Geological Characteristics and Genesis of the Jiamoshan MVT Pb–Zn Deposit in the Sanjiang belt, Tibetan Plateau

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Abstract: The carbonate-hosted Pb–Zn deposits in the Sanjiang metallogenic belt on the Tibetan Plateau are typical of MVT Pb–Zn deposits that form in thrust-fold belts. The Jiamoshan Pb–Zn deposit is located in the Changdu area in the middle part of the belt, and it represents a new style of MVT deposit that was controlled by karst structures in a thrust-fold system. Such a karst-controlled MVT Pb–Zn deposit in thrust settings has not previously been described in detail, and we therefore mapped the geology of the deposit and undertook a detailed study of its genesis. The karst structures that host the Jiamoshan deposit were formed in Triassic limestones along secondary reverse faults, and the orebodies have irregular tubular shapes. The main sulfide minerals are galena, sphalerite, and pyrite that occur in massive and lamellar form. The ore-forming fluids belonged to a Mg2+-Na–K2SO4–Cl–F–NO3–H2O system at low temperatures (120–130°C) but with high salinities (19–22% NaCl eq.). We have recognized basal brine as the source of the ore-forming fluids on the basis of their H−O isotopic compositions (−145‰ to −93‰ for δDvSMOW and −2.2‰ to 13.00‰ for δ18OvH2O), the ratios of Cl/Br (14−1196) and Na/Br (16−586) in the hydrothermal fluids, and the C−O isotopic compositions of calcite (−5.0‰ to 3.7‰ for δ13CvPDB and 15.1‰ to 22.3‰ for δ18OvPDB). These fluids may have been derived from evaporated seawater trapped in marine strata at depth or from Paleogene–Neogene basins on the surface. The δ34S values are low in the galena (−3.2‰ to 0.6‰) but high in the barite (27.1‰), indicating that the reduced sulfur came from gypsum in the regional Cenozoic basins and from sulfates in trapped paleo-seawater by bacterial sulfate reduction. The Pb isotopic compositions of the galena samples (18.3270–18.3482 for 206Pb/204Pb, 15.6345–15.6390 for 207Pb/204Pb, and 38.5503–38.5582 for 208Pb/204Pb) are similar to those of the regional Triassic volcanic–arc rocks that formed during the closure of the Paleo-Tethys, indicating these arc rocks were the source of the metals in the deposit. Taking into account our new observations and data, as well as regional Pb–Zn metallogenic processes, we present here a new model for MVT deposits controlled by karst structures in thrust-fold systems.

Key words: ore genesis, thrust-fold belt, karst-controlled style, Jiamoshan Pb-Zn deposit, MVT (Mississippi Valley Type) Pb-Zn deposit, Sanjiang metallogenic belt, Tibetan Plateau

I Introduction

The Sanjiang metallogenic belt is located along the northeastern margin of the Tibetan Plateau between the Jinhaijiang and Bangonghu–Nuijiang suturets (Fig. 1a), and it is one of the most important polymetallic metallogenic belts in China (Hou et al., 2007; Deng et al., 2014). Since the Cenozoic continent–continent collision between India and Eurasia, ore-deposits of three types of metallogenic system have been formed in this belt, namely, porphyry Cu–Mo–Au deposits (Hou et al., 2003, 2006), orogenetic Au deposits (Sun et al., 2009; Deng et al., 2015), and sediment-hosted Pb–Zn deposits (Hou et al., 2008; Song et al., 2011). Deposits that belong to sediment-hosted Pb–Zn metallogenic systems are located in six areas (Fig. 1a) from the Simao Basin in the southeast (e.g., Li S et al., 2013; Wei et al., 2015) to the Tianshuihai area in the northwest of the Sanjiang belt (Xie et al., 2014; Dong et al., 2015; Gao et al., 2020) and form a huge Pb–Zn metallogenic belt that is 2000 km in length. The Pb–Zn ores in this belt are as large as 640 Mt with 0.76–3.2% Pb and 0.3–6.1% Zn (Liu et al., 2017 and references therein) and show great economic value and scientific research potential. These deposits form a set of Mississippi Valley Type (MVT) Pb–Zn deposits developed in a thrust-fold belt (Hou et al., 2008; Liu et al., 2016, 2017; Zhang H R et al., 2017), and based on various ore-controlling factors and mineralization characteristics, they were categorized into three styles: a thrust-related fractures and fold-controlled style (Liu et al., 2011, 2015; Wang et al., 2018), a facies transition-controlled style (Liu et al., 2013), and an evaporite dome-controlled style (Leach et al., 2017).
Fig. 1. Geological maps of the Changdu area in the Sanjiang belt, Tibetan Plateau.

(a) Tectonic position of the Changdu region after Spurlin et al. (2005) and Yin and Harrison (2000); (b) Map of regional structures showing the distribution of Pb–Zn deposits, modified from TARGS (2006); (c) Regional cross-section, modified from TARGS (2006).

Carbonate-hosted Pb–Zn deposits in Fig. 1a: 1, Huoshaoyun; 2, Sachakou; 3, Chaqupacha; 4, Nabaozhalong; 5, Dongmozhazhua; 6, Mohailaheng; 7, Jiamoshan; 8, Zhaofayong; 9, Lalongla; 10, Huachangshan; 11, Jinding; 12, Changdong; 13, Luoboshan.
However, for such a huge Pb–Zn belt, further research is needed to understand the regional variations and patterns in mineralization and to develop a more comprehensive metallogenic model for these Pb–Zn deposits in thrust–fold belts. Zhang et al. (2012) discovered that the mineralization of the Chaqupacha Pb–Zn deposit in the Tuotuohe area in the northern part of the Sanjiang belt (Fig. 1a) was controlled by karst caves. Subsequently, we found a similar interpretation could be applied to the Pb–Zn deposits in the Changdu area in the middle part of the Sanjiang belt (Fig. 1a), and also to deposits in the Simao Basin in the southern part of the Sanjiang belt (Fig. 1a). It came to be realized, therefore, that a new style of Pb–Zn deposits controlled by karst structures was well represented in this belt. The Jiamoshan Pb–Zn deposit which is located in the Changdu area (Fig. 1a, b) was first found around the year of 2006 and did not receive detail research before. However, after the geological reconnaissance in this deposit, we found it was the most suitable karst-controlled deposit for the geological research in the Sanjiang belt for its systematic adit engineering. In this context, we selected this deposit for a project that would involve detailed geological mapping, thin section observations, and geochemical analyses to determine the genesis of this karst-controlled Pb–Zn deposit. The systematic information obtained would lead to a better understanding of MVT Pb–Zn deposits in the thrust–fold belt generally and of the regional patterns in Pb–Zn metallogenesis.

2 Regional Geology

The Changdu area which hosts the Jiamoshan deposit is in the eastern part of the Qiangtang Terrane (Fig. 1a). The lithology in this area includes marine and volcanic sedimentary rocks of Devonian to Carboniferous and terrestrial sedimentary rocks of Cretaceous to Neogene (RGSS, 1993; TARGS, 2006). Older metamorphic rocks of Proterozoic age underlie these sedimentary sequences (Fig. 1b). The Changdu area is located where the Sanjiang belt changes in trend from NNE–SSW to WNW–ESE (Fig. 1a), and the rocks in the area underwent strong tectonic deformation during the Mesozoic–Cenozoic collisional orogeny. Regional-scale nappes and strike-slip faults formed during the India–Eurasia collision in the Cenozoic and now constitute the most remarkable deformation structures of the region (Tang et al., 2006; Liu et al., 2016). The major nappes consist of two thrust systems with opposite dips, known as the eastern and western thrust systems (Fig. 1c). The Changdu Basin is located between the two sets of reverse faults (Fig. 1c). Lithology involved in the thrust structures includes Proterozoic to Lower Ordovician metamorphic basement rocks and upper Paleozoic to Mesozoic clastic rocks (Tang et al., 2006). Strike-slip faults such as the Gongjue–Mangkang fault and strike-slip pull-apart basins such as the Gongjue and Mangzong basins formed during the period of post-thrusting extension, and now occur together with the regional thrust faults (Fig. 1a). The most typical carbonate-hosted cluster of Pb–Zn deposits within the Changdu area is the Zhaofayong cluster and it is hosted by reverse faults that belong to the Lancangjiang thrust fault set (Fig. 1b). This ore cluster, from south to north, contains three carbonate-hosted Pb–Zn deposits which are the Lalongla, Zhaofayong, and Jiamoshan deposits (Fig. 1b).

3 Deposit Geology

The Jiamoshan Pb–Zn deposit, which has reserves of Pb + Zn ore of 1 Mt with 10.40–13.40% Zn and 15.70–18.60% Pb (Liu et al., 2016), is located in the northern part of the Zhaofayong ore cluster (Fig. 1b). The mining area of Jiamoshan is dominated by three sedimentary sequences (Fig. 2) known as the Bolila, Adula, and Dongdaqiao formations. The Late Triassic Bolila Formation is at the base of the stratigraphic column and
consists of limestones that contain crinoids and shell fossils and chert nodules. It crops out over 75% of the mine area and hosts the orebodies (Fig. 2). The Upper Triassic Adula Formation overlies the Bolila Formation and consists of sandstones, siltstones, and mudstones (Fig. 2). It crops out in the southeastern part of the mine area. The Middle Jurassic Dongdaqiao Formation is the youngest of the three sedimentary sequences, and it consists of purple–red feldspar- and quartz-bearing sandstones and mudstones. It is in thrust contact with the underlying Triassic strata via F₁ fault and crops out in the southwestern part of the mine area (Fig. 2).

Two large reverse faults (F₁ and F₂) with similar strikes (310°) and dips (55°–75°) define the structural fabric of the area (Fig. 2). In addition, there are three other sets of faults in the area. The first and second sets are reverse faults that strike NE–SW and dip towards the NW, and reverse faults that strike NW–SE and dip to the NE (Fig. 2). Both sets of faults dip at angles larger than 60°, and they constitute a system of conjugate faults that formed at the same time as faults of F₁ and F₂. The third set consists of ill-defined faults that strike WNW–ESE with NNE and SSW dip directions (Fig. 2). Karst caves are numerous in the limestones of the Bolila Formation, and most are found between the F₁ and F₂ faults (Fig. 2) and along the NE–SW and NW–SE striking faults (Fig. 3). The karst caves on the surface are empty with just some residual iron oxides (Fig. 4a), but the caves underground are filled completely or partly with Pb–Zn ores (Fig. 4b–d).

All the Pb–Zn orebodies were deposited in karst caves (Fig. 3), making up two major groups of underground orebodies that have been explored by adit prospecting (Fig. 2). The orebodies are irregular tubular shapes (Fig. 3), range in size from 1 m × 1 m to 20 m × 10 m across, and can extend downwards as far as 20 to 30 m. The different orebodies do not connect with each other (Fig. 3). An abundance of breccias with sub angular limestone blocks occurs in the orebodies and the blocks measure from several centimeters to several meters across (Fig. 4c). These breccias are ore-barren breccias and the cements consist of sulfides and calcite (Fig. 4d).

The sulfide minerals in the deposit are simple galena, sphalerite, and pyrite, but a wide variety of non-sulfide minerals produced by the hydrothermal system includes calcite, dolomite, and barite, and secondary oxidation produced smithsonite, cerusite, anglesite, siderite, gypsum, willemite, hydrozincite, and limonite. The ores are massive (Fig. 5a, b) and lamellar (Fig. 5c, d), and both types form the cement of the big breccias in the mineralized karst caves (Figs. 4c, 5e). The massive ores are made up of galena (Fig. 5a, b, d) and sphalerite (Fig. 5d) along with cogenetic calcite (Fig. 5a). The lamellar ores are made up of residual galena and sphalerite along with secondary Pb–Zn–Fe oxidized minerals (Fig. 5d). Two mineralization events can be recognized in the deposit (Fig. 6), and each event had a hydrothermal stage and an oxidation stage. In mineralization event I (Fig. 6), coarse euhedral crystals of hydrothermal minerals (calcite, galena, and sphalerite) were precipitated to fill open spaces (Fig. 5a, b, e) among the breccia blocks during stage I, which is most economically important. The sulfides of this stage underwent secondary oxidation to
produce stage II smithsonite, cerusite, anglesite, and siderite (Fig. 5e). In mineralization event II (Fig. 6), fine-grained and crustiform hydrothermal galena and sphalerite in stage III grew over the minerals formed during mineralization event I (Fig. 5f). There were two hydrothermal sub-stages (III₁ and III₂) (Fig. 5g, h) in stage III and both underwent a later period of secondary oxidation (stage IV) (Fig. 6). The secondary Pb–Zn minerals including smithsonite, cerusite, and anglesite all contribute to the Pb–Zn ore resource in the deposit.

4 Sampling and Analytical Methods

Ore and gangue samples were collected from both the surface and trench in the Jiamoshan deposit. Polished thin sections were prepared from the hand specimens for petrographic observation. Calcite grains in double-sided polished thin sections were prepared for fluid inclusion studies. Single mineral grains were separated and handpicked from fresh samples for geochemical analyses including fluid inclusion electrolyte, and carbon, oxygen, hydrogen, sulfur and lead isotopes.

The calcite fluid inclusions that formed during mineralization event II (Fig. 6) were too small to be tested under the microscope, so systematic microthermometric analyses were performed only on fluid inclusions in stage I calcite precipitated during mineralization event I (Fig. 6). The double-polished wafers of calcite were placed on a Linkam THMSG-600 heating–freezing stage mounted on a Leica DM LB microscope equipped with a Nikon 50× long working-distance lens at the Institute of Geology, Chinese Academy of Geological Sciences, Beijing, China. The stage was calibrated for temperatures between −56.6 and 0.0 °C by guide fluid inclusions hosted in quartz before measurements. The precision for ice melting-temperatures is within ±0.1°C and for homogenization temperatures is within ±1°C. The homogenization temperatures were tested before the ice-melting temperatures in order to avoid fluid-inclusion stretching associated with ice formation, which would have increased vapor/liquid ratios and yielded erroneously high homogenization temperatures. In addition, homogenization temperatures were measured in the order of progressively higher vapor/liquid ratios so as to avoid stretching lower-temperature fluid inclusions before they were measured.

The bulk gas compositions of fluid inclusions in stage I calcite and barite of mineralization event I and sub-stage
Fig. 5. Typical ore types, minerals, textures, and structures in the Jiamoshan Pb-Zn deposit.
III$_2$ calcite of mineralization event II were analyzed using a Shimadzu GC-2010 and SEG Heater in the Key Laboratory of Metallogeny and Resource Assessments, Institute of Mineral Resources, Chinese Academy of Geological Sciences, Beijing, China. The calcite and barite samples were crushed to grain sizes between 0.30 and 0.45 mm, and then immersed in aqua regia at 70 to 80°C for 4 hours and then leached 20 times by doubly distilled water in an ultrasonic cleaning bath. The leached samples were dried for 5 hours at 70 to 80°C. Decrepitation analyses were performed between 100 and 500°C. The carrier gas (He) and the other standard gases were all from the National Standard Material Center of China. The nonaqueous volatiles in fluid inclusion vapor phases within stage I calcite were checked using Laser Raman microprobe analyses in the Beijing Research Institute of Uranium Geology, Beijing, China. This work used double-sided polished wafers and a HR800 micro-Raman spectrometer equipped with a holographic notch filter and a 100× lens. The analyses were undertaken with a 633 nm excitation line produced by an Ar$^+$ laser. The Raman intensities were measured using a charge-coupled device detector from 100 to 4300 cm$^{-1}$ with an integration time of 30 s per analysis, a laser spot diameter of 1 μm, and a spectrometer resolution of 0.14 cm$^{-1}$.

The ionic compositions of fluid inclusions in the stage I and sub-stage III$_2$ calcites were also analyzed using an ion chromatograph HIC-SP, manufactured by Shimadzu Corporation, in the same laboratory noted above. The crushed calcite grains were cleaned, dried, decrepitated at 500 °C in a vacuum, and then leached repeatedly using Millipore water in an ultrasonic cleaning bath at room temperature until the electrical conductivity of the leachate was the same as the Millipore water. All collected leachate was volumetrically fixed to 30 ml and then analyzed. The detection limits of the cationic and anionic compositional analyses were ppm and ppb, respectively. The analytical methods we used have been described in detail by Yang et al. (2007).

The carbon, oxygen, and hydrogen isotopes of the calcites from the two mineralization events (Fig. 6) were all analyzed at the Laboratory of Isotope Geology, Ministry of Land and Resources of China, Beijing, China, using a Finnigan MAT-251 mass spectrometer. The carbon and oxygen isotope analyses were undertaken on CO$_2$ generated by 100% orthophosphoric acid and using the approach of McCrea (1950). The hydrogen isotopic compositions of fluid inclusions in the calcites were determined from H$_2$ collected during the decrepitation of carbonate under a vacuum at 500°C before reduction by Zn to obtain H$_2$ at 400°C. Analytical reproducibility was better than ±0.2% for δ$^{18}$O and δ$^{13}$C, and ±2% for δD. Isotope data for O and H are reported relative to Vienna SMOW (Standard Mean Ocean Water) and for C relative to Vienna PDB (Pee Dee Belemnite).

The sulfur isotopes of galena related to both mineralization events (Fig. 6) and for barite that formed during mineralization event I (Fig. 6) were analyzed by using the same mass spectrometer at the same laboratory noted above. The SO$_2$ was prepared from galena by using Cu$_2$O or from purified BaSO$_4$ by using V$_2$O$_5$. Analytical reproducibility was better than ±0.2% for δ$^{34}$S. The data are reported relative to V-CDT.

The Pb isotopes of the galena formed in both mineralization events (Fig. 6) were analyzed in the same laboratory for sulfur analysis by using a British Nu Plasma HR MC–ICPMS. The analytical procedures involved dissolution of samples using HF and HClO$_4$ in crucibles, followed by the use of AG1-X8 basic anion exchange resin to purify Pb. The analytical methods we used have been described in detail by He et al. (2005). The analytical results for the standard NBS 981 were $^{208}$Pb/$^{206}$Pb = 2.16736 ± 0.00066, $^{207}$Pb/$^{206}$Pb = 0.91488 ± 0.00028 (+2σ), $^{206}$Pb/$^{204}$Pb = 16.9386 ± 0.0131 (+2σ), $^{207}$Pb/$^{204}$Pb = 15.4968 ± 0.0107 (+2σ), and $^{208}$Pb/$^{204}$Pb = 36.7119 ± 0.0331 (+2σ), in agreement with the reference values.

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Coarse sulfide stage (I)</th>
<th>Oxidation stage (II)</th>
<th>Fine sulfide stage (III)</th>
<th>Oxidation stage (IV)</th>
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<td>Galena</td>
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<td>Sphalerite</td>
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<td>Pyrite</td>
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<td>Calcite</td>
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<td>Dolomite</td>
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<td>Barite</td>
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<td>Smithsonite</td>
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<td>Cerusite</td>
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<td>Anglesite</td>
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<td>Siderite</td>
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<td>Gypsum</td>
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<td>Willemite</td>
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<tr>
<td>Hydrozincite</td>
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<td>Limonite</td>
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The mineral content from more to less.

Fig. 6. Paragenetic sequence of minerals in the Jiamoshan Pb–Zn deposit.
5 Results

5.1 Fluid inclusions
Fluid inclusions in the stage I calcite are rare. However, those found were easy to analyze. They formed in the growth zones of the calcite crystals (Fig. 7a) and were irregular or rectangular in shape with major axes 4 to 13 μm long (Fig. 7b). The inclusions contained two phases at room temperature, an aqueous liquid (L) and some kind of carbonic vapor phase (V), with volume ratio values of V/(V + L) = 0.01−0.05 (Fig. 7b).

5.1.1 Results of the microthermometric analyses
We undertook microthermometric analyses of 63 stage I calcite-hosted fluid inclusions for this study (Table 1). The salinities (Table 1) were calculated from ice-melting temperatures (Tm,ice) using the equations of Bodnar (1993) and Chi and Ni (2007) for two-phase fluid inclusions. The low eutectic melting temperatures (Tem) (ranging between −43.4 and −42.0°C) obtained from two fluid inclusions hosted in two calcite samples meant that densities (Table 1) are in the range of 0.83–1.04 g/cm3 from Liu and Shen (1999). The densities obtained are in the range of 273 to 1107 m (Table 1).

The analytical results of the carbon and oxygen isotopes in stage I calcite of mineralization event I and sub-stage III2 of mineralization event II are summarized in Table 4. Therefore, just the effects of fluid pressures were considered when the ore-forming pressures and depths were calculated from the microthermometric results. The ore-forming pressures were calculated using the empirical equation P = P0 × T/T0 (P0 = 219 + 2620S, T0 = 374 + 920S, where T is the homogenization temperature and S is the salinity) of Shao (1988), and the depths were calculated using the empirical equation of P = 2.7 × 0.0981 × H (the unit for P is bar and for H is m) from Shepherd et al. (1985). The calculated ore-forming pressures during stage I are in the range of 72–293 bar (Table 1), with peak values of 150–190 bar (Fig. 8d). The depths at the time when ores were precipitated ranged from 273 to 1107 m (Table 1).

5.1.2 Fluid inclusion compositions
The elemental compositions of fluid inclusions hosted by the stage I calcite and barite forming in mineralization event I and the sub-stage III2 calcite in mineralization event II are similar (Table 2). The cations are mainly Ca2+, Na+, K+, and Mg2+, although the Ca2+ can only be detected in the fluid inclusions hosted by barite. The anions are mainly SO42−, Cl−, F−, and NO3− and with minor Br−. However, the NaCl contents of the fluid inclusions hosted by sub-stage III2 calcite are higher than those of the stage I inclusions, and the Na+ and Cl− contents in the sub-stage III2 fluid inclusions can be as high as 21.55 ppm and 67.56 ppm (Table 2), respectively. They are the maximum contents of these two elements in all the fluid inclusions in the minerals from the two stages. The vapor phase of the fluid inclusions hosted in both stage I and sub-stage III2 minerals is mainly an oxidized vapor phase including N2, CO2, H2O, O2, and CO (Table 3). However, some reduced vapor phase, mainly CH4, is also present (Table 3). The Laser Raman analyses also found CH4 in the fluid inclusions hosted by stage I calcite (Fig. 9).

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Table 1 Microthermometric data of fluid inclusions hosted the calcite precipitating in the stage I in the Jiamoshan Pb-Zn deposit

<table>
<thead>
<tr>
<th>Sample no.</th>
<th>Host mineral</th>
<th>Tmwt (°C)</th>
<th>Thv-L (°C)</th>
<th>Salinity (wt% NaCl eq.)</th>
<th>Density (g/cm³)</th>
<th>Pressure (bar)</th>
<th>Depth (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>JMS10-3-3-1 Calcite</td>
<td>-19.8 to -12.0 (17)</td>
<td>132.3 to 243.9 (18)</td>
<td>15.96 to 22.24 (17)</td>
<td>1.02 to 1.09 (15)</td>
<td>158 to 249 (15)</td>
<td>596 to 941 (15)</td>
<td></td>
</tr>
<tr>
<td>JMS10-3-4-2 Calcite</td>
<td>-43.4 (1)</td>
<td>-18.6 to -12.6 (12)</td>
<td>92.2 to 223.9 (15)</td>
<td>16.5 to 21.33 (12)</td>
<td>0.83 to 1.10 (13)</td>
<td>123 to 291 (13)</td>
<td>464 to 1098 (13)</td>
</tr>
<tr>
<td>JMS10-3-10 Calcite</td>
<td>-42.0 (1)</td>
<td>-18.4 to -9.3 (14)</td>
<td>114.2 to 211.2 (16)</td>
<td>13.18 to 21.26 (14)</td>
<td>0.95 to 1.10 (15)</td>
<td>72 to 288 (15)</td>
<td>273 to 964 (15)</td>
</tr>
<tr>
<td>JMS10-2-5 Calcite</td>
<td>-18.6 to -7.2 (12)</td>
<td>125.2 to 222.5 (14)</td>
<td>10.73 to 21.40 (12)</td>
<td>0.93 to 1.09 (14)</td>
<td>88 to 293 (14)</td>
<td>331 to 1107 (14)</td>
<td></td>
</tr>
</tbody>
</table>

Notes: Numbers in parentheses refer to numbers of fluid inclusions analyzed; Tmwt, eutectic temperature of ice and salts containing H2O; Thv-L, homogenization temperature.

Table 2 Bulk ionic compositions (in ppm) and elemental molar ratios of calcite- and barite-hosted fluid inclusions from the Jiamoshan Pb-Zn deposit

<table>
<thead>
<tr>
<th>Sample no.</th>
<th>Mineral</th>
<th>Stage</th>
<th>Li</th>
<th>Na</th>
<th>K</th>
<th>Mg</th>
<th>Ca</th>
<th>F</th>
<th>Cl</th>
<th>Br</th>
<th>NO3</th>
<th>SO4^2-</th>
<th>Na/Br</th>
<th>Cl/Br</th>
<th>Na/K</th>
</tr>
</thead>
<tbody>
<tr>
<td>JMS10-2-5-1 Calcite</td>
<td>I</td>
<td>0.00</td>
<td>13.83</td>
<td>0.00</td>
<td>2.27</td>
<td>-</td>
<td>5.50</td>
<td>43.57</td>
<td>0.08</td>
<td>1.53</td>
<td>9.99</td>
<td>586</td>
<td>1196</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>JMS10-3-1 Calcite</td>
<td>I</td>
<td>0.00</td>
<td>1.92</td>
<td>0.00</td>
<td>0.73</td>
<td>-</td>
<td>4.66</td>
<td>2.02</td>
<td>0.29</td>
<td>2.20</td>
<td>8.80</td>
<td>23</td>
<td>16</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>JMS10-3-8 Calcite</td>
<td>I</td>
<td>0.00</td>
<td>3.28</td>
<td>1.04</td>
<td>0.85</td>
<td>-</td>
<td>6.43</td>
<td>4.54</td>
<td>0.72</td>
<td>3.62</td>
<td>26.06</td>
<td>16</td>
<td>14</td>
<td>5</td>
<td></td>
</tr>
<tr>
<td>JMS10-4-1 Calcite</td>
<td>I</td>
<td>0.37</td>
<td>7.12</td>
<td>6.11</td>
<td>4.95</td>
<td>-</td>
<td>12.98</td>
<td>29.73</td>
<td>0.63</td>
<td>3.71</td>
<td>46.34</td>
<td>40</td>
<td>107</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td>JMS10-25-1 Barite</td>
<td>I</td>
<td>0.00</td>
<td>1.85</td>
<td>1.76</td>
<td>0.00</td>
<td>44.93</td>
<td>0.43</td>
<td>2.39</td>
<td>0.34</td>
<td>2.24</td>
<td>23.61</td>
<td>19</td>
<td>16</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td>JMS10-3-3-3-2 Calcite</td>
<td>II_2</td>
<td>0.00</td>
<td>21.55</td>
<td>1.50</td>
<td>8.03</td>
<td>-</td>
<td>2.19</td>
<td>67.56</td>
<td>2.45</td>
<td>1.92</td>
<td>18.92</td>
<td>31</td>
<td>62</td>
<td>24</td>
<td></td>
</tr>
<tr>
<td>JMS10-3-6 Calcite</td>
<td>II_2</td>
<td>0.00</td>
<td>1.75</td>
<td>0.00</td>
<td>0.00</td>
<td>-</td>
<td>4.06</td>
<td>2.18</td>
<td>0.24</td>
<td>2.13</td>
<td>7.41</td>
<td>25</td>
<td>20</td>
<td>-</td>
<td></td>
</tr>
</tbody>
</table>

Table 3 GC bulk analyses of calcite- and barite-hosted fluid inclusions from the Jiamoshan Pb-Zn deposit

<table>
<thead>
<tr>
<th>Sample name</th>
<th>mineral</th>
<th>Stage</th>
<th>CH4</th>
<th>C2H2</th>
<th>C2H4</th>
<th>C2H6</th>
<th>CO2</th>
<th>H2O</th>
<th>O2</th>
<th>N2</th>
<th>CO</th>
<th>R</th>
</tr>
</thead>
<tbody>
<tr>
<td>JMS10-2-5-1 Calcite</td>
<td>I</td>
<td>0.11</td>
<td>0.04</td>
<td>0.00</td>
<td>17.25</td>
<td>23.18</td>
<td>7.25</td>
<td>50.08</td>
<td>2.09</td>
<td>0.13</td>
<td></td>
<td></td>
</tr>
<tr>
<td>JMS10-3-1 Calcite</td>
<td>I</td>
<td>0.07</td>
<td>0.03</td>
<td>0.00</td>
<td>15.86</td>
<td>3.68</td>
<td>12.05</td>
<td>65.89</td>
<td>2.43</td>
<td>0.16</td>
<td></td>
<td></td>
</tr>
<tr>
<td>JMS10-3-8 Calcite</td>
<td>I</td>
<td>0.03</td>
<td>0.02</td>
<td>0.00</td>
<td>13.30</td>
<td>6.77</td>
<td>12.21</td>
<td>66.51</td>
<td>1.17</td>
<td>0.09</td>
<td></td>
<td></td>
</tr>
<tr>
<td>JMS10-4-1 Calcite</td>
<td>I</td>
<td>0.03</td>
<td>0.02</td>
<td>0.00</td>
<td>12.44</td>
<td>15.31</td>
<td>10.71</td>
<td>60.29</td>
<td>1.19</td>
<td>0.10</td>
<td></td>
<td></td>
</tr>
<tr>
<td>JMS10-25-1 Barite</td>
<td>I</td>
<td>0.20</td>
<td>0.28</td>
<td>0.12</td>
<td>15.37</td>
<td>17.91</td>
<td>3.92</td>
<td>53.81</td>
<td>8.39</td>
<td>0.58</td>
<td></td>
<td></td>
</tr>
<tr>
<td>JMS10-3-3-3-2 Calcite</td>
<td>II_2</td>
<td>0.03</td>
<td>0.01</td>
<td>0.00</td>
<td>9.54</td>
<td>24.63</td>
<td>11.39</td>
<td>54.40</td>
<td>0.00</td>
<td>0.00</td>
<td></td>
<td></td>
</tr>
<tr>
<td>JMS10-3-6 Calcite</td>
<td>II_2</td>
<td>0.02</td>
<td>0.01</td>
<td>0.00</td>
<td>10.28</td>
<td>5.49</td>
<td>13.91</td>
<td>69.34</td>
<td>0.95</td>
<td>0.10</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Notes: R=(CH4+C2H2+C2H4+C2H6+CO)/CO2

Fig. 8. Histograms of (a) homogenization temperatures, (b) salinities, and (c) densities of the hydrothermal fluid, and (d) the ore-forming pressures of the Jiamoshan Pb-Zn deposit.
Stage I samples have $\delta^{13}$C$_{VPDB}$ values of $-4.9\%$ to $3.7\%$ and sub-stage III$_2$ samples have values of $-5.0\%$ to $3.5\%$. Analyses of oxygen isotopes in stage I samples yielded $\delta^{18}$O$_{V-SMOW}$ values of 18.8% to 21.9% and sub-stage III$_2$ samples yielded values of 15.1% to 22.3%. The results of hydrogen and oxygen isotopes in hydrothermal fluids hosted in the stage I calcite and barite and the sub-stage III$_2$ calcite are also given in Table 4. The $\delta^{13}$O$_{V-SMOW}$ values are $-145\%$ to $-93\%$ for stage I samples and $-136\%$ to $-96\%$ for sub-stage III$_2$ samples, and the $\delta^{18}$O$_{V-SMOW}$ values are $-2.22\%$ to $7.19\%$ for stage I samples and $5.78\%$ to $13.00\%$ for sub-stage III$_2$ samples.

5.2.2 Sulfur isotopes

The sulfur isotopic compositions for stage I galena and barite and sub-stage III$_2$ galena are summarized in Table 5. The $\delta^{34}$S$_{V-CDT}$ values for the galena of both stages are similar, $-3.2\%$ to $0\%$ for five stage I samples and $0.6\%$ for one sub-stage III$_1$, sample, but the $\delta^{34}$S$_{V-CDT}$ values for stage I barite are quite different and very positive up to $27.1\%$.

5.2.3 Pb isotopes

The Pb isotopic compositions for the same galena samples analyzed for sulfur isotopes are summarized in Table 6. The Pb isotopic ratios for all the galena samples are similar. For the stage I galena (5 samples) the values are $^{206}$Pb/$^{204}$Pb $= 18.3270$ to $18.3482$, $^{207}$Pb/$^{204}$Pb $= 15.6345$ to $15.6390$, and $^{208}$Pb/$^{204}$Pb $= 38.5503$ to $38.5582$, and for the sub-stage III$_2$ galena (1 sample) the values are $^{206}$Pb/$^{204}$Pb $= 18.3345$, $^{207}$Pb/$^{204}$Pb $= 15.6416$, and $^{208}$Pb/$^{204}$Pb $= 38.5699$.

6 Discussion

6.1 Nature and origin of the Jiamoshan hydrothermal fluid

For the stage I ore-forming fluid in the Jiamoshan Pb–Zn deposit, two eutectic melting temperatures (Te$_{en}$) were obtained from the microthermometric work with typical values of $-43.4\%$ and $-42.0\%$ (Table 1). These values indicate that the fluid was not a simple H$_2$O–NaCl system, which has a eutectic melting temperature of $-21.2\%$ (Davis et al., 1990), but had some other components such as CaCl and MgCl, which gave lower eutectic melting temperatures (Davis et al., 1990; Goldstein and Aizenshtat, 1994). The bulk ionic compositions of stage I fluid inclusions indicate abundant Ca$^{2+}$ as well as significant amounts of Mg$^{2+}$ and K$^+$ in addition to Na$^+$ and Cl$^-$ in the fluid. Taking into account the contents of negative ions, which include abundant SO$_4^{2-}$ and significant amounts of F$^-$ and NO$_3^-$ (Table 2), the stage I ore-forming fluid must have been a complex Mg$^{2+}$–Na$^+$–K$^+$–SO$_4^{2-}$–Cl$^-$–F$^-$–NO$_3^-$–H$_2$O system. The vapor phase in the stage I fluid was made up mainly of N$_2$, CO$_2$, H$_2$O, and O$_2$ (Table 3), but a reduced organic vapor phase, including CH$_4$ and other hydrocarbons (Table 3 and Fig. 9), cannot be ignored, and the R value ($R = (CH_4 + C_2H_6 + C_2H_4 + C_2H_2 + CO)/CO_2$), a vapor phase reduction parameter, was as much as 0.58 (Table 3), indicating the fluid existed in a weakly reducing and oxidizing environment (Xu et al., 2005). The homogenization temperatures and the salinities of fluid inclusions hosted by the stage I calcite indicate that the stage I ore-forming fluid was characterized by low temperatures (peak value 130–140°C), high salinities (peak value 23–24% NaCl eq.), and medium to high densities (peak value 1.04–1.06 g/cm$^3$) (Fig. 8), similar to a basinal brine (Fig. 10a).

The hydrogen and oxygen isotope analyses confirm a basinal brine origin for these ore-forming fluids. The values of $\delta^{18}$O$_{V-SMOW}$ and $\delta^{13}$O$_{V-SMOW}$ for the stage I and sub-stage III$_2$ fluids differ from those of magmatic water (Fig. 10b; Taylor, 1974). Moreover, for every stage, the plotted $\delta^{18}$O$_{V-SMOW}$ and $\delta^{13}$O$_{V-SMOW}$ values show a linear relationship and form positively sloped lines intersecting the meteoric water line (Fig. 10b; Craig, 1961; Crawford et al., 2014). This characteristic is similar to basinal brines in sedimentary basins worldwide (Fig. 10b; Kesler et al., 1997a). A basinal brine origin is therefore indicated for the

Table 4 C–O isotopic composition of calcite and H–O isotopic composition of calcite and barite from different mineralization stages during the Jiamoshan deposit formation

<table>
<thead>
<tr>
<th>Sample no.</th>
<th>Mineral</th>
<th>Mineralization stage</th>
<th>$\delta^{13}$C$_{VPDB}$ (%)</th>
<th>$\delta^{18}$O$_{VPDB}$ (%)</th>
<th>$\delta^{18}$O$_{SMOW}$ (%)</th>
<th>$\delta^{13}$O$_{SMOW}$ (%)</th>
<th>$\delta^{18}$O$_{SMOW}$ (%)</th>
<th>T (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>JMS10-4-1</td>
<td>Calcite</td>
<td>I</td>
<td>$-4.9$</td>
<td>$-11.7$</td>
<td>$18.8$</td>
<td>$-132$</td>
<td>$4.17$</td>
<td>125</td>
</tr>
<tr>
<td>JMS10-3-9</td>
<td>Calcite</td>
<td>I</td>
<td>$-3.7$</td>
<td>$-9.3$</td>
<td>$21.4$</td>
<td>$-119$</td>
<td>$6.71$</td>
<td>125</td>
</tr>
<tr>
<td>JMS10-3-4</td>
<td>Calcite</td>
<td>I</td>
<td>$3.6$</td>
<td>$-8.9$</td>
<td>$21.7$</td>
<td>$-93$</td>
<td>$7.07$</td>
<td>125</td>
</tr>
<tr>
<td>JMS10-2-5-1</td>
<td>Calcite</td>
<td>I</td>
<td>$2.7$</td>
<td>$-8.8$</td>
<td>$21.9$</td>
<td>$-101$</td>
<td>$7.19$</td>
<td>125</td>
</tr>
<tr>
<td>JMS10-25-1</td>
<td>Barite</td>
<td>I</td>
<td>-</td>
<td>$-5.0$</td>
<td>$15.3$</td>
<td>$-136$</td>
<td>$5.78$</td>
<td>204</td>
</tr>
<tr>
<td>JMS10-3-1</td>
<td>Calcite</td>
<td>III$_1$</td>
<td>$-5.0$</td>
<td>$-15.3$</td>
<td>$15.1$</td>
<td>$-136$</td>
<td>$13.60$</td>
<td>204</td>
</tr>
<tr>
<td>JMS10-3-3-2</td>
<td>Calcite</td>
<td>III$_2$</td>
<td>$3.5$</td>
<td>$-8.3$</td>
<td>$22.3$</td>
<td>$-96$</td>
<td>$9.08$</td>
<td>204</td>
</tr>
<tr>
<td>JMS10-3-6</td>
<td>Calcite</td>
<td>III$_3$</td>
<td>$3.0$</td>
<td>$-12.1$</td>
<td>$18.4$</td>
<td>$-122$</td>
<td>$9.08$</td>
<td>204</td>
</tr>
<tr>
<td>JMS10-3-8</td>
<td>Calcite</td>
<td>III$_4$</td>
<td>$-4.6$</td>
<td>$-14.5$</td>
<td>$15.9$</td>
<td>$-131$</td>
<td>$6.61$</td>
<td>204</td>
</tr>
</tbody>
</table>

Notes: $\delta^{18}$O$_{SMOW} = 1.0386/\delta^{13}$C$_{VPDB} + 30.86$ (Friedman and O’Neil, 1977). $\delta^{13}$O$_{VPDB}$ calculations use different fractionation equations for different minerals: 1000m$\delta^{13}$O$_{calcite-fad}$ = $2.78 	imes 10^{-6}$– $2.89$ (O’Neil et al., 1969); 1000m$\delta^{13}$O$_{calcite-fad}$ = $3.25 	imes 10^{-6}$– $5.1$ (Lloyd, 1968). T for the stage I is the peak homogenization temperature value got from this paper and for the sub-stage III$_2$ is from the unpublished work of Yingchao Liu.
ore-forming fluids of both stages of mineralization. It is noteworthy that the $\Delta V_{\text{SMOW}}$ values for the ore-forming fluids in the Jiamoshan deposit are relatively lower than those of other MVT deposits around the world (Fig. 10b; Kesler et al., 1997a), but they are similar to Mesozoic and Cenozoic meteoric water in the Sanjiang belt (Fig. 10b; Xu and Mo, 2000; Wang et al., 2018) and organic water (Fig. 10b). This suggests that the ore-forming fluids were mixed with the meteoric water of the region or that the primary ore-forming fluids reacted with organic matter. Taking into account the organic gas phase in the fluid inclusions (Table 3 and Fig. 9), the latter possibility seems most likely.

The compositions of the stage I and sub-stage III$_2$ fluid inclusions hosted by calcite are similar (Tables 2, 3), even though they relate to two separate mineralization events (Fig. 6). The values of Cl/Br and Na/Br in the fluids of both stages are lower than those of seawater (Fig. 11a; Fuge and Johnson, 1986; McCaffrey et al., 1987).

However, they are both distributed along the seawater-evaporation trend (Fig. 11a; Fontes and Matray, 1993), indicating that a possible origin for the fluids in both stages might be evaporated seawater that had been trapped in marine strata.

The carbon and oxygen isotopic data for the stage I and sub-stage III$_2$ calcites are also similar (Table 4). Firstly, their $\delta ^{13}C_{\text{PDB}}$ and $\delta ^{18}O_{\text{VSMOW}}$ values are apparently different from those of magmatic rocks (Fig. 11b; Shieh and Schwarcz, 1974), indicating that the carbon and oxygen could not have come from magmatic activities. Secondly, the calcite samples can be classified into two groups according to their $\delta ^{13}C_{\text{PDB}}$ values, with group I calcites having positive values (up to 3.7‰) and group II calcites having negative values (up to −5.0‰) (Fig. 11b). For the group I samples, the $\delta ^{13}C_{\text{PDB}}$ values are similar to those of the carbonate rocks in the region (Fig. 11b; Liu et al., 2013), indicating that the carbon in these calcites came from the dissolved carbonate rocks during water–rock reactions. For the group II samples, the $\delta ^{13}C_{\text{PDB}}$ values are apparently lower than those of the regional carbonate rocks, but they trend towards the $\delta ^{13}C_{\text{PDB}}$ values of sedimentary organic matter (Fig. 11b; Irwin et al., 1977). Taking into account the hydrocarbons in the fluid inclusions, the carbon of the group II calcite might have been derived from sedimentary organic matter as a result of reactions between the fluids and organic matter.

In summary, the hydrothermal fluids for the two mineralization events represented by stage I and sub-stage.

### Table 5 Isotopic composition of sulfides and barites from the Jiamoshan Pb-Zn deposit

<table>
<thead>
<tr>
<th>Sample no.</th>
<th>Mineral</th>
<th>Stage</th>
<th>$\delta ^{34}S$ (‰)</th>
</tr>
</thead>
<tbody>
<tr>
<td>JMS10-2-1</td>
<td>Galena</td>
<td>I</td>
<td>−1.7</td>
</tr>
<tr>
<td>JMS10-3-4-1</td>
<td>Galena</td>
<td>I</td>
<td>0</td>
</tr>
<tr>
<td>JMS10-3-5</td>
<td>Galena</td>
<td>I</td>
<td>−1.1</td>
</tr>
<tr>
<td>JMS10-3-1</td>
<td>Galena</td>
<td>III$_2$</td>
<td>0.6</td>
</tr>
</tbody>
</table>

### Table 6 Pb isotopic composition of galena from the Jiamoshan Pb-Zn deposit

<table>
<thead>
<tr>
<th>Sample no.</th>
<th>Mineral</th>
<th>Stage</th>
<th>$\Delta V_{\text{SMOW}}$</th>
<th>$\beta$</th>
<th>$\gamma$</th>
</tr>
</thead>
<tbody>
<tr>
<td>JMS10-2-1</td>
<td>Galena</td>
<td>I</td>
<td>18.3482 ± 0.0016</td>
<td>15.6390 ± 0.0014</td>
<td>38.5571 ± 0.0036</td>
</tr>
<tr>
<td>JMS10-3-4-1</td>
<td>Galena</td>
<td>I</td>
<td>18.3337 ± 0.0019</td>
<td>15.6354 ± 0.0018</td>
<td>38.5516 ± 0.0046</td>
</tr>
<tr>
<td>JMS10-3-5</td>
<td>Galena</td>
<td>I</td>
<td>18.3336 ± 0.0016</td>
<td>15.6364 ± 0.0015</td>
<td>38.5542 ± 0.0036</td>
</tr>
<tr>
<td>JMS10-5-4</td>
<td>Galena</td>
<td>I</td>
<td>18.3270 ± 0.0009</td>
<td>15.6345 ± 0.0008</td>
<td>38.5503 ± 0.0021</td>
</tr>
<tr>
<td>JMS10-3-1</td>
<td>Galena</td>
<td>III$_2$</td>
<td>18.3399 ± 0.0007</td>
<td>15.6381 ± 0.0005</td>
<td>38.5812 ± 0.0014</td>
</tr>
<tr>
<td>JMS10-5-3</td>
<td>Galena</td>
<td>III$_2$</td>
<td>18.3345 ± 0.0009</td>
<td>15.6416 ± 0.0008</td>
<td>38.6099 ± 0.0021</td>
</tr>
</tbody>
</table>

**Notes:** $\Delta V_{\text{SMOW}}=1000 \times \beta \left(\frac{\delta ^{206}Pb}{\delta ^{204}Pb} - 1\right)$; $\gamma = 1000 \times (\gamma \left(\frac{\delta ^{207}Pb}{\delta ^{204}Pb} - 1\right)$; $\beta$ and $\gamma$ represent the $\delta ^{207}Pb/\delta ^{204}Pb$ and $\delta ^{208}Pb/\delta ^{204}Pb$ for the mantle.
III$_2$ inclusions had similar origins, and they were probably derived from basinal brines. The basinal brines could have come from water in the marine strata of sedimentary formations in the region that originated as evaporated seawater that was trapped at depth, or it could also come from near the surface in the local Paleogene–Neogene basins (Fig. 1a, b). The hydrothermal fluids reacted strongly with the country rocks and with organic matter in the local sedimentary rocks.

6.2 Origin of the ore-forming materials

Sulfur is one of the main components of the economic minerals in hydrothermal deposits, so the sulfur isotopes in the sulfides and sulfates will provide important information about the sources of the ore-forming materials. The sulfur isotopes of the barite in the Jiamoshan deposit (Table 5 and Fig. 12) are totally different from the sulfur isotopes of contemporaneous seawater, which had $\delta^{34}$S values of 11‰ to 17‰ during the Late Triassic–Middle Jurassic (Claypool et al., 1980; Bottrell and Newton, 2006). This indicates that the sulfur in the barite could not have been directly sourced from residual sulfates in the trapped paleo-seawater of the local marine strata. The sulfur isotope compositions of galena samples from both the two mineralization events in the Jiamoshan deposit are similar (Table 5 and Fig. 12). The samples analyzed for their sulfur isotopes were just galena and barite, so it is difficult to ascertain whether or not the sulfur isotope fractionation was in equilibrium with the hydrothermal fluid, and this means that the total sulfur isotopic value cannot be calculated. Therefore, we can not explained that the sulfur came from the mantle (e.g., Chaussidon et al., 1989) according to the $\delta^{34}$S values of the galena samples which are near 0‰ (Table 5 and Fig. 12). Whether or not the sulfur isotopic fractionation was in equilibrium, values as low as $-3.2$‰ can be explained in terms of sulfur that came from sulfates undergone bacterial sulfate reduction (BSR; Warren, 2000; Machel, 2001). The fractionation of sulfur isotopes also provides useful information. For the purposes of discussion, sulfur isotopic data for gypsum taken from the regional Paleogene–Neogene basins (Fig. 1) have been collected (Fig. 12; Liu et al., 2013). The $\delta^{34}$S values for the Paleogene–Neogene gypsum minerals are lower than that of the Jiamoshan barite but higher than those of the Jiamoshan galena samples (Fig. 12). The differences in these values of $\delta^{34}$S are between 2‰ to 14‰ (Fig. 12), and this phenomenon can be explained by sulfur isotopic fractionation. The original sulfur might have come from SO$_4^{2-}$ derived from the gypsum by dissolution, but the sulfur with heavy isotopes fractionated to the barite while the sulfur with light isotopes fractionated to the galena during the hydrothermal evolution. Thus, the barite received heavier sulfur isotopes and the sulfide received lighter sulfur isotopes. According to the degree of sulfur isotopic fractionation between the gypsum and the galena (Fig. 12), this kind of process could occur during BSR (Warren, 2000; Machel, 2001). It is noteworthy that marine sulfates in the paleo-seawater trapped in the host carbonate rocks have sulfur isotopic values of about 11‰ to 17‰ (Claypool et al., 1980; Bottrell and Newton, 2006), which also lie between those of the galena and the barite. So, they might also have provided the sulfur for the mineralization by BSR in the Jiamoshan Pb–Zn deposit.

We cannot exclude either of these two possibilities, so we consider them both as having possibly contributed to the reduced sulfur in the Pb-Zn sulfide minerals of the Jiamoshan deposit.

Pb isotopic compositions are difficult to change during geological processes due to their high density, so it is appropriate to use them as indicators of sources of metals in ore deposits (e.g., Stacey et al., 1980; Zhu, 1995). Generally speaking, Pb minerals can provide more precise Pb isotope data, so stage I and sub-stage III$_2$ galena
samples that represent the two mineralization events were chosen for Pb isotope analyses. Although only a few samples were analyzed, the results provide important information on the origin of the Jiamoshan ore-forming metals. Firstly, the Pb isotopic ratios of the galena samples from both stages are exactly the same (Table 6 and Fig. 13), indicating that the metal sources for the two mineralization events were the same. Secondly, all the data of $^{206}$Pb, $^{207}$Pb, and $^{208}$Pb indicate that the lead came from the orogen (Fig. 13a, b), which is consistent with the collisional orogenic tectonic setting for the deposit (Liu et al., 2016, 2017). Furthermore, more information can be obtained by comparing these data with the regional Pb isotopic compositions (Fig. 13). We found that the values of $^{206}$Pb/$^{204}$Pb, $^{207}$Pb/$^{204}$Pb, and $^{208}$Pb/$^{204}$Pb for the galena samples in the Jiamoshan deposit differ from those of the regional carbonate and clastic rocks of Carboniferous to Triassic age (Fig. 13; Tian et al., 2011; Liu et al., 2013), which are either the carbonate hosts of the Pb–Zn deposits in the region or occur in the same general area as the deposit (e.g., Liu et al., 2013, 2015, 2016). The values of the Pb ratios also differ from those of the Cenozoic potassic volcanic rocks (Fig. 13), which are widespread in the Paleogene–Neogene basins of the region (Deng et al., 2001). However, the Pb isotopic compositions of the galena samples all lie within the range of the values for the Late Triassic volcanic-arc rocks in the region (Fig. 13; Hou et al., 1995; Leng et al., 2014), and they are approximately the same as the Pb isotopic compositions of sulfides in the VMS deposits hosted by these volcanic-arc rocks (Fig. 13; Zhu et al., 2001; Zhang F et al., 2017). Therefore, just as previous researchers’ thought that the ore-forming metals for the VMS deposits came from the host volcanic-arc rocks (e.g., Stuart et al., 1999; Mortensen et al., 2008), our new Pb isotopic data for the Jiamoshan deposit lead us to suggest that the Jiamoshan ore-forming metals also came from the Triassic volcanic-arc rocks. The differences in value of $^{207}$Pb/$^{204}$Pb and $^{208}$Pb/$^{204}$Pb between the galena samples in the Jiamoshan deposit and the mantle also provide consistent information. These differences can be represented by $\Delta \beta$ and $\Delta \gamma$ (Zhu et al., 1998). $\Delta \beta$ is defined as $1000 \times (\beta_{\text{M}} - 1)$ and $\Delta \gamma$ is defined as $1000 \times (\gamma_{\text{M}} - 1)$ where $\beta$ and $\gamma$ represent the values of $^{207}$Pb/$^{204}$Pb and $^{208}$Pb/$^{204}$Pb for the samples and $\beta_{\text{M}}$ and $\gamma_{\text{M}}$ represent the values of $^{207}$Pb/$^{204}$Pb and $^{208}$Pb/$^{204}$Pb for the mantle (Zhu et al., 1998). The plot of $\Delta \beta$ vs. $\Delta \gamma$ (Fig. 14) defined by the data of Pb isotopes in China provided useful information for determining the origin of the lead (Zhu et al., 1998) and the data from Jiamoshan deposit can be compared with them (Fig. 14). This comparison indicates that the Pb isotopic compositions of the Jiamoshan galena samples are the same as those of magmatic lead from a subduction zone (Fig. 14), which is consistent with the origin of the volcanic-arc rocks discussed above. In fact, Late Triassic volcanic-arc rocks are widespread in the middle part of the Sanjiang belt (Yang et al., 2012, 2019). These arcs were formed during the subduction of the Paleo-Tethys oceanic crust, and the arc rocks became involved in the Cenozoic
consistent with those of other MVT deposits formed in the Sanjiang thrust–fold belt (e.g., Hou et al., 2008; Liu et al., 2017; Song et al., 2019). However, the Jiamoshan deposit has ore-controlled structures and orebody occurrences that are unique in this Pb-Zn metallogenic belt. The Pb-Zn orebodies in the Jiamoshan deposit were controlled by karst caves where they formed irregularly shaped tubular bodies. Therefore, we consider that the Jiamoshan deposit represents a new style of MVT Pb-Zn mineralization formed in thrust–fold belts, and name it the karst-controlled style. To understand the origin of the ore-controlled karst caves (i.e., meteoric karst or hydrothermal karst; Sass–Gustkiewicz et al., 1982; Sangster, 1988) in the thrust–fold belts requires more detailed research. The existence of unmineralized paleo-karst caves in the area of the Jiamoshan deposit (Fig. 2) suggests that at least some pre-ore meteoric karst caves were formed, and this idea is given support by the existence of pre-ore karst caves in the vicinity of the Chaqupacha Pb-Zn deposit in the northern part of the Sanjiang belt (Zhang et al., 2012). However, the dissolution collapse breccias cemented by sulfides (Fig. 4c) in the Jiamoshan deposit demonstrate that the karst caves underwent further dissolution by hydrothermal fluids during the mineralization events. Karst caves in the thrust–fold belt could have formed at any time after the formation of the host rocks, but our geological observations show that all the karst-caves formed along reverse faults in the vicinity of the deposit (Figs. 2, 4a), which suggests that they formed after the regional thrusting, when the host rocks would have been taken down to shallower levels. This view is consistent with the relationship between the ore-controlled karst caves in the Changdong Pb-Zn deposit and the regional thrust faulting in the Simao Basin in the southern part of the Sanjiang belt (unpublished data from Yingchao Liu) and some MVT Zn-Pb deposits in the Tethyan belt (Song et al., 2017).

Therefore, the Jiamoshan deposit can be taken as the style example of karst-controlled MVT Pb-Zn deposits in a thrust–fold belt. A model for the Pb-Zn metallogenic processes of the Jiamoshan deposit are summarized in Fig. 15, and this model takes into account the geology and genesis of the deposit described here, as well as previous ideas on ore-forming processes in other MVT Pb-Zn deposits of the Sanjiang thrust–fold belt (Liu et al., 2016, 2017). The processes for mineralization events I and II in this deposit were similar, but as an example, we describe below the processes that took place during event I.

(1) When the regional thrusting took place, the strata in the area of the later-formed Jiamoshan deposit were uplifted by reverse faulting to arrive at a relatively shallow level in the crust. The carbonate rocks were soluble in the ground water and some meteoric karst caves formed (Fig. 15a), providing good structural traps. Paleogene–Neogene basinal brines that percolated down from the surface, and evaporated seawater that was trapped in the marine strata in the vicinity of the later ore-deposit, migrated and converged in these structural traps. Then, aided by organic matters from the sediments, the bacteria reduced the sulfates in the basinal brines and seawater in theses traps. Much H₂S was produced and therefore reduced sulfur...
(2) As the regional compression continued, the evaporated seawater trapped in the marine strata in the region was released and began to move along the detachment belt of the regional thrust fault system (Hou et al., 2008; Liu et al., 2016). During this process, water–rock reactions caused the fluids to leach materials from the widespread Triassic volcanic-arc rocks in the area, forming a regionally extensive fluid rich in metals such as Pb and Zn (Fig. 15a).

(3) When regional compression ceased, some of the thrust or reverse faults changed to extensional faults, opening spaces above the thrust faults. The falling pressure in the overlying strata led the fluid in the detachment belt to move rapidly upwards to arrive in the structural traps which were also reduced sulfur traps at that time and characterize the deposit now (Hou et al., 2008; Liu et al., 2016). The metals derived from the Triassic arc rocks were precipitated from the fluids by reactions with the reduced sulfur. Because the process of sulfide precipitation would form significant quantities of HCl (Goldstein and Aizenshtat, 1994; Machel, 2001), the carbonate host rocks were dissolved triggering the formation of dissolution collapse breccias (Sangster, 1988; Dzulynski and Sass-Guskiewiez, 1989). During this process the earlier-formed meteoric karst caves evolved to provide further open spaces for sulfide precipitation (Fig. 15b).

7 Conclusions

The formation of the Jiamoshan Pb–Zn deposit was controlled by karst structures that in turn were controlled by thrust and reverse faults, thus allowing orebodies to form with irregular tubular shapes. The main sulfide minerals of the deposit are galena, sphalerite, and pyrite that form massive and lamellar ore types. The deposit represents a new style of karst-controlled MVT Pb–Zn mineralization in a thrust–fold belt. The ore-forming fluid in this deposit belonged to the 

\[ \text{Mg}^{2+} - \text{Na}^{+} - \text{K}^{+} - \text{SO}_{4}^{2-} - \text{Cl}^{-} - \text{F}^{-} - \text{NO}_{3}^{-} - \text{H}_{2}\text{O} \]

system, and it was formed from basinal brines. The fluids were of low temperature but high salinity, and they were sourced from evaporated seawater trapped at depth in the local marine strata and might be also from Paleogene–Neogene basins at the surface. The sulfur for the sulfide minerals was produced by the process of bacterial sulfate reduction of a combination of sulfate derived by the solution of gypsum in the regional Paleogene–Neogene basins, and dissolved sulfates in the trapped paleo-seawater. The ore-forming metals were derived from the regional Triassic volcanic-arc rocks that formed during the closure of the Paleo-Tethys. The ore forming process was preceded by regional thrusting and reverse faulting that produced suitable structures for the formation of meteoric paleo-karst caves. Reduced sulfur reservoirs were produced in these caves during this process. Later post-thrusting extension then opened up pathways for the migration of regional basinal-brine fluids that were rich in Pb and Zn. They moved upwards into the karst structures that acted as structural traps, mixed with the reduced sulfur staying there, and triggered the precipitation of the Pb–Zn sulfides of the Jiamoshan Pb–Zn deposit.

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