuffs of the Yamasun Formation and late Carboniferous rhyolites of the Tugutubulake Formation (Xiao et al., 2013). The arc belt hosts several Fe deposits and Cu–Ag–Pb–Zn skarn deposits (Chen et al., 2012a; Hou et al., 2014).

The Central Tianshan Massif, bounded by the Aqikuduke Fault to the north and the Xingxingxia Fault to the south (Xiao et al., 2004) (Fig. 1c), consists of a Precambrian crystalline basement with minor volumes of
early and late Paleozoic volcano-sedimentary formations (Qin Kezhang et al., 2002; Shu Liangshu et al., 2002; Liu et al., 2004b; Li et al., 2007; Lei et al., 2011). It hosts hydrothermal magnetite deposits (e.g., Tianhu; Huang et al., 2015) and sedimentation reformation type Pb–Zn deposits (e.g., Caixia Shan; Li et al., 2016). Granitoid intrusions of various ages, mostly I-type, are widespread in the Central Tian Shan (Wang Jingbin et al., 2006).

2.2 Ore deposit geology

The Weiquan Ag–polymetallic deposit (91°43′E, 41°52′N) is located in the western part of the Aqishan–Yamansu arc belt, East Tian Shan, close to the E–W trending Yamansu Fault (Fig. 1c). The main strata exposed in the Weiquan area belong to the lower Carboniferous Yamansu and upper Carboniferous Tugutubuke formations, which are separated by a fault. The 1830 m thick Yamansu Formation consists mainly of intermediate–acid tuffs, sedimentary tuffs, tuffaceous sandstones, clastic rocks, and carbonate rocks, and it strikes roughly E–W and dips to the south at 35°–75°. The Tugutubuke Formation crops out to the southwest of the Weiquan area, and consists of andesites, tuffs, and breccia. These Carboniferous formations were intruded by the Bailingshan pluton, and they exhibit a widespread thermal contact metamorphism. The lower Carboniferous Yamansu Formation is the main wall rock of the Weiquan deposit.

Throughout the Weiquan area, a series of secondary faults are well developed (Fig. 2), and they can be subdivided into two groups according to their crosscutting relationships: (1) the nearly NW–SE and E–W trending faults that pre-date and host the main mineralization; and (2) the later NE–SW trending faults that cut the NW–SE and E–W trending faults.

Carboniferous intrusive rocks are widespread in the Weiquan deposit area, and they cover an area of 5 km², mainly in the southern and southwestern parts of the mine area. The rocks belong to the calc-alkaline series and are classified as I-type granites (Zhang Zengjie et al., 2003b). They are part of the Bailingshan pluton (297±3 Ma; Wang

![Fig. 2. Simplified geological map of the Weiquan deposit (based on a map by the No. 1 Geological Party of the Xinjiang Bureau of Geology and Mineral Resources, 2004).](image-url)
Longsheng et al., 2005), which consists of granite, granodiorite, diorite, quartz diorite, and monzogranite. The Bailingshan granitoids probably represent melts that were derived from juvenile lower crust and mixed with mantle-derived material, and with the mantle components playing an increasingly prominent role over time (Zhang et al., 2016). In addition, there are dikes of diorite porphyry and granite porphyry in the mine area, and the dikes of diorite porphyry cut the orebody, as seen in underground tunnels (Fig. 4f). This indicates that the diorite porphyry developed after the mineralization, and the age of the diorite porphyry therefore provides an upper time limit for the age of ore formation.

Thirty-two Cu–Ag and 24 Pb–Zn orebodies have been identified in the mine area, including three relatively large Cu–Ag orebodies (named I, II, and III) and one representative Pb–Zn orebody known as L1-6 (Fig. 2). The orebodies occur as veins, lenses, or layers, and are hosted mainly in the garnet skarn zone. The No. I Cu–Ag orebody is aligned E–W and dips generally to the north at 42°–67° (avg. 50°). It has an average length of 190 m (maximum 270 m), a thickness of 0.67–14.78 m (avg. 5.05 m), and an average grade of 331.1 ppm Ag (range from 70.2 ppm to 1902 ppm) and 0.73% Cu (range from 0.3% to 2.4%). The explored orebodies extend down-dip for more than 300 m (Fig. 3). The No. II Cu–Ag orebody occurs about 560 m northeast of the No. I Cu–Ag orebody. It strikes 78° and dips 60°–64° (avg. 62°) to the north, and its outcrop extends for 200 m, averages 5.32 m in thickness (ranges from 0.44 m to 12.22 m), and has an average grade of 328.1 ppm Ag (range from 90.9 ppm to 1085.9 ppm) and 0.21% Cu (maximum 4.63%). The explored orebodies extend down-dip for more than 30 m, and exhibit vein and lenticular shapes on the surface. The blind No. III Cu–Ag orebody is located between the No. I and No. II Cu–Ag orebodies, and it dips generally to the north at 57°. It has an average thickness of 17.22 m and an average grade of 421.5 ppm Ag (maximum 1540 ppm). The one representative Pb–Zn orebody (L1-6) dips generally to the south at 53°–61° (avg. 57°). It extends for 400–1400 m, has a thickness of 1.02–31.26 m (average 11.19 m), and an average grade of 1.29% Zn (range from 0.54% to 3.12%) and 0.33% Pb (range from 0.16% to 1.62%). The principal economic ore mineral is sphalerite (Fig. 4i) with lesser amounts of galena (Fig. 4h). We propose that the Cu–Ag and Pb–Zn mineralization took place roughly in two separate periods: the earlier Cu–Ag mineralization produced large amounts of calcite, argentite, natural silver, and natural bismuth, along with minor peacreeite and bornite; the later Pb–Zn mineralization produced massive sphalerite, galena, chalcopyrite, and pyrite, as well as minor tetrahedrite.

More than 30 minerals have been identified in the Weiquan deposit. The main ore minerals are native silver, native bismuth, argentite, chalcopyrite, sphalerite, galena, limonite, and pyrite. The subordinate metal minerals are chalcocite, matildite, magnetite, bornite, covellite, arsenopyrite, and tetrahedrite, and rare metal minerals include pyrrhotite, pyrrhotite, apahenite, and hematite. The main gangue minerals are garnet, actinolite, quartz, epidote, chlorite, and calcite, with smaller amounts of biotite, plagioclase, and gypsum, amongst others. Of note, we found native bismuth in carbonate (Fig. 5g), which is relatively rare (Liu Yan et al., 2017). The ore textures include euhedral granular, subhedral granular, anhedral granular, emulsion, skeleton, and metasomatic relict textures. The main ore structures are massive, disseminated, and vein structures.

Due to thermal metamorphism and subsequent metasomatism, the wall rocks of the Weiquan deposit were strongly altered to form a typical calc-silicate mineral assemblage, dominated by garnet and diopside,
Fig. 4. Photographs of the Weiquan ore deposit.
(a), Tuff and skarnized tuff of the Yumansu Formation; (b), garnet aggregates in altered tuff; (c), skarn cut by retrograde stage epidote; (d), epidote replacing and pseudomorphing garnet, and pyrite + galena + sphalerite replacing bands of garnet and filling the intertices; (e), calcite vein in garnet skarn; (f), orebody cut by the late-stage diorite porphyry; (g), scattered actinolite in skarn; (h), massive galena in skarn; (i), massive sphalerite and minor chalcopyrite in skarn; (j), native silver flakes or film in clinoelite affected by carbonatization; (k), quartz–pyrite veinslets overprinting garnet skarn; and (l), quartz–calcite veinslets overprinting garnet skarn. Abbreviations: Act = actinolite, Cal = calcite, Ccp = chalcopyrite, Ep = epidote, Gt = garnet, Py = pyrite, Qtz = quartz, Siv = native silver, Sp = sphalerite.

and an intensely developed retrograde mineral assemblage, dominated by epidote, chlorite, actinolite, magnetite, quartz, and calcite. The hydrothermal alteration in the Weiquan deposit can be divided into the following three zones from orebody to wall rock: (1) a carbonatization zone (polymetallic sulfides and calcite); (2) a chloritization–actinolitization zone (epidote, chlorite, and actinolite); and (3) a pyritization zone (epidote, chlorite, and pyrite).

Based on field investigations and petrographic observations, the Weiquan endogenous mineralization can be divided into a skarn stage (Stage I), a retrograde alteration stage (Stage II), a sulfide stage (Stage III), and a carbonate stage (Stage IV). Finally, a period of supergene mineralization affected the orebody. The paragenetic sequences are illustrated and summarized in Fig. 6.

Similar to many skarn systems, the earliest paragenetic stage (Stage I) is characterized by the formation of the anhydrous assemblage of garnet and pyroxene (Fig. 4a–b, 5d), and there was no extensive ore mineralization at this
Fig. 5. Photomicrographs of representative samples from the Weiquan deposit.
(a), Stage I Gt replaced by retrograde stage Ep and Cal, plane-polarized light; (b), retrograde stage actinolite; cross-polarized light; (c), Stage I Gt cut by retrograde stage calcite; plane-polarized light; (d), pyrite in interstitial spaces of the garnets; reflected light; (e), chalcopyrite and sphalerite occurring in the interstitial spaces of the garnets; reflected light; (f), pyrite with inclined-granular structure was replaced by magnetite; reflected light; (g), calcite + matildite + native bismuth assemblage showing matildite replacing native bismuth; reflected light; (h), argentite and native silver replacing arsenopyrite; reflected light; (i), chalcopyrite was replaced by the assemblage of argentite + native silver; reflected light; (j), native silver replacing pearceite, and was replaced by argentite; reflected light; (k), paragenesis of pyrite, chalcopyrite, tetrahedrite and magnetite; reflected light; (l), the relationship of sulfide assemblages in the interstitial sulfides; reflected light; (m), the magnetite was replaced by quartz-chalcopyrite-tetrahedrite; reflected light; (n), paragenesis of quartz, chalcopyrite, tetrahedrite and sphalerite; reflected light; (o), sulfide assemblages showing chalcopyrite occurring as veinlet filling microfractures in pyrite; reflected light; Abbreviations: Act = actinolite, Ar = argentite, Bsm = native bismuth, Cal = calcite, Cep = chalcopyrite, Ep = epidote, Gn = galena, Gt = garnet, Mag = magnetite, Prt = pearceite, Py = pyrite, Qtz = quartz, Slv = native silver, Sp = sphalerite, Td = tetrahedrite.
### Table: Mineral Paragenesis for the Weiquan Deposit

<table>
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<th>Minerals</th>
<th>Skarn stage (Stage I)</th>
<th>Retrograde alteration (Stage II)</th>
<th>Sulfide stage (Stage III)</th>
<th>Carbonate stage (Stage IV)</th>
<th>Supergene period</th>
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Fig. 6. Mineral paragenesis for the Weiquan deposit.

Stage II is characterized by the replacement of garnet and pyroxene by epidote, chlorite, actinolite, quartz, calcite, and magnetite (Figs. 4c, 4g, 5a–b). Magnetite occurs mainly in exoskarn, either in disseminated or massive forms (Fig. 5f, 5k, 5m). Epidote is a representative mineral of the retrograde alteration stage (Fig. 5a). Stage III is the main stage of sulfide mineralization and is characterized by the assemblage of pyrite, chalcopyrite, magnetite, galena, sphalerite, tetrahedrite, natural silver, natural bismuth, argentite, matildite, and pearsite (Fig. 5e–o). Stage IV, the carbonate stage, is characterized by the formation of low-temperature minerals, especially calcite, along with minor quartz and a little sulfide, and these Stage IV minerals occur either as veinlets crosscutting skarn and sulfides, or as fracture fill within skarn minerals (Figs. 4l, 4e, 5c).

### 3 Sampling and Analytical Methods

#### 3.1 LA–ICP–MS zircon U–Pb dating

Samples were collected from unaltered tuff (14WQ-01, location: 91°44′6″E, 41°52′12″N) and the diorite porphyry dike that cuts the orebodies (14WQ-08, location: mine stope 1009 level) (Fig. 2).

Zircons were extracted by conventional heavy liquid and magnetic techniques, and then purified by hand-picking under a binocular microscope at the Hebei Regional Geological Survey Research Institute in Langfang. Representative zircon grains were mounted in epoxy resin and polished to expose the grain centers, and all the grains were then photographed under transmitted light, reflected light, and cathodoluminescence (CL) to examine inner structures, fluid inclusions, and cracks before selecting suitable crystals and sites for analysis. CL imaging was carried out using a LEO1450VP scanning electron microscope (SEM) at the Institute of Geology and Geophysics, Chinese Academy of Sciences, Beijing, China.

Zircon U–Pb and trace element analyses were carried out at the State Key Laboratory of Geological Processes and Mineral Resources, China University of Geosciences (Beijing). The instrument couples a quadrupole ICP–MS
(Agilent 7500a) and a UP-193 Solid-State laser (193 nm, New Wave Research) with an automatic positioning system. Helium was used as the carrier gas with a flux of 0.8 L/min to provide efficient aerosol transport to the ICP and minimize aerosol deposition. Concentrations of uranium, thorium, lead, and trace elements were calibrated using 25Si as an internal standard and NIST 610 as the external reference standard. U–Pb isotope fractionation effects were corrected using zircon 91500 (Wiedenbeck et al., 1995) as the external standard. Zircon standard TEMORA (417 Ma) from Australia (Black et al., 2003) was used as a secondary standard to monitor the deviations of age measurements and calculations. The instrument protocols and data acquisition methods used during this study are the same as those described by Song et al. (2010). The isotopic ratios and element concentrations of the zircons were calculated using Glitter (ver. 4.0, Macquarie University). Concordia plots and weighted means were obtained using Isoplot/Ex (3.0) (Ludwig, 2003). Age uncertainties are cited as 2σ values, and weighted mean 206Pb/238U ages are quoted at 95% confidence levels.

3.2 C–H–O Isotopes

Seven representative calcite samples related to mineralization were selected for C–H–O isotopic analyses, and one representative quartz sample related to mineralization was selected for H–O isotopic analyses at the Analytical Laboratory of the Beijing Research Institute of Uranium Geology. The fresh samples were crushed to 40–60 mesh size. The samples were chosen on the basis of field investigations and microscopic observations, and the grains of fresh calcite and quartz were selected by hand from ore samples under a binocular microscope. The purity of each mineral separate was greater than 99%. All of the mineral separates were cleaned in an ultrasonic bath before being powdered in an agate mortar.

The C–O isotopic compositions of calcite crystals were obtained using a Finnigan MAT253 isotope ratio mass spectrometer. The calcite was reacted with pure phosphoric acid to produce CO2. The isotopic compositions were measured following the detailed procedures outlined by Zhou et al. (2013). The analytical precisions (2σ) were ±0.1‰ for C isotopes and ±0.2‰ for O isotopes. The δ13C and δ18O values are reported relative to Pee Dee Belemnite (PDB). The values of δ18O are calculated from δ18O = (1000x−1000)/δPDB + 1000 (Coplen et al., 1983).

The H–O isotopic compositions were measured by a MA1253 mass spectrometer. Oxygen was liberated from the quartz by reaction with BrF5 (Clayton et al., 1972). The δD values of the water extracted from fluid inclusions in the minerals were determined using the Zn reduction method (Coleman et al., 1982). The results are reported relative to Standard Mean Ocean Water (SMOW), with analytical precisions of ±1‰ for the H isotopes. The mineral–H2O fractionation factors of oxygen isotopes are from O’Neil et al. (1969) (calcite–H2O: 1000lnα =2.78×10^6/T^2−3.39) and Clayton et al. (1972) (quartz–H2O: 1000lnα =3.38×10^6/T^2−3.40). The calculations of the fractionation factors were performed using the mean value of the total homogenization temperatures of the associated fluid inclusions.

3.3 S and Pb isotopes

Most of the samples selected for S and Pb isotope analyses were collected from the 1099 level of the underground tunnels in the Weiqian deposit, and the remainder were from drill cores. S isotope analyses of argentite, chalcocite, pyrite, chalcopyrite, and sphalerite were completed at the Analytical Laboratory of the Beijing Research Institute of Uranium Geology. The Pb isotope compositions of argentite, chalcocite, chalcopyrite, pyrite, pyrrhotite, sphalerite, and galena were measured at the State Key Laboratory for Mineral Deposits Research, Nanjing University. Mineral separates were carefully crushed and sieved through 40–60 meshes and sorted to a purity of >99% using a binocular microscope. For the sulfur and lead isotope analyses, 200 mesh sulfides were used.

The S isotopic compositions were determined using a Delta V Plus isotope ratio mass spectrometer with an accuracy of better than ±0.2‰. The δ34S of sulfide was measured from SO2 obtained by placing the sulfide–CuO composite (at a weight ratio of 1:7) into a vacuum system heated to 980°C (Robinson and Kusakabe, 1975). The standard materials used for determining sulfur isotopic ratios in sulfides were GBW-04414 and GBW-04415 Ag2S with δ34S values of −0.07 ± 0.13‰ and 22.15±0.14‰, respectively. The results are reported relative to the standard of Vienna Canon Diablo Troilite (V-CDT) sulfide.

The Pb isotopic compositions were analyzed using a Thermo-Finnigan Triton T1 thermal ionization mass spectrometer (TIMS). Samples were completely dissolved in an ultrapure mixed solution of HNO3 and HCl, and then a basic anion exchange resin was used to purify the Pb. Samples and standards were corrected for mass fractionation using thallium and an exponential law (Albarède et al., 2004). The Pb isotopic ratios were corrected according to the values of NBS-981. The 2σ errors on the T1-corrected NIST 981 values were measured during the period of sample analyses.
4 Results

4.1 Zircon U–Pb dating

CL images of representative zircon grains are shown in Fig. 7 and analytical data are listed in Table 1. The mineralogy and compositional features of the zircons were taken into consideration when determining the genetic type. In general, oscillatory zoning, high Th and U concentrations, and high values of Th/U indicate a magmatic origin, while obscure zoning, low Th and U concentrations, and low values of Th/U indicate a metamorphic origin (Sun Min and Guan Hong, 2001).

The CL images of zircons from sample 14WQ-01 show oscillatory zoning (Fig. 7a), indicating a magmatic origin. Zircons from the sample are colorless with no obvious inclusions. The crystals range in length from 30 to 50 μm, with length/width ratios of 2:1 to 3:1. The Th, U, and Pb concentrations in 5 zircon grains are in the ranges (110–1861) ppm, (138–863) ppm, and (8.64–66.78) ppm, respectively, with Th/U ratios of 0.61–2.41. The high Th/U ratios also indicate a magmatic origin. All the analyses plot on or near the concordia line. Their $^{206}\text{Pb}/^{238}\text{U}$ ages range from 321 to 333 Ma (1σ error), and they yield a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 326.5±4.5 Ma (n=5, MSWD=0.87) (Fig. 8a and Table 1). This age is interpreted as the time of deposition of the tuffaceous strata in the Yamansu Formation associated with the Weiquan deposit.

The zircons in sample 14WQ-08 are 80–120 μm long and have length/width ratios of 2:1 to 3:1. Eleven analyses conducted on the rims showed Th concentrations of (3–837) ppm and U concentrations of (9–1079) ppm, yielding Th/U ratios of 0.30 to 1.74, consistent with a magmatic origin. Their common Pb concentrations are (0.48–59.72) ppm. All the analyses plot on the concordia line, and they yield a concordia mean age of 298.5±1.5 Ma (n=11, MSWD=0.14).

Trace element analyses of the zircons from sample 14WQ-08 reveal that the magmatic rims display significant depletions in LREEs with (La/Yb)$_n$=0.00007–0.00055, positive Ce anomalies with δCe=18.21–38.33, and large negative Eu anomalies with δEu=0.06–0.47 (Fig. 9 and Table 2). These characteristics are typical of magmatic zircons (Hoskin and Ireland, 2000).

4.2 Carbon and oxygen isotopes

The results of our analyses of carbon and oxygen isotopes are presented in Table 3 and Fig. 10. The average fluid inclusion temperature (Tf) for each stage was used to calculate the δ$^{18}$O$_{water}$ value. Eighty-three primary fluid
<table>
<thead>
<tr>
<th>Spot</th>
<th>Element (ppm)</th>
<th>Measured isotope ratio</th>
<th>Apparent age (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Tb/U</td>
<td>3^208Pb/206Pb</td>
<td>208Pb/206Pb</td>
</tr>
<tr>
<td></td>
<td></td>
<td>207Pb/206Pb</td>
<td>1.0</td>
</tr>
<tr>
<td>1</td>
<td>560.07</td>
<td>341.25</td>
<td>1.64</td>
</tr>
<tr>
<td>2</td>
<td>233.31</td>
<td>366.47</td>
<td>0.61</td>
</tr>
<tr>
<td>3</td>
<td>1139.24</td>
<td>863.38</td>
<td>1.32</td>
</tr>
<tr>
<td>4</td>
<td>110.43</td>
<td>137.98</td>
<td>0.80</td>
</tr>
<tr>
<td>5</td>
<td>1860.67</td>
<td>771.77</td>
<td>2.41</td>
</tr>
</tbody>
</table>

Table 2 LA–ICP–MS analytical results of zircon trace elements (ppm)

<table>
<thead>
<tr>
<th>Spot</th>
<th>Y</th>
<th>La</th>
<th>Ce</th>
<th>Pr</th>
<th>Nd</th>
<th>Sm</th>
<th>Eu</th>
<th>Tb</th>
<th>Dy</th>
<th>Ho</th>
<th>Er</th>
<th>Yb</th>
<th>Lu</th>
<th>(La/Yb)</th>
<th>ΣREE</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>360.21</td>
<td>0.047</td>
<td>4.39</td>
<td>0.059</td>
<td>0.27</td>
<td>0.81</td>
<td>0.30</td>
<td>4.83</td>
<td>2.09</td>
<td>29.83</td>
<td>13.60</td>
<td>64.31</td>
<td>200.23</td>
<td>28.88</td>
<td>0.00017</td>
</tr>
<tr>
<td>2</td>
<td>545.72</td>
<td>0.046</td>
<td>5.49</td>
<td>0.042</td>
<td>0.22</td>
<td>1.14</td>
<td>0.49</td>
<td>7.54</td>
<td>3.26</td>
<td>45.30</td>
<td>18.37</td>
<td>87.96</td>
<td>226.12</td>
<td>39.26</td>
<td>0.00012</td>
</tr>
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<td>3</td>
<td>441.03</td>
<td>0.056</td>
<td>4.89</td>
<td>0.031</td>
<td>0.38</td>
<td>0.29</td>
<td>0.50</td>
<td>6.52</td>
<td>2.75</td>
<td>37.05</td>
<td>15.18</td>
<td>72.72</td>
<td>190.40</td>
<td>34.30</td>
<td>0.00018</td>
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<td>4</td>
<td>521.56</td>
<td>0.025</td>
<td>5.44</td>
<td>0.034</td>
<td>0.28</td>
<td>0.89</td>
<td>0.37</td>
<td>7.02</td>
<td>3.19</td>
<td>41.90</td>
<td>17.18</td>
<td>81.08</td>
<td>21.41</td>
<td>256.48</td>
<td>37.80</td>
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<tr>
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<td>367.74</td>
<td>0.060</td>
<td>4.05</td>
<td>0.065</td>
<td>1.20</td>
<td>0.29</td>
<td>6.50</td>
<td>2.54</td>
<td>31.25</td>
<td>12.31</td>
<td>56.82</td>
<td>13.83</td>
<td>159.99</td>
<td>21.91</td>
<td>0.00027</td>
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<td>713.21</td>
<td>0.049</td>
<td>4.70</td>
<td>0.037</td>
<td>0.47</td>
<td>1.64</td>
<td>0.48</td>
<td>10.90</td>
<td>4.37</td>
<td>58.62</td>
<td>23.25</td>
<td>105.95</td>
<td>26.91</td>
<td>304.54</td>
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<td>7</td>
<td>2144.32</td>
<td>0.463</td>
<td>6.18</td>
<td>3.01</td>
<td>42.83</td>
<td>41.47</td>
<td>19.16</td>
<td>98.30</td>
<td>25.56</td>
<td>242.14</td>
<td>69.69</td>
<td>266.87</td>
<td>58.84</td>
<td>607.66</td>
<td>82.53</td>
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<tr>
<td>8</td>
<td>1616.34</td>
<td>0.099</td>
<td>16.72</td>
<td>0.174</td>
<td>2.79</td>
<td>6.18</td>
<td>0.35</td>
<td>31.98</td>
<td>11.43</td>
<td>137.73</td>
<td>51.90</td>
<td>226.40</td>
<td>52.72</td>
<td>551.79</td>
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</tr>
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<td>9</td>
<td>575.83</td>
<td>0.068</td>
<td>6.82</td>
<td>0.045</td>
<td>0.40</td>
<td>1.55</td>
<td>0.40</td>
<td>8.08</td>
<td>3.19</td>
<td>44.16</td>
<td>18.20</td>
<td>88.22</td>
<td>23.10</td>
<td>277.05</td>
<td>40.10</td>
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<td>10</td>
<td>1950.77</td>
<td>0.072</td>
<td>15.31</td>
<td>0.180</td>
<td>3.69</td>
<td>8.23</td>
<td>0.58</td>
<td>40.37</td>
<td>14.26</td>
<td>171.96</td>
<td>62.45</td>
<td>262.61</td>
<td>61.12</td>
<td>620.15</td>
<td>80.57</td>
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<tr>
<td>11</td>
<td>1258.39</td>
<td>0.085</td>
<td>6.53</td>
<td>0.077</td>
<td>1.98</td>
<td>4.52</td>
<td>1.20</td>
<td>22.2</td>
<td>8.24</td>
<td>107.47</td>
<td>41.86</td>
<td>190.82</td>
<td>46.76</td>
<td>529.76</td>
<td>71.95</td>
</tr>
</tbody>
</table>
4.4 Sulfur and lead isotopes

The sulfur isotopic compositions determined during this study, as well as data from Wang Longsheng et al. (2005), are listed in Table 5 and shown in Fig. 12. In detail, the $\delta^{34}$S values range from $\sim 3.0\%$ to $1.4\%$ for pyrite, $\sim -2.3\%$ to $-1.0\%$ for sphalerite, $\sim -4.0\%$ to $-2.7\%$ for chalcopyrite, and $\sim -6.9\%$ to $-4.4\%$ for argentite. In addition, the $\delta^{34}$S value for one chalcanthite sample is $\sim 3.0\%$, and the $\delta^{34}$S value for one pyrrhotite sample is $\sim 1.9\%$. There is a trend of $\delta^{34}$S_Argentite $<$ $\delta^{34}$S_Chalcocite $<$ $\delta^{34}$S_Chalcopyrite $<$ $\delta^{34}$S_Sphalerite $<$ $\delta^{34}$S_Pyrrhotite $<$ $\delta^{34}$S_Pyrite, indicating an overall equilibrium of S-isotopes among the sulfides during the formation of the Weiquan deposit.

The lead isotope data for Weiquan sulfide minerals and associated whole rocks are listed in Table 6 and shown in Fig. 13. One whole-rock sample has $^{206}$Pb/$^{204}$Pb, $^{207}$Pb/$^{204}$Pb, and $^{208}$Pb/$^{204}$Pb ratios of 18.2060, 15.5674, and 38.0511, respectively. Twelve sulfide samples have $^{206}$Pb/$^{204}$Pb ratios that range from 17.9848 to 18.2785 (avg).

**Table 3** C–O isotope compositions (%) of the Weiquan Ag–polymetallic deposit

<table>
<thead>
<tr>
<th>Spot no.</th>
<th>Sample No.</th>
<th>Class</th>
<th>Mineral</th>
<th>$\delta^{13}$C_(CH) (%)</th>
<th>$\delta^{18}$O_CaCO_3 (%)</th>
<th>$\delta^{18}$O_SMOW (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>ZK2206-175</td>
<td>Ore</td>
<td>Calcite</td>
<td>1.4</td>
<td>$-17.5$</td>
<td>12.9</td>
</tr>
<tr>
<td>2</td>
<td>ZK1407-121</td>
<td>Barren</td>
<td>Calcite</td>
<td>1.5</td>
<td>$-18.2$</td>
<td>12.1</td>
</tr>
<tr>
<td>3</td>
<td>ZK609-513</td>
<td>Ore</td>
<td>Calcite</td>
<td>2.3</td>
<td>$-17.0$</td>
<td>13.4</td>
</tr>
<tr>
<td>4</td>
<td>D1</td>
<td>Ore</td>
<td>Calcite</td>
<td>1.9</td>
<td>$-17.4$</td>
<td>13.0</td>
</tr>
<tr>
<td>5</td>
<td>D1-2</td>
<td>Ore</td>
<td>Calcite</td>
<td>$-2.5$</td>
<td>$-21.0$</td>
<td>9.3</td>
</tr>
<tr>
<td>6</td>
<td>D2</td>
<td>Ore</td>
<td>Calcite</td>
<td>1.2</td>
<td>$-17.8$</td>
<td>12.5</td>
</tr>
<tr>
<td>7</td>
<td>D9</td>
<td>Ore</td>
<td>Calcite</td>
<td>$-1.0$</td>
<td>$-18.4$</td>
<td>12.0</td>
</tr>
</tbody>
</table>

**Table 4** H–O isotope compositions (%) of the Weiquan Ag–polymetallic deposit

<table>
<thead>
<tr>
<th>Sample no.</th>
<th>Mineralization stage</th>
<th>Mineral</th>
<th>$\delta^{18}$O_SMOW (%)</th>
<th>Average $T_1$ (°C)</th>
<th>$\delta^{18}$O_D</th>
<th>$\delta^{18}$O_SMOW</th>
</tr>
</thead>
<tbody>
<tr>
<td>ZK2206-175</td>
<td>III</td>
<td>Calcite</td>
<td>12.9</td>
<td>217</td>
<td>4.7</td>
<td>$-96.6$</td>
</tr>
<tr>
<td>ZK609-513</td>
<td>III</td>
<td>Calcite</td>
<td>13.4</td>
<td>217</td>
<td>5.2</td>
<td>$-91.1$</td>
</tr>
<tr>
<td>D1</td>
<td>III</td>
<td>Calcite</td>
<td>13.0</td>
<td>217</td>
<td>4.8</td>
<td>$-78.0$</td>
</tr>
<tr>
<td>D1-2</td>
<td>III</td>
<td>Calcite</td>
<td>9.3</td>
<td>217</td>
<td>1.1</td>
<td>$-111.7$</td>
</tr>
<tr>
<td>D1-3</td>
<td>III</td>
<td>Quartz</td>
<td>12.5</td>
<td>217</td>
<td>1.8</td>
<td>$-66.1$</td>
</tr>
<tr>
<td>D2</td>
<td>III</td>
<td>Calcite</td>
<td>12.5</td>
<td>217</td>
<td>4.3</td>
<td>$-84.4$</td>
</tr>
<tr>
<td>D9</td>
<td>III</td>
<td>Calcite</td>
<td>12.0</td>
<td>217</td>
<td>3.8</td>
<td>$-91.1$</td>
</tr>
<tr>
<td>ZK1407-121</td>
<td>IV</td>
<td>Calcite</td>
<td>12.1</td>
<td>146</td>
<td>$-0.3$</td>
<td>$-115.6$</td>
</tr>
</tbody>
</table>

$\delta^{18}$O calculations use different fractionation equations for different minerals: 1000$\delta^{18}$O_{water} = 3.38 × $10^{-3}$T2 − 3.40 (after Clayton et al., 1972); 1000$\delta^{18}$O_{calcite-water} = 2.78 × $10^{-3}$T2 − 3.39 (after O’Neil et al., 1999).
Table 5 Sulfur isotope analysis results of sulfides in the Weiquan ore deposit

<table>
<thead>
<tr>
<th>Spot no.</th>
<th>Sample no.</th>
<th>Sample</th>
<th>Mineral</th>
<th>$\delta^{34}S_{	ext{crst}}$ (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>ZK1407-123.5</td>
<td>Host rock</td>
<td>Sphalerite</td>
<td>$-1.0$</td>
</tr>
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<td>2</td>
<td>ZK1409-510.4</td>
<td>Host rock</td>
<td>Pyrite</td>
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</tr>
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<td>3</td>
<td>ZK2206-473.3</td>
<td>Ore</td>
<td>Sphalerite</td>
<td>$-2.3$</td>
</tr>
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<td>4</td>
<td>D2</td>
<td>Ore</td>
<td>Argentite</td>
<td>$-6.9$</td>
</tr>
<tr>
<td>5</td>
<td>D3</td>
<td>Ore</td>
<td>Argentite</td>
<td>$-4.4$</td>
</tr>
<tr>
<td>6</td>
<td>D5</td>
<td>Ore</td>
<td>Pyrite</td>
<td>$-1.2$</td>
</tr>
<tr>
<td>7</td>
<td>D6-1</td>
<td>Ore</td>
<td>Chalcopyrite</td>
<td>$-3.0$</td>
</tr>
<tr>
<td>8</td>
<td>D6-2</td>
<td>Ore</td>
<td>Chalcopyrite</td>
<td>$-4.0$</td>
</tr>
<tr>
<td>9</td>
<td>D10</td>
<td>Ore</td>
<td>Pyrite</td>
<td>$1.4$</td>
</tr>
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<td>11</td>
<td>WQ-18-1</td>
<td>Quartz diorite</td>
<td>Pyrite</td>
<td>$-0.6$</td>
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<td>12</td>
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<td>Quartz diorite</td>
<td>pyrrhotine</td>
<td>$-1.9$</td>
</tr>
</tbody>
</table>

Fig. 12. Histogram showing the $\delta^{34}S$ values for sulfide minerals from the Weiquan deposit.

18.0674), $^{207}$Pb/$^{206}$Pb ratios that range from 15.5188 to 15.6536 (avg. 15.5434), and $^{208}$Pb/$^{204}$Pb ratios that range from 37.8125 to 38.4650 (avg. 37.9417). As shown in Fig. 13, all the Weiquan data yield similar lead isotope compositions. The results indicate that the lead isotopic ratios are relatively uniform and that the sulfides share the same source. In addition, the calculated $\mu$, $\omega$, and Th/U values range from 9.35 to 9.58, 35.11 to 37.87, and 3.63 to 3.83, respectively, indicating that the U and Th concentrations in the lead source area are lower.

5 Discussion

5.1 Age and tectonic setting of ore formation

The precise dating of host rocks can be used to constrain the timing and duration of magmatic hydrothermal events, which is crucial to understanding the rock-forming processes and geodynamic setting (Leng et al., 2013; Deng et al., 2014; Wang et al., 2015d; Zhang et al., 2016b).

The Weiquan deposit is a skarn deposit that shows an obvious spatial relationship between mineralization and magmatic activity. Our new field and petrographic evidence support a genetic relationship between ore formation and magmatic activity, for the following reasons. First, skarn minerals such as garnet and epidote are well-developed in the deposit (Fig. 4), and these are typically the products of magmatically-hydrothermal activity. Second, the H–O isotope data for the calcites show that the early fluids were of magmatic origin, which means that the mineralization was associated with a phase of the Bailingshan granitic magmatism.

Mao et al. (2002) and Han et al. (2006) proposed that the Bailingshan granitoids provided the heat and material required for the Weiquan mineralization, and the age of a sample collected from the Bailingshan pluton, southeast of the Weiquan deposit, is considered to give the upper time limit of the mineralization. Wang Bixiang et al. (1989) reported a whole-rock Rb–Sr isochron age of 315.7 Ma for the Bailingshan pluton, and Wang Longsheng et al. (2005) obtained a U–Pb age of 297±3 Ma for zircons from the Bailingshan granite. Zhou Taofa et al. (2010) reported a U–Pb age of 317.7±3.5 Ma for zircons from the Bailingshan granodiorite, and Zhang et al. (2016) reported a similar age for the Bailingshan granodiorite of 317.7±1.8 Ma, but younger ages for the Bailingshan monzonitic granite (313.7±2.1 Ma) and granite (307.2±2.3 Ma). The variations in the ages reported in these earlier studies show that the timing of the mineralization in the Weiquan deposit remains equivocal.

In this paper we provide new LA–ICP–MS zircon U–Pb data to constrain the age of the Weiquan wall-rock tuffs and the post-ore-deposit dike of diorite porphyry. The unaltered Weiquan tuff was deposited at 326.5±4.5 Ma (Fig. 8a), and this age is consistent with the U–Pb age of 334.0±2.5 Ma for zircons from rhyolites in the Yamansu formation (Luo Ting et al., 2012) as well as the U–Pb ages of 324.4±0.94 and 323.47±0.95 Ma for zircons from basalts and skarns in the Yamansu formation (Hou et al., 2014). The diorite porphyry dike was emplaced at 298.5±1.5 Ma (Fig. 8b). Based on the geological and crosscutting relationships, the age of the tuff (ca.
298.5±1.5 Ma) represents the lower time limit of ore formation, and the age of the diorite porphyry (ca. 326.5±4.5 Ma) represents the upper time limit.

Previous studies have indicated a complex tectonic evolution for the East Tianshan orogenic belt, including subduction, accretion, and the subsequent collision of the Siberian and Tarim cratons (Ma Ruishi et al., 1993, 1997; Xiao et al., 2004, 2013; Han et al., 2006, 2010; Santosh et al., 2011; Deng et al., 2014). Studies of Carboniferous andesites and granitoids in the East Tianshan have shown that they display a subduction-related component (Sun et al., 2008; Tang et al., 2010; Wang Yinhong et al., 2014). The high-pressure–ultrahigh-pressure (HP–UHP) metamorphic belt of the Chinese West Tianshan was formed as a result of subduction of the Tarim plate beneath the Yili–Central Tianshan plate during the late Palaeozoic (Su et al., 2010; Fan Tianfeng et al., 2015; Lü Zeng and Zhang Lifei, 2016). In addition, a new U–Pb age of ca. 320 Ma for zircons from the South Tianshan eclogites indicates that the collision of the Tarim Craton and the Yili–Central Tianshan Block occurred during the late Carboniferous (Su et al., 2010). Furthermore, the presence of ophiolite (ca. 310 Ma) and widespread bimodal volcanic rocks (ca. 290 Ma; Qin Kezhang et al., 2002; Qin et al., 2003, 2011; Zhang et al., 2015; Wang and Zhang, 2016c) indicate the collision occurred at ca. 310–290 Ma, followed by early Permian post-collisional tectonics. A number of Cu–Au, Pb–Zn, and other base metal deposits, along with associated magmatic rocks, were formed during the period from subduction to post-collisional tectonics. We conclude, therefore, that the Weinquan deposit was formed in the late Carboniferous (326.5–298.5 Ma) during a period that spanned the subduction to post-collisional tectonic events.

5.2 Source of ore-forming fluids

5.2.1 C–O isotopes

Different geological materials are characterized by different $\delta^{13}$C$_{PDB}$ values. For example, organic matter has an average $\delta^{13}$C$_{PDB}$ value of −25‰ (Hoeft, 2009), atmospheric CO$_2$ has −8‰ $\delta^{13}$C$_{PDB}$ (Schidlowski and Tiedt, 1998) or −11‰ to −7‰ $\delta^{13}$C$_{PDB}$ (Hoeft, 1997), soluble CO$_2$ in fresh water has −20‰ to −9‰ $\delta^{13}$C$_{PDB}$ (Hoeft, 1997), the continental crust has −7‰ (Faure, 1986), the mantle −7‰ to −5‰ (Hoeft, 2009), and igneous rocks −30‰ to −3‰ $\delta^{13}$C$_{PDB}$ (Hoeft, 1997). The $\delta^{13}$C$_{PDB}$ values for calcites in the Weinquan deposit show a broad range from −2.5‰ to +2.3‰, but are most commonly between 1.2‰ and 1.9‰, which is slightly higher than the $\delta^{13}$C$_{PDB}$ value for marine carbonate (about 0‰; Hoeft, 2009).

The calcites in the Weinquan ores and barren veins have similar C and O isotopic compositions, indicating a common origin. On the $\delta^{13}$C$_{PDB}$ vs. $\delta^{18}$O$_{SMOW}$ diagram, one calcite sample plots within the granite field (Fig. 10), indicating the carbon in the early ore-forming fluids was provided by magmatic water. The other calcite samples plot in the field between mantle and marine carbonates, but close to the mantle value, indicating that the CO$_2$ or CO$_2$ in the ore-forming fluids was sourced mainly from a magmatic system. In addition, the samples show a positive correlation between $\delta^{13}$C$_{PDB}$ and $\delta^{18}$O$_{SMOW}$ (Fig. 10), which is observed in both magmatic and hydrothermally deposited carbonates (Comin-Chiaramonte et al., 2005; Marks et al., 2009; Gwalani et al., 2010). We infer, therefore, that the C-bearing portion of the ore-forming fluids in the Weinquan deposit was derived from deep sources and was of magmatic origin. This further supports the notion that the ore-forming fluids of the Weinquan deposit had a mantle connection. Considering the samples also show clear signs of sedimentary hybridization, we infer that the ore-forming fluids were weakly contaminated by sedimentary rock. The carbon-bearing hydrothermal fluids of magmatic origin underwent fluid–rock reactions with carbonaceous strata during their upwards flow, thus forming the ore-forming fluids of the early and middle stages.

5.2.2 H–O isotopes

A study of hydrogen (H) and oxygen (O) isotopes is a common and effective method for constraining the source and evolution of ore-forming fluids. The $\delta^{18}$O values of calcite in the Weinquan deposit range from 9.3‰ to 13.4‰, with an average of 12.2‰ (Table 4). This range is largely consistent with the $\delta^{18}$O values of granites formed by the melting of crust (10.0‰–12.0‰; Chen Yanjing et al., 2006b), indicating that the formation of hydrothermal calcite in the Weinquan deposit was related to magmatic fluids and associated intrusive rocks. The $\delta^2$D values of the ore-forming fluids of the middle stage range from −111.7‰ to −66.1‰, and the $\delta^{18}$O$_{H_2O}$ values range from 1.1‰ to 5.2‰. The $\delta^2$D value of the late stage hydrothermal fluids is −115.6‰, and the $\delta^{18}$O$_{H_2O}$ value is −0.3‰. As shown in Fig. 11, the $\delta^{18}$O$_{H_2O}$ and $\delta^2$D values of the fluids from the main ore stage at Weinquan show a relatively narrow range of $\delta^{18}$O$_{H_2O}$ values and a relatively broad range of $\delta^2$D values, and plot near the magmatic water field, which suggests that magmatic fluids made a significant contribution to the ore-forming fluids.

However, compared with typical magmatic water (−40‰ to −80‰; Taylor, 1997), the $\delta^2$D values of the fluids involved in the Weinquan deposit are much lower (Fig. 11). Previous researchers have variously ascribed depleted $\delta^2$D isotopic features to interactions between
water and organic matter (Sheppard, 1986; Munoz et al., 1994; Polya et al., 2000), the influence of altitude on meteoric water distal to the site of ore deposition (Wilkinson et al., 1995; Donoghue et al., 2010), or the boiling of hydrothermal fluids (Imai et al., 1998; Gleeson et al., 1999; Fifearek and Rye, 2005). Moreover, a uniform population of $\delta D$ data may be explained by the degassing of a magma chamber (Rye, 1993). The contribution of meteoric water to the low $\delta D$ values of our samples ($-115.6\%$ to $-66.1\%$) can be ignored, because the value of modern meteoric water in the Tianshan region ($-70\%$; Chen et al., 2012a) is even higher, but it is possible that these low $\delta D$ values resulted from Raleigh fractionation during the progressive loss of the vapor phase as boiling fluids ascended towards the surface.

One sample from the barren late calcite stage yielded a remarkably low value, and plots to the right of the meteoric water line (Fig. 11), indicating that some meteoric water was incorporated into the ore-forming system during Stage IV. On the $\delta D$ vs. $\delta^{18}O$ diagram, most of the isotopic values for calcite from Stages III to IV show a decreasing trend, and the evolutionary trend of the $\delta^{18}O$ values indicates that the contribution of magmatic fluids was gradually reduced from the early to late stages of mineralization, also implying the involvement of meteoric water.

5.3 Origin of the ore-forming constituents

5.3.1 S isotopes

The $\delta^{34}S$ values of coexisting fluids can be calculated from the $\delta^{34}S$ values of sulfides, on the basis of available fractionation factors between aqueous sulfur species and sulfide minerals, which are closely related to the temperature and $fO_2$ during sulfide precipitation (Ohmoto, 1972; Ohmoto and Rye, 1979). Considering that some pyrite samples are intergrown with minor magnetite, the sulfur isotopic fractionation between sulfides and fluids would probably have been minor (<2%; Ohmoto and Rye, 1979), which means that the measured $\delta^{34}S$ values of sulfides at Weiquan are nearly equal to those of the fluids.

Many researchers are of the opinion that the sulfur in skarn deposits is of magmatic origin (e.g., Ault, 2004; Kamvong and Zaw, 2009), but some researchers have argued for the derivation of the sulfur from the host carbonate rocks (e.g., Ishihara et al., 2000, 2002). The $\delta^{34}S$ values of sulfides in the Weiquan deposit exhibit a large range (−6.9% to 1.4%; mean=−2.2%, n=13), and these values are similar to but a little more negative than the $\delta^{34}S$ values of sulfides from the majority of magmatic hydrothermal deposits (−3% to +1%; Hoefs, 2009). The mixing of different sources or the oxidation effect of a reduced fluid may explain a large range of $\delta^{34}S$ values (Liu et al., 2007; Deng et al., 2015). In the case of the Weiquan deposit, we suggest that the large range of $\delta^{34}S$ may be attributed to mixed sources, and a sedimentary contribution, for example, is suggested by the three relatively low $\delta^{34}S$ values of −6.9%, −4.4%, and −4.0%.

We propose, therefore, that the sources of sulfur in the Weiquan deposit were most likely magmatic, but possibly with contributions from sediments.

5.3.2 Pb isotopes

Pb isotope data are useful geochemical tracers and can indicate crustal evolution, the source of Pb, and deposit genesis (Lu Zhicheng et al., 2000). The Pb isotopic compositions of the sulfides from Weiquan show a generally narrow range, with $^{206}Pb/^{208}Pb$ ratios of 17.9848 to 18.2785, $^{207}Pb/^{206}Pb$ ratios of 15.5188 to 15.6536, and $^{208}Pb/^{206}Pb$ ratios of 38.7125 to 38.4650 (Table 6). These values indicate that the sulfides share similar Pb isotopic compositions, and they might therefore have the same source of Pb. The Pb isotopic compositions of the host rock are relatively consistent with those of the sulfides at Weiquan, indicating a somewhat uniform lead source, and this indicates that the host rock was probably an important source of ore-forming fluids as a result of fluid–rock reactions. The majority of the data for the Weiquan deposit are distributed in the domain between the orogenic belt and mantle evolution lines on the $^{207}Pb/^{206}Pb$ vs. $^{206}Pb/^{204}Pb$ diagram (Fig. 13a), but they are near the upper

<table>
<thead>
<tr>
<th>Spot no.</th>
<th>Sample no.</th>
<th>Sample name</th>
<th>Sulfide/Rock</th>
<th>$^{206}Pb/^{208}Pb$</th>
<th>$^{207}Pb/^{206}Pb$</th>
<th>$^{208}Pb/^{206}Pb$</th>
<th>$\mu$</th>
<th>$\omega$</th>
<th>Th/U</th>
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<tr>
<td>1</td>
<td>ZK1407-123.5a</td>
<td>Host rock</td>
<td>Pyrite</td>
<td>18.1431</td>
<td>15.5302</td>
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<td>35.11</td>
<td>3.65</td>
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<tr>
<td>2</td>
<td>ZK1407-123.5b</td>
<td>Host rock</td>
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<td>18.0564</td>
<td>15.5248</td>
<td>37.8572</td>
<td>9.35</td>
<td>35.34</td>
<td>3.66</td>
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<tr>
<td>3</td>
<td>ZK1409-510.40</td>
<td>Host rock</td>
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<td>15.5188</td>
<td>37.8339</td>
<td>9.35</td>
<td>35.47</td>
<td>3.67</td>
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<tr>
<td>4</td>
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<td>15.5343</td>
<td>37.8905</td>
<td>9.37</td>
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<tr>
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<td>Sphalerite</td>
<td>18.0675</td>
<td>15.5264</td>
<td>37.8695</td>
<td>9.36</td>
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<td>3.66</td>
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<td>15.5279</td>
<td>37.8723</td>
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<td>35.52</td>
<td>3.67</td>
</tr>
<tr>
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<td>15.5675</td>
<td>38.0666</td>
<td>9.44</td>
<td>36.66</td>
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</tr>
<tr>
<td>8</td>
<td>D6</td>
<td>Ore</td>
<td>Chalcopyrite</td>
<td>17.9848</td>
<td>15.5235</td>
<td>37.8125</td>
<td>9.36</td>
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<tr>
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<td>15.5500</td>
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<td>35.51</td>
<td>3.67</td>
</tr>
<tr>
<td>13</td>
<td>D1049</td>
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<td>Rock</td>
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<td>15.5674</td>
<td>38.0511</td>
<td>9.42</td>
<td>35.73</td>
<td>3.67</td>
</tr>
</tbody>
</table>
crust growth curve on the $^{208}\text{Pb}^{206}\text{Pb}$ vs. $^{206}\text{Pb}^{204}\text{Pb}$ diagram (Fig. 13b).

Furthermore, Pb from the mantle has an average $\mu$ value of 8.92 and an average $\omega$ value of 31.844, while Pb from the upper crust has an average $\mu$ value of 9.58 and an average $\omega$ value of 41.860 (Doe and Zartman, 1979). All of the sulfides from Weiquan are characterized by moderate radiogenic Pb isotope values with $\mu$ values of 9.35 to 9.58 and $\omega$ values of 35.11 to 37.87, and these are higher than those of the mantle but lower than those of the upper crust, indicating that the source of Pb was mixed. We suggest, therefore, that the lead was derived from a deep-seated magma, but that some crustal material was involved.

5.4 Ore genesis

The Aqishan–Yamansu arc belt contains early Carboniferous basalts, andesites, dacites, and tuffs of the Yamansu Formation and lower Carboniferous rhyolites of the Tugutubulake Formation (Ma Ruishi et al., 1993). It contains Fe-(Cu) and Cu–Ag–Pb–Zn skarn deposits, including Hongyuantan, Bailingshan, Duotoushan, Weiquan, and Yamansu. These deposits formed during the emplacement of granitic intrusions and associated hydrothermal activity, and they replaced Carboniferous and Neoproterozoic carbonate and calcareous elastic rocks (Mao et al., 2005). Previous researchers have suggested that the syn-arc and post-arc granitic magmatism of the Aqishan–Yamansu arc belt occurred in three episodes during the period from ca. 349 to 228 Ma (Zhou Taofa et al., 2010; Lei Ruxiong et al., 2013): Carboniferous (ca. 349–303 Ma), early Permian (ca. 297–272 Ma), and middle to late Triassic (ca. 246–228 Ma). There is a general eastwards younging trend in this magmatic activity. Geo-dynamic studies indicate that the late Paleozoic granitoids are arc-related, whereas the Triassic granitoids were emplaced in a post-collisional setting (Zhou Taofa et al., 2010; Lei Ruxiong et al., 2013). The Carboniferous granitoids occur mainly in the western part of the Aqishan–Yamansu arc belt, including the Xifengshan, Changtiaoshan, Hongyuantan, and Bailingshan plutons of granite and granodiorite. The early Permian plutons are distributed mainly in the middle part of the arc belt and are represented by the Longdong monzonite and Duotoushan granites. The middle to late Triassic granitoids are distributed mainly in the eastern part of the belt and include the Tudun and Yamansubei granites (Zhang et al., 2016).

The results of our zircon U–Pb isotope dating suggest that the Weiquan deposit was formed in the period 326.5–298.5 Ma and that it was related to the late Carboniferous granitic magmatism of the Aqishan–Yamansu arc belt. Considering the tectonic evolution of the East Tianshan, we conclude that the gold mineralization in the Weiquan area was formed in a tectonic setting spanning a period from subduction to post-collision.

Feng Jing et al. (2008) showed that the host rocks of the Weiquan deposit contain higher than normal Ag and Cu concentrations, which suggests the volcanic–pyroclastic rocks contributed a portion of the ore-forming metals in the Weiquan deposit. Taking into account our C–H–O–S–Pb isotope data, we conclude that the metals of the Weiquan ore field were derived mainly from the adjacent granodiorite intrusions, but that there was an additional contribution from the host rocks, and that the ore-forming fluids were dominated by magmatic–hydrothermal fluids mixed with smaller amounts of meteoric fluid.

Based on the results described above and the regional tectonic background, we propose the following model (Fig. 14) for the overall metallogenic of the Weiquan

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![Fig. 13. Diagram showing the lead isotopic compositions of the Weiquan ore deposit.](image-url)  
(a) $^{208}\text{Pb}^{206}\text{Pb}$ vs. $^{206}\text{Pb}^{204}\text{Pb}$ and (b) $^{208}\text{Pb}^{206}\text{Pb}$ vs. $^{206}\text{Pb}^{204}\text{Pb}$ (base map from Zartman and Doe, 1981).
(2) The H–O isotope systematics suggest that the ore-forming fluids of the Weiquan deposit were derived from magmatic sources that were mixed with some meteoric water in the later stages. The δ18O values of the calcite suggest that the CO3^2– or CO2 in the ore-forming fluids was sourced mainly from a magmatic system, but with some contamination by sedimentary rocks. The δ34S values of sulfides indicate that the sulfur in the Weiquan deposit was derived primarily from magmatic sources and partly from the host sediments. Lead isotopic compositions show that the metallic elements of the ores came from a mixed crust–mantle source.

(3) Petrographic observations, geological features, and C–H–O–S–Pb isotope systematics indicate that the Weiquan deposit is a skarn deposit. The mineralization was related to a magmatic–hydrothermal system, and the skarns and ores were formed by fluids that were derived from a deep-seated active magma chamber with the involvement of smaller amounts of meteoric water and contributions from the host volcano-sedimentary rocks of the Yamansu Formation. The existence of the Weiquan deposit indicates that the Aqishan–Yamansu arc belt in East Tianshan has considerable potential for further exploration for skarn deposits.

6 Conclusions

(1) The Weiquan deposit is a Ag–polymetallic deposit in East Tianshan that is hosted by lower Carboniferous volcano-sedimentary rocks of the Yamansu Formation. The LA–ICP–MS zircon U–Pb age data show that the Weiquan deposit was formed in the period 326.5–298.5 Ma.

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References


Hoefs, J., 2009. *Stable Isotope Geochemistry* (sixth ed.). Berlin,
Heidelberg: Springer Verlag, 1–285.


Shen, P., Pan, H.D., Zhou, T.F., and Wang, J.B., 2014b. Petrography, geochemistry and geochronology of the host porphyries and associated alteration at the Tuwu Cu deposit, NW China: A case for increased depositional efficiency by


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