Fluid Inclusion and Noble Gas Studies of the Dongping Gold Deposit, Hebei Province, China: A Mantle Connection for Mineralization?

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Abstract

The Dongping gold deposit (>100 t Au) occurs about 200 km inboard of the northern margin of the North China craton. The deposit is mainly hosted by syenite of a middle Paleozoic alkalic intrusive complex that was emplaced into Late Archean basement rocks. Both groups of rocks are intruded by Late Jurassic to Early Cretaceous crustal-melt granite dikes and stocks, some within a few kilometers of the deposit. The gold ores were deposited during this latter magmatic period at about 150 Ma, a time that was characterized by widespread regional north-south compression that formed the east-west–trending Yanshan deformational belt. The ores include both the telluride mineral-bearing, low sulfide quartz veins and the highly K-feldspar-altered syenite, with most of the resource concentrated in two orebodies (1 and 70).

Fluid inclusion microthermometry indicates heterogeneous trapping of low-salinity (e.g., 5–7 wt % NaCl equiv) fluids that varied from a few to 60 mole percent nonaqueous volatile species. Laser Raman spectroscopy confirms that the vapor phase in these inclusions is dominated by CO₂ but may be comprised of as much as 9 mole percent H₂S and 20 mole percent N₂; methane concentrations in the vapor phase are consistently <1 mole percent. The variable phase ratios are consistent with fluid immiscibility during ore formation. Fluid inclusion trapping conditions are estimated to be 250° to 375°C and 0.6 to 1.0 kbar. Helium isotope studies of fluid inclusions in ore-stage pyrites indicate ³He/⁴He ratios of 2.1 to 5.2 Ra (Ra = 1.4 × 10⁻⁶ for air) for orebody 1 and 0.3 to 0.8 Ra for orebody 70. The former data suggest that at least 26 to 65 percent mantle helium occurs in the fluids that deposited the veins in orebody 1. The lower values for orebody 70, which is characterized by a more disseminated style of gold mineralization, are interpreted to reflect an increased interaction of ore fluids with surrounding crustal rocks, which may have contributed additional ⁴He to the fluids.

A mantle source for at least some of the components of the gold-forming fluid is consistent with upwelling of hot asthenosphere and erosion of as much as 100 to 150 km of cool Archean lithosphere beneath the craton during this time. The Dongping deposit is located along the 100-km-wide north-south gravity lineament, which marks the western border of the thinned crust. As both regional metamorphism of Mesoproterozoic and younger cover rocks, and widespread granite magmatism, also occurred at ca. 150 Ma, it is unclear as to whether one or both of these also contributed fluid and/or metals to the hydrothermal system. Importantly, these new data suggest that economically significant gold deposits of similar mineral style and fluid composition, which are scattered along the margins of the craton, may all be products of a fluid originally partly sourced within the mantle.

Introduction

The telluride-rich Dongping gold deposit contains one of the largest economic resources within a series of gold systems that are spread over 2,000 km along the northern margin of the North China craton (Nie, 1997; Miller et al., 1998; Hart et al., 2002). It is located in Chongli county, northwestern Hebei province, and was discovered by a local farmer in 1985. Exploration was conducted in the Dongping area in 1987 by Branch 80 of the Gold Headquarters of the Ministry of Metallurgical Industry (MMI), and construction of the now-operating mine began the next year (Song, 1990, 1991; Wang et al., 1994; Zhang and Mao, 1995). Results from the geologic exploration revealed that the Dongping deposit had reserves of more than 100 metric tons (t) Au with an average grade of 6 g/t and that significant additional potential exists at depth and in the surrounding areas. The gold-bearing zones are consistently associated with tellurium-bearing minerals, most commonly calaverite (Zhang and Mao, 1995). As with many large tonnage gold deposits in China, annual production by local workers is relatively small by western standards and currently is estimated to be about 1 t Au.

Song and Zhao (1996) summarized the geology and geochemistry of the Dongping gold deposit and its resource potential. Since the discovery of the deposit, many scientific...
studies have been conducted that are concerned with the geology and geochemistry of the gold ores (Song, 1990, 1991; Wang et al., 1994; Xu et al., 1994; Zhang and Mao, 1995; Mo et al., 1996; Zhao, 1996), the associated Shuiquangou alkaline complex (Wang and Jiang, 1992; Wei and Su, 1994; Zhang and Mao, 1995; Jiang et al., 1996; Mo, 1997; Zhang, 1997), radiogenic isotope dating and stable isotope studies (Lu and Wang, 1992; Lu et al., 1993; Wang and Jiang, 1993; Jiang and Nie, 2000; Miao et al., 2002), and the ore-forming environment (Li et al., 1996; Niu et al., 1996; Nie, 1998). As many well-known gold-bearing telluride deposits have temporal and spatial links with alkaline igneous rocks (e.g., Cripple Creek in Colorado, Thompson et al., 1985; Emperor in Fiji, Ahmad et al., 1987), most researchers studying the telluride-rich Dongping gold deposit inferred that it was spatially and genetically related to the Shuiquangou alkaline complex. However, recent age data indicate that the Shuiquangou alkaline complex originated in the Variscan (middle to late Paleozoic), whereas the gold deposit formed in the middle to late Mesozoic (Hart et al., 2002; Jiang and Nie, 2000). Therefore, at best, the alkaline complex was only a favorable host rock for emplacement of the Dongping ores and there is no direct genetic link.

This paper constrains the temperatures, pressures, and compositions of the ore-forming fluids at the Dongping deposit and examines the isotopic compositions of noble gases in fluid inclusions to help constrain the sources of the ore fluids. The work comprises some of the most detailed published P-T-X data for any lode gold deposit in China. It is particularly significant since Dongping is one of the largest of the middle to late Paleozoic gold deposits in China but genetically remain poorly understood (see Hart et al., 2002; Zhou et al., 2002). The combined fluid inclusion and noble gas data from Dongping place important new constraints on conditions of ore formation, both at that deposit and, by inference, throughout much of northern China.

Geologic Setting

The Dongping deposit is located within the northern margin of the North China craton, 10 km south of the Chongli-Shangyi fault zone (Figs. 1, 2A). This east-to-northeast trending normal fault system marks the boundary between the Inner Mongolia Shield (or Inner Mongolian axis) to the north and the intracontinental, gold-hosting Yanshan deformational belt to the south (Davis et al., 2001), which is generally characterized by the broad Yanliao depression in the study area. The fault roughly parallels the more northerly Kangbao-Chifeng-Weichang regional fault zone, which separates the above-mentioned deformed rocks of the North China craton margin from those of the late Paleozoic Xing’an (or Greater Hinggan)-Inner Mongolia orogenic belt to the north. The latter collisional orogen extends westward to the Chinese Tian Shan and into the Central Asian countries and may be interpreted as the easternmost part of the Altaid orogenic belt (Ren et al., 1999; Yakubchuk et al., 2002).

The Chongli-Shangyi fault zone separates Paleoproterozoic metamorphic cover rocks to the north from Late Archean rocks to the south. The latter, the main lithologies in the area of the Dongping deposit, include gneiss, migmatite, and granulite of the Late Archean Sanggan Group dated at 2790 ± 155 Ma by the whole-rock Rb-Sr isochron method and at 2715 ± 21 Ma by the U-Pb zircon method (Nie, 1998). Low-grade to unmetamorphosed rocks of the Middle Proterozoic Changcheng Group, dated at 1776 Ma by U-Pb zircon methods (Cheng, 1994), occur locally and represent a sequence of clastic marine sedimentary rocks.

The Precambrian rocks are intruded by alkaline igneous rocks of the middle Paleozoic Shuiquangou complex, which is described in more detail below. Late Jurassic felsic to intermediate volcanic rocks unconformably overlie the Precambrian rocks and the Shuiquangou complex. For a more detailed description of the local stratigraphy, we refer readers to Nie (1998).

Magmatic rocks near the Dongping deposit are mainly represented by the Variscan Shuiquangou alkaline complex, which is exposed over an area 55 km long, from east to west, and 5 to 8 km wide (Fig. 2A). This large composite batholith is composed of several syenitic-monzonitic lithologies, including alkaline felsspar syenite, quartz-alkali feldspar syenite, pyroxene-hornblende syenite, hornblende-alkali feldspar syenite, pyroxene-hornblende monzonite, and hornblende monzonite. On the basis of the mineral assemblage, petrochemistry, and rare earth element geochemistry, Zhang and Mao (1995) divided the Shuiquangou complex into calc-alkaline, weakly alkaline, and alkaline series. Lead isotope data suggest a mantle origin for the intrusive bodies (Nie, 1998). Whole-rock Rb-Sr dates reported by Song and Zhao (1996) suggested emplacement of the complex ca. 327 to 309 Ma, but new SHRIMP U-Pb zircon dates indicate these to be cooling ages and crystallization occurred at 390 ± 6 Ma (Yumin Qiu, oral commun., 1999). This plutonic complex is most likely part of the Baoerhantu magmatic arc, which was developed within the northern part of the North China craton during southward-directed subduction of the Paleo-Asian Ocean throughout much of the Paleozoic (e.g., Yue et al., 2001). The paleosubduction zone is marked by the Suolun-Linxí suture (also referred to as the Solonker suture, Junggar-Hgen fault, or the Kangbao-Chifeng-Weichang regional fault zone of Fig. 1) between the craton and the Paleozoic Mongolian accretionary belt, approximately 200 km north of the Dongping area.

A number of Yanshanian (late Mesozoic) biotite granite plutons and dikes intrude the margins of the alkaline complex and the adjacent basement rocks. Potassium-argon biotite ages from these plutons and dikes range from 187 to 177 Ma (Nie et al., 1998). However, as pointed out by Davis et al. (2001), new U-Pb zircon work has shown that many of the Chinese K-Ar biotite ages from Mesozoic plutons in the Yanshan belt are inexplicably erroneously too old. The Shangshuiquan monzogranite, located within basement rocks about 5 km southwest of the Dongping deposit, has a U-Pb age of 142.5 ± 1.3 Ma (Miao et al., 2002). This age is consistent with other U-Pb ages for Late Jurassic-Early Cretaceous magnetism in the region (e.g., Davis et al., 2001). Unpublished radiogenic isotope data (Yumin Qiu, oral commun., 2001) suggest that the Yanshanian plutons are mainly products of crustal melting.

The structural evolution of the auriferous Yanshan deformational belt is complex and poorly understood (Davis et al., 2001). Regional features in the Dongping area include...
northeast- and north-trending normal and reverse faults, synclines filled with Jurassic strata, basement-cored anticlines, and amphibolite facies metamorphism of Proterozoic-Paleozoic cover sequences. The extensive Mesozoic tectonism seen throughout the region results from a combination of broadly contemporaneous Late Jurassic-Early Cretaceous contraction, due to final collision between Siberia and amalgamated Mongolia-North China, and extension caused by rollback of the northwest-dipping Pacific slab beneath northern China (Davis et al., 2001).

Gold Mineralization

The Dongping gold deposit is hosted mainly by syenite in the southern part of the Shuiquangou complex and, to a lesser degree, by rocks of the Archean basement (Fig. 2B). Mineralization is marked by a series of parallel auriferous quartz vein swarms and altered wall rocks, which occur along fault zones that strike from 0° to 35° and dip to the northwest at 45° to 75°. Among these, vein systems 1 and 70 are the largest orebodies, containing about 80 percent of the total gold resource at the Dongping deposit. These two orebodies were originally defined to be distinct mineralized zones, but more recent exploration indicates that they are actually connected (Fig. 2B). Individual ore-bearing veins are commonly a few hundred to a few thousand meters long, 0.3 to 8 m wide, and extend down-dip for 200 to 600 m. A detailed structural analysis of this important gold deposit has not yet been carried out.
Intense K-feldspar alteration was developed on both sides of the main quartz veins (Fig. 3), most commonly a few tens of centimeters to a few meters wide, although it may locally reach a maximum width of 20 m. Frequently, the alteration zones also contain gold and, therefore, also constitute a part of the orebodies. Gold ores in altered country rock are particularly common in orebody 70, in contrast to the more common occurrence of auriferous quartz–K-feldspar veins in orebody 1 and other orebodies.

Although various studies have subdivided the mineralization into three or four nearly identical paragenetic stages (e.g., Zhang and Mao, 1995; Nie, 1998), our field observations indicate that the mineralization consists of one main stage of massive auriferous quartz veins and coeval, gold-bearing, quartz–K-feldspar veinlets and extremely red, K-feldspar–altered wall rock. The average abundance of sulfide minerals is about 3 vol percent throughout the various orebodies, although there is much variation with sulfide abundance and grain size being greater in the veins. Pyrite is the dominant sulfide mineral in the orebodies, with lesser sphalerite, galena, and chalcopyrite. Tellurides are ubiquitous in the ore zones, and magnetite and hematite are also common. Generally, the greater the sulfide mineral content, the higher the gold grade within the veins; high gold grades in wall rock correlate with more intense zones of K-feldspar alteration. The silver content of the ores ranges from 0.2 to 6.5 ppm (Song and Zhao, 1996). Although carbonate stockwork veinlets and barite veins are rarely found to cut the orebodies, they are both postgold mineralization features and were not studied in the present work.

There have been numerous isotopic dating studies of the Dongping deposit. Wang and Jiang (1992) used the K-Ar method to date hydrothermal K-feldspar in the orebodies and obtained an age of 157 Ma. Lu et al. (1993) used both the K-Ar and 40Ar-39Ar methods to similarly date K-feldspar and obtained ages of 148 ± 2 and 157 ± 0.9 Ma, respectively. Wang (1992) carried out additional 40Ar-39Ar dating of K-feldspar from both the Dongping deposit and the adjacent Houguon deposit and reported ages of 177 ± 5 and 173 ± 5 Ma, respectively. The most recent Ar-Ar geochronology, using hydrothermal micas, indicates that gold veining occurred at about 150 Ma (Hart et al., 2002), as was suggested by Lu et al. (1993). Mineralization is, therefore, interpreted to be coeval with the Late Jurassic-Early Cretaceous magmatic episode within the Yanshan belt.

Fluid Inclusion Studies

Microthermometric data from the Dongping deposit have been previously published in two other English language journals (Zhang and Mao, 1995; Nie, 1998), but those data are of a preliminary nature and were mainly simply translated from poorly documented Chinese sources and lack detailed interpretation. Nie (1998) originally suggested that there were four distinct stages of mineralization consisting of (1)
gold-bearing K-feldspar–quartz stockworks and veins, (2) disseminated sulfide and gold zones, (3) gold-bearing quartz veins, and (4) barren calcite-quartz veins. More recently, however, he has stated that the first three styles of gold mineralization reflect a single event (Feng-Jun Nie, oral commun., 2000).

Zhang and Mao (1995) briefly mentioned data for 13 vein samples. They noted final fluid inclusion homogenization temperatures ranging between 240° to 350°C for quartz and quartz-feldspar veins, salinities ranging between 1.5 to 10.0 wt percent NaCl equiv, based on ice melting temperatures (which must be maximum values because of the presence of clathrates), the common presence of CO₂, and estimated fluid inclusion trapping pressures of 500 to 690 bars. Nie (1998) reported homogenization temperatures of 220° to 340°C and salinities of 5 to 21 wt percent NaCl equiv for fluid inclusions from auriferous quartz–K-feldspar veins, and 190° to 270°C and 1 to 7 wt percent NaCl equiv for inclusions in later auriferous quartz veins and disseminated sulfide and gold zones. Coexisting liquid- and vapor-rich inclusions were interpreted as representative of fluid immiscibility in some samples. Nie (1998) claimed that homogenization temperatures represented trapping temperatures, trapping pressures were 550 to 650 bars, and the ore fluids contained 28 to 42 mole percent

FIG. 3. Geologic cross section of exploration line 15 in the Dongping gold deposit (from Song and Zhao, 1996). Auriferous quartz veins are commonly accompanied by halos of auriferous K-feldspar–altered rocks, both of which define the ore. The host rock is a 390 Ma syenite body.
CO₂ in the quartz-feldspar veins and 35 to 54 mole percent CO₂ in the quartz veins. However, measurements were only presented for the final homogenization temperatures and it is, therefore, difficult to evaluate many of the interpretations and conclusions of Nie (1998).

We present new, detailed microthermometric and laser Raman spectroscopy data for ore-stage fluid inclusions from the Dongping deposit. Doubly polished thin sections were prepared from nine representative samples of the gold- and telluride-bearing quartz veins and quartz–K-feldspar veins. Sections were studied with a petrographic microscope and a Chaixmeca heating-freezing stage at the Chinese Academy of Geological Sciences. The behavior of different types of fluid inclusions was documented during heating and cooling experiments, and the temperatures of various phase changes in different types of inclusions were measured. Selected individual inclusions were also analyzed by laser Raman spectroscopy. Phase diagrams of corresponding fluid systems and/or thermodynamic equations were used to determine the thermodynamic parameters such as temperature, pressure, and composition of the ore-forming fluids.

**Types and features of fluid inclusions**

Quartz is the principal transparent mineral that is paragenetically related to gold mineralization and contains abundant fluid inclusions. Measurements in quartz were typically restricted to inclusions that appeared primary or pseudosecondary in nature (Roedder, 1984). These inclusions have relatively regular shapes (negative crystal or incomplete negative crystal shapes) and are dispersed in quartz grains, or occur along microcracks within quartz grains, but are not found cutting the crystal boundaries. Inclusions occurring as groups usually have similar degrees of fill and homogenization temperatures, as well as relatively consistent compositions, with CO₂ and H₂O as the dominant components. Aqueous-type inclusions are more abundant than the CO₂ types in the Dongping samples and occur as one-, two-, or multiphase inclusions. Many of these aqueous inclusions frequently coexist with the CO₂-type inclusions. Their behavior of different types of fluid inclusions was documented during heating and cooling experiments, and the temperatures of various phase changes in different types of inclusions were measured. Selected individual inclusions were also analyzed by laser Raman spectroscopy. Phase diagrams of corresponding fluid systems and/or thermodynamic equations were used to determine the thermodynamic parameters such as temperature, pressure, and composition of the ore-forming fluids.

Secondary inclusions have not been studied. These secondary inclusions have not been studied.

Primary or pseudosecondary fluid inclusions in gold-bearing quartz may be classified as CO₂ type (or CO₂-rich type) and H₂O type (H₂O-rich type) based upon their phase relationships and chemical compositions at room temperature. Their basic features are described below.

The CO₂-type inclusions most commonly contain significant, but quite variable, amounts of liquid CO₂. These have relatively regular shapes (usually negative crystal or incomplete negative crystal shapes) and are about 5 to 30 µm (mainly 8–15 µm) in maximum dimension. At room temperature, they appear as three-, two-, and one-phase CO₂-rich inclusions, although the one-phase probably contain a small amount of water. The CO₂-type inclusions are generally reddish to grayish when viewed at room temperature, and the V₁₀₂ color is slightly darker than L₁₀₂. The three-phase CO₂-type fluid inclusions (Fig. 4A–F), consisting of vapor, L₁₀₂ and L₁₁₂, are the most common of the three CO₂-rich varieties of inclusion. The CO₂ phase in the latter inclusions varies widely in volume, from 10 to >90 percent of the inclusion and mainly from 20 to 60 percent. Some of the two-phase fluid inclusions also nucleate a third phase during slight cooling below room temperature. Two- and one-phase (i.e., pure CO₂) CO₂-rich inclusions are much less common. The two-phase inclusions are characterized by V₁₀₂ of 10 to 25 percent and the remainder of the inclusion contains L₁₀₂ (and likely a small amount of water). The one-phase CO₂-rich inclusions nucleate a second phase (e.g., V₁₀₂ + L₁₁₂) when cooled below room temperature.

**Aqueous-type inclusions**

Aqueous-type inclusions are more abundant than the CO₂ types in the Dongping samples and occur as one-, two-, or multiphase inclusions. Many of these aqueous inclusions frequently coexist with the CO₂-type inclusions. Their behavior of different types of fluid inclusions was documented during heating and cooling experiments, and the temperatures of various phase changes in different types of inclusions were measured. Selected individual inclusions were also analyzed by laser Raman spectroscopy. Phase diagrams of corresponding fluid systems and/or thermodynamic equations were used to determine the thermodynamic parameters such as temperature, pressure, and composition of the ore-forming fluids.

**Microthermometric measurements**

Microthermometric measurements were carried out on a calibrated Chaixmeca heating-freezing stage (+60°C to −180°C). The heating rate was 0.1° to 0.2°C/min between 0° and 10°C and below −56.6°C, with an accuracy of ±0.1°C, whereas rates were about 3° to 5°C/min between 10° to 31°C and 5° to 10°C/min at higher temperatures, both with an accuracy of about ±1°C.
Fig. 4. Photomicrographs of fluid inclusion types from gold-bearing veins of the Dongping deposit. (A)-(F) show the variable phase ratios for a group of three-phase CO$_2$-type inclusions in quartz (at room temperature in plane-polarized light), with (B) and (C) clearly showing some of the adjacent, coexisting CO$_2$- and H$_2$O-type inclusions. A. Volume of CO$_2$ is 70 percent and homogenization is to a vapor phase. B. Volume of CO$_2$ is >75 percent and homogenization is to the LCO$_2$ phase. C. Volume of CO$_2$ is 55 percent and homogenization is to a liquid. D. Volume of CO$_2$ is 85 percent and homogenization is to the LCO$_2$ phase. E. Volume of CO$_2$ is 35 percent and homogenization is to a liquid. F. Volume of CO$_2$ is 20 percent and homogenization is to a vapor. G. and I. Vapor-rich H$_2$O inclusion that homogenizes to a vapor phase. H. Two-phase vapor-liquid H$_2$O inclusion that homogenizes to a liquid phase. J. Three-phase halite (H)-vapor-liquid H$_2$O inclusion.
CO₂-type inclusions

A total of 63 CO₂-type inclusions were measured in seven gold-bearing quartz vein samples from orebodies 1 and 70, and these data are summarized in Table 1 and Figure 5. The melting temperatures of CO₂ typically range from –56.6 to –57.9°C (Fig. 5), suggesting only minor contamination of the CO₂ with other volatiles. A few data with melting temperatures warmer than –56.6°C may reflect slight inaccuracies with the calibration of the stage or too rapid heating cycles. The clathrate-melting temperatures obtained from analyses of 34 inclusions range from 5.1° to 7.6°C. Using the equation of Bozzo et al. (1973), and assuming no effect from minor methane or nitrogen, the salinities of the three-phase CO₂-type inclusions range between 4.7 to 8.9 wt percent NaCl equiv (Table 1, Fig. 6) and mainly 5 to 7 wt percent NaCl equiv. Using final homogenization temperatures (see below) and the above salinities, the resulting densities of the aqueous ($\rho_{H_2O}$) phase range between 0.668 and 0.826 g/cm³ (Table 1), using the NaCl-H₂O reference table of Liu and Shen (1999).

The CO₂ homogenization temperatures for the 63 CO₂-type inclusions ranged widely between 0.3° to 31.1°C, with a mode between 26.0° and 30.9°C (Table 1, Fig. 7). Of these samples, 19 homogenized to a vapor when heated and the remainder to a liquid (Fig. 7). Given these temperatures, as well as the final homogenization temperatures described below, we can obtain corresponding CO₂ densities mainly between 0.2 and 0.8 g/cm³, when using the relationships of Liu and Shen (1999). There is a distinct gap in the density data between about 0.37 and 0.52 g/cm³. Also, the three inclusions with anomalously low CO₂ homogenization temperatures (0.3°, 7.4°, and 14.9°C), each homogenizing to a liquid in two-phase, CO₂-rich inclusions, have relatively high corresponding densities of 0.913, 0.873, and 0.815 g/cm³, respectively.

We measured final homogenization temperatures $T_{h(total)}$ in 39, three-phase, CO₂-type inclusions (Table 1, Fig. 8). Two-thirds of these homogenized into a liquid at temperatures of 247° to 365°C and the remainder homogenized by vapor expansion at 255° to 372°C. The similar homogenization temperatures are suggestive of overlapping trapping conditions for both groups of inclusions.
### Table 1. Microthermometric Data for CO₂-Type Inclusions in Quartz

<table>
<thead>
<tr>
<th>Sample</th>
<th>T_m(CO₂) (°C)</th>
<th>T_m(cl) (°C)</th>
<th>T_h(CO₂) (°C)</th>
<th>Salinity</th>
<th>Density (g/cm³)</th>
<th>Volume %</th>
<th>CO₂</th>
<th>H₂O</th>
<th>Total CO₂</th>
<th>H₂O</th>
<th>X(CO₂)</th>
<th>X(H₂O)</th>
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<tr>
<td>57.8</td>
<td>5.9</td>
<td>29.9</td>
<td>360 (Critical)</td>
<td>7.64</td>
<td>0.331</td>
<td>0.676</td>
<td>0.521</td>
<td>45</td>
<td>55</td>
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<td>0.858</td>
<td>0.022</td>
<td>80</td>
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<tr>
<td>57.8</td>
<td>7.0</td>
<td>24.8</td>
<td>365</td>
<td>5.77</td>
<td>0.238</td>
<td>0.677</td>
<td>0.479</td>
<td>45</td>
<td>55</td>
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<td>0.563</td>
<td>0.017</td>
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<td>5.9</td>
<td>27.6</td>
<td>333</td>
<td>7.64</td>
<td>0.277</td>
<td>0.712</td>
<td>0.209</td>
<td>95</td>
<td>5</td>
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<td>0.686</td>
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<td>324</td>
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<td>0.755</td>
<td>0.498</td>
<td>50</td>
<td>50</td>
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<td>56.7</td>
<td>7.3</td>
<td>25.2</td>
<td>330</td>
<td>5.23</td>
<td>0.699</td>
<td>0.706</td>
<td>0.700</td>
<td>90</td>
<td>10</td>
<td>0.65</td>
<td>0.362</td>
<td>0.018</td>
<td>&lt;80</td>
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<td>0.741</td>
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<td>56.7</td>
<td>7.3</td>
<td>28.1</td>
<td>324</td>
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<td>0.006</td>
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<td>0.729</td>
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<td>15</td>
<td>0.64</td>
<td>0.354</td>
<td>0.006</td>
<td>&lt;80</td>
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</tr>
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</table>

Notes: T_m(CO₂) = melting temperature of CO₂, T_m(cl) = melting temperature of CO₂ clathrate, T_h(CO₂) = partial homogenization temperature of CO₂ inclusions, T_h(total) = total homogenization temperature of three-phase CO₂ inclusions; X(CO₂), X(H₂O), and X(NaCl) are mole fractions of CO₂, H₂O, and NaCl in CO₂-type inclusions, respectively; P = pressure during total homogenization of three-phase CO₂ inclusions, V = vapor phase, L = liquid phase.
Total densities for the three-phase fluids, assuming a CO2-NaCl-H2O system, are calculated using the equation of Liu and Shen (1999):

$$\rho_{\text{total}} = \phi_{\text{CO2}} \rho_{\text{CO2}} + (1 - \phi_{\text{CO2}}) \rho_{\text{aq}}, \quad (1)$$

where $\rho_{\text{total}}$ is the total density ($g/cm^3$) of fluids, $\phi_{\text{CO2}}$ is the CO2 volume/total inclusion volume during CO2 partial homogenization, $\rho_{\text{CO2}}$ is the density ($g/cm^3$) of the CO2 phase during CO2 partial homogenization, and $\rho_{\text{aq}}$ is the density ($g/cm^3$) of the aqueous liquid phase during CO2 partial homogenization. Calculations indicate that the total densities of the three-phase, CO2-type inclusions range from 0.284 to 0.813 g/cm³ but mainly between 0.479 and 0.794 g/cm³.

The $X_{\text{CO2}}$ of three-phase CO2-type inclusions can be determined from the $\rho_{\text{CO2}}$ and $\rho_{\text{total}}$ obtained above and using the $X_{\text{CO2}}$-$\phi_{\text{CO2}}$-$T_{\text{th(total)}}$-$\rho_{\text{aq}}$ diagram (Schwartz, 1989) for the CO2-H2O 6 wt percent NaCl equiv system. Thus, $X_{\text{CO2}}$ is about 0.04 to 0.95, with a range of 0.2 to 0.6, typical of most inclusions; $\phi_{\text{CO2}}$ is essentially consistent with the $X_{\text{CO2}}$ (12–95%) measured under the microscope. Substituting the measured salinities of aqueous phases of this type of inclusion into the equation for total salinity (e.g., Liu and Shen, 1999) results in the following:

$$X_{\text{NaCl(aq)}} = \frac{18.0152 \cdot \omega \cdot 18.0152 \cdot \omega + 58.4428 \times (100 - \omega)}{18.0152 \cdot \omega + 58.4428}, \quad (2)$$

where $X_{\text{NaCl(aq)}}$ is the mole fraction of NaCl in the aqueous phase and $\omega$ is the salinity of that phase. Then, substituting $X_{\text{NaCl(aq)}}$ in the following equation:

$$X_{\text{NaCl}} = X_{\text{NaCl(aq)}} \cdot [1 - X_{\text{CO2}}], \quad (3)$$

we get the total mole fraction of NaCl in the three-phase, CO2-type inclusions as 0.004 to 0.026 (Table 1). The mole fraction of H2O in these inclusions, determined from the equation $X_{\text{H2O}} = 1 - X_{\text{NaCl}} - X_{\text{CO2}}$, is between 0.166 and 0.944 (Table 1).

In the $X_{\text{CO2}}$-$\phi_{\text{CO2}}$-$T_{\text{th(total)}}$-$\rho_{\text{aq}}$ diagram for CO2-H2O 6 wt percent NaCl equiv (Schwartz, 1989), the internal pressure of three-phase CO2 inclusions during total homogenization can be derived from the $X_{\text{CO2}}$ and $\phi_{\text{CO2}}$ obtained above. Calculated pressures are in the range of <0.4 to 1.8 kbars and mainly in the range of 0.6 to 1.2 kbars (Table 1), when the $\phi_{\text{CO2}}$ is estimated from diagrams showing relationships at 40°C. Because the partial homogenization temperatures of the CO2 phase are mainly between 26° and 31°C, and the $\phi_{\text{CO2}}$ values do not show much difference from those at 40°C, estimation of the $\phi_{\text{CO2}}$ values during partial homogenization of CO2-type inclusions is valid.

In summary, the significant microthermometric parameters of three-phase, CO2-type inclusions are as follows: (1) total homogenization temperatures are 247° to 365°C for inclusions homogenizing to a liquid and 255° to 372°C for inclusions homogenizing to a vapor; (2) trapping pressures are estimated to be 0.6 to 1.2 kbars; (3) salinities of fluids are mainly 5 to 7 wt percent NaCl equiv; (4) total densities of fluids are mainly 0.479 to 0.794 g/cm³, being 0.412 to 0.813 g/cm³ for inclusions that homogenize to a liquid and 0.284 to 0.800 g/cm³ for those that homogenize to a vapor; and (5) $X_{\text{CO2}}$ is mainly 0.2 to 0.6, although the range may vary from 0.04 to 0.95.

**Aqueous-type inclusions**

A total of 231 H2O-NaCl inclusions were measured in quartz and feldspar, which can be subdivided into 206 liquid rich, 18 vapor rich, and seven daughter minerals-bearing inclusions; the latter two are solely in quartz. All the microthermometric data are summarized in Table 2. As shown in Figure 9, the homogenization temperatures of the aqueous inclusions range from 160° to 430°C, with most defining a bimodal distribution from 250° to 380° and 160° to 250°C. The lower temperature mode reflects inclusions observed and classified as secondary, which were trapped subsequent to vein formation and deposition of the ore minerals. They represent some of the aqueous inclusions in the quartz but all of the observed inclusions in the K-feldspar samples as mentioned above. The relatively few vapor-rich fluid inclusions give some of the highest homogenization temperatures for the aqueous inclusions, with most of the measurements

<table>
<thead>
<tr>
<th>Homogenization</th>
<th>Numbers and types</th>
<th>Ice-melting</th>
<th>Salinity</th>
<th>Density (g/cm³)</th>
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</thead>
<tbody>
<tr>
<td>temperature (°C)</td>
<td>of inclusions¹</td>
<td>temperatures (°C)</td>
<td>(wt % NaCl equiv)²</td>
<td></td>
</tr>
<tr>
<td>&lt;150–230</td>
<td>Two-phase (34); polyphase (4)</td>
<td>3.3–5.0 (n = 16)</td>
<td>5.4–7.9</td>
<td>0.864–0.946</td>
</tr>
<tr>
<td>251–290</td>
<td>Two-phase (45); polyphase (2)</td>
<td>3.0–4.2 (n = 12)</td>
<td>5.0–6.7</td>
<td>0.749–0.852</td>
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<tr>
<td>291–350</td>
<td>Two-phase (110); polyphase (1); vapor-rich (7)</td>
<td>3.0–5.2 (n = 51)</td>
<td>5.0–8.1</td>
<td>0.694–0.786</td>
</tr>
<tr>
<td>351–390</td>
<td>Two-phase (17); vapor-rich phase (10)</td>
<td>3.3–5.2 (n = 4)</td>
<td>5.4–8.1</td>
<td>0.667–0.728</td>
</tr>
</tbody>
</table>

¹The digit in the parenthesis is the number of inclusion measured
²The few high salinities for rare polyphase inclusions are not listed here but are described in the text
Inclusion in quartz
- Aqueous (Th to liquid)
- Aqueous (Th to vapor)
- Polyphase
- 3-phase CO2-rich

Inclusion in feldspar
- Aqueous (Th to liquid)
- Polyphase

The polyphase inclusions are small in size, with correspondingly very small daughter crystals, so phase changes were measured in only seven inclusions in quartz. Of these, (1) one is a saturated saline inclusion in which vapor and liquid phases homogenized and the daughter crystal melted both at a temperature of 281°C, (2) two were unsaturated saline inclusions in which vapor-liquid homogenization temperatures (344°C and 276°C) are higher than the melting temperatures (248° and 158°C) of daughter crystals, and (3) the other four are supersaturated saline inclusions in which daughter crystals (halite) melted at 333°, 332°, 338°, and 343°C, whereas their corresponding homogenization temperatures (172°, 173°, 196°, and 226°C). Because the oversaturated saline inclusions represent cavities in which the daughter minerals and solution were captured at the same time, we assume that the melting temperatures of the daughter minerals do not reflect the salinities of the original solutions. Therefore, we chose the three former inclusions to calculate the salinities, which are 29.6, 34.6, and 36.7 wt percent NaCl equiv, respectively (Fig. 6).

Estimated fluid densities for the H2O-NaCl system can be determined from the appropriate table of Liu and Shen (1999; relationships are consistent with those of Shepherd et al., 1985, and references therein), knowing the homogenization temperatures and salinities of the aqueous fluid inclusions. The three daughter mineral-bearing polyphase inclusions with estimated salinities show relatively high densities of 1.022, 1.038, and 1.081 g/cm³, whereas the 83 other aqueous inclusions have densities of 0.642 to 0.946 g/cm³, with most ranging from 0.694 to 0.852 g/cm³ (Table 2). As a generalization, the densities of the aqueous inclusions from the Dongping deposit have a negative correlation with homogenization temperatures. The fluid inclusions with temperatures >350°C have relatively low densities of 0.667 to 0.728 g/cm³, those with TH of 230° to 350°C have densities of 0.694 to 0.852 g/cm³, and those with TH <250°C have high densities of 0.864 to 0.946 g/cm³. The latter are the postmineralization secondary fluid inclusions with high degrees of filling and occurring along late fractures.

**Laser Raman spectroscopy analysis for fluid inclusions**

Nine inclusions were chosen (one- and three-phase CO2 type and one- and two-phase aqueous-type inclusions) for quantitative laser Raman spectroscopy analysis of the volatile species such as CO2, H2S, CH4, CO, N2, C2H2, and higher hydrocarbons within the vapor and liquid phases. The analyses were conducted at the Xi’an Institute of Geology and Mineral Resources with a JY Inc RAMANOR-U1000 laser Raman spectrograph made in France. The results from the analyses are listed in Table 3. The 514.5-nm laser line from a spectra physics model 2020 3W Ar+ laser was used as the exciting source. Laser power was 45 mW for a micrometer spot size, and the spectral bandpass was approximately 9.2 cm⁻¹. Calibration was carried out using a silica metal standard. Detection limits were 0.06 mole percent for CO2, 0.04 mole percent for CH4, 0.02 mole percent for N2, and 0.03 mole percent for H2S. Resulting data are consistent with the microthermometry, as CO2 and H2O are the main volatiles in most of the measured inclusions. Small amounts of N2, H2S, CH4, CO, and C2H2 were also detected. The seven analyzed two- and three-phase inclusions contain significant amounts of CO2. The vapor phase of the CO2-type inclusions contains about 70 to 96 mole percent CO2, and the vapor phase of the two two-phase aqueous-type inclusions contains 75 to 80 mole percent CO2. In contrast, the two postgold, one-phase aqueous-type inclusions contain no CO2 in their vapor phase and only very minor CO2 (1.2 mol %) dissolved in the liquid phase of one of the inclusions (Table 3). Moreover, the vapor phase in the two- and three-phase CO2- and aqueous-type inclusions contains as much as 9 mole percent H2S and varies between about 2 to 20 mole percent N2; some of the more elevated N2 concentrations characterize the aqueous-type inclusions. Approximately 0.5 to 1.0 mole percent CH4 occurs within the vapor in the CO2-type inclusions, but the species was not detected in the two-phase aqueous-type inclusions.
As described above, there are two main, relatively early, spatially associated fluid inclusion types: the CO$_2$-rich type and aqueous type. These have nearly identical final homogenization temperatures and salinities, as was shown in the microthermometric studies. Moreover, both have significant amounts of CO$_2$, N$_2$, and H$_2$S in their vapor phase. We, therefore, infer that they are derived from the same parent fluid and underwent phase separation during the ore-forming process. The Raman study has shown that the aqueous-type inclusions (the high-temperature mode) are, in actuality, low end-member CO$_2$-type inclusions. They contain relatively small vapor bubbles of CO$_2$ ± H$_2$S, N$_2$ and thus were too small to observe clathrate formation during freezing experiments. These CO$_2$-bearing inclusions contrast strongly with the postgold, aqueous, secondary inclusions with lower salinity, lower homogenization temperatures, high fill degrees, high densities, and ≤1 mole percent CO$_2$.

### Noble Gas Isotopes of Fluid Inclusions

As suggested by Giggenbach (1986), isotopes of noble gases may be valuable discriminators between fluids of mantle versus crustal versus meteoric origins. Helium isotope composition is a useful tool for tracing fluids, given that depleted upper mantle $^{3}$He/$^{4}$He ratios of 7 to 9 Ra (Ra = atmospheric ratio) and deep mantle plume-derived ratios of 9 to 32 Ra are very distinct from both the atmospheric ratio (1.0) and ratios of shallow crustal (0.01–0.05 Ra) hydrothermal fluids (Mamyrin and Tolstikhin, 1984; Azbel and Tolstikhin 1990; Graham et al., 1999; Kendrick et al., 2001). Furthermore, $^{40}$Ar/$^{36}$Ar ratios in excess of the atmospheric value of 295.5 indicate some degree of crustal or mantle enrichment of a hydrothermal fluid (e.g., Bohle and Irwin, 1992).

### Sampling and analytical procedure

We chose seven vein samples from orebodies 1 and 70 in the Dongguoping deposit for study of noble gas isotopes in the trapped inclusion fluids. Pyrite is known to be a suitable trap for noble gases (Stuart et al., 1994; Jean-Baptiste and Fouquet, 1996; Burnard et al., 1999), particularly because it retains helium over geologic periods, in contrast to quartz and feldspar (Mark Kendrick, writ. commun., 2002). We purposely selected pyrite separates from veins that had the fewest generations of postgold, secondary fluid inclusions trails, in case such fluids were also trapped in significant volumes in the pyrite in these veins. All the samples were prepared and analyzed at the Isotope Laboratory, Institute of Mineral Resources, Chinese Academy of Geological Sciences. The helium and argon isotope compositions were analyzed by bulk extraction of fluids using the invacuo crushing method. This method can reduce the effects from in situ-produced radiogenic isotopes and adsorbed atmospheric gases, overcoming problems due to the low blank. Also, the large fluid volumes, relative to laser extraction methods, allow for greater precision in the data. The pure pyrite separates, 0.2 to 0.8-mm diam, were first washed ultrasonically in distilled water, then in acetone, and finally oven dried. The online screw-type crushers were constructed from modified vacuo valves. Six samples, about 1 g each, were loaded into crushers at the same time. The samples were baked online at 100° to 150°C for 24 h to remove adsorbed atmospheric gases before crushing. It should be noted, however, that adsorbed argon is not entirely removed by these procedures and thus our argon measurements represent minimum estimates because of possible mixing with air with a $^{36}$Ar/$^{40}$Ar value of 295.5.

The gases released by crushing were exposed to a titanium sponge pump at 780°C and a zircon-aluminum pump at 300°C for 20 min to remove the active gases, such as H$_2$, N$_2$, O$_2$, CO$_2$, CH$_4$, and H$_2$O, and any organic substances. Then, the remaining gases were exposed to a second cold zircon-aluminum pump for 10 min and argon was condensed onto liquid N$_2$-cooled charcoal for 10 min. Comparatively pure helium travels directly into the analyzing system of the mass spectrometer. Trace impure gases, such as hydrogen, which entered the system along with the helium, were removed by a liquid N$_2$-cooled titanium sublimation pump located adjacent to the spectrometer ion source.

The helium isotope compositions were measured on a MI-1201H inert gas mass spectrometer made by Firm “SELMI” based in Sunny, Ukraine. A Faraday cup was used to measure $^4$He and $^4$He was measured on an electron multiplier. The mass resolution of 1,200 enables $^3$He to be distinguished from the HD-H$_3$ doublet peak. Argon was adsorbed from the charcoal filter and separated from xenon at ~75°C. The $^{36}$Ar was measured by a Faraday cup and $^{36}$Ar and $^{38}$Ar by an electron multiplier. A standard was measured before analyzing the samples and the results from the samples were normalized to the standard, which was made by purifying the air and

<table>
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<th>Sample no.</th>
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<th>Liquid phase (mol %)</th>
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<td>H$_2$S</td>
<td>CH$_4$</td>
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<td>Monophase CO$_2$ inclusion</td>
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<td>11.1</td>
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<td>3.5</td>
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<td>74.7</td>
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<td>Two-phase V-L H$_2$O inclusion</td>
<td>80.0</td>
<td>5.1</td>
<td>14.9</td>
</tr>
</tbody>
</table>

1 Instrument type is a JY Inc. RAMANORU 1,000-type laser Raman probe made in France.
periodically examining the air during its use. All the analytical results are based on the internationally accepted values for atmospheric $^{3}$He/$^{4}$He of 1.4 × 10⁻⁶, and $^{40}$Ar/$^{36}$Ar of 295.5. Helium results are reported as R/Ra ratios, where R is the $^{3}$He/$^{4}$He ratio of the sample and Ra is that of the air. The hot blank level of $^{4}$He was 2 × 10⁻¹¹ cm³ STP, for which $^{3}$He/$^{4}$He ratio of 1 × 10⁻⁶ was assumed, whereas the blank level of $^{40}$Ar was 5 × 10⁻⁹ cm³ STP/g. The effect of the blank on the measurement result was negligible. The analytical precision of standard gas was 1 percent and that of samples was included in Table 4.

Results

The pyrites analyzed were well-formed euhedral grains with no evidence of subsequent deformation and show a paragenesis with quartz and K-feldspar in the auriferous quartz veins and feldspar-rich alteration halos. Therefore, we are confident that the extracted fluids were related to mineralization and thus coeval with the unmixing CO₂-type and high-temperature aqueous-type fluid inclusions in the quartz that hosts the pyrite and gold.

The $^{3}$He/$^{4}$He ratios of pyrite in the ore veins range from 0.3 to 5.2 Ra (Table 4), all significantly higher than crustal values. The $^{3}$He/$^{4}$He value of fluid extracted from the pyrite of orebody 1 (2.1–5.2 Ra) is higher than that of orebody 70 (0.3–0.8 Ra). Although the cause for the difference is uncertain, we speculate that this reflects the abundant mineralization in altered wall rock in orebody 70, and the possibility of greater interaction between fluids and the granitoid host rock, which could have contributed radiogenically produced $^{4}$He to the circulating ore fluids. Despite the order of magnitude difference in isotope ratios between the orebodies, the ore fluids in both main ing ore fluids. Despite the order of magnitude difference in isotope ratios between the orebodies, the ore fluids in both main orebodies of the Dongping deposit are markedly enriched in $^{3}$He relative to typical crustal values. Their $^{3}$He/$^{4}$He ratio is 6 to 100 times higher than that of fluids typical of a crustal origin (Manyr and Tolstikhin, 1984; Azbel and Tolstikhin 1990), which requires helium input from a mantle source into the ore fluids. Many syngenetic and epigenetic metallic ore deposits, including porphyry copper and orogenic gold deposits, have been reported as showing involvement of mantle-derived components based on helium isotope studies (Simmons et al., 1987; Stuart et al., 1995; Hu et al., 1998; Burnard et al., 1999; Kendrick et al., 2001; Mao et al., 2002a). All of the $^{3}$He/$^{4}$He ratios in these other deposits are between crustal and mantle end members, similar to what is seen at the Dongping deposit, implying variable degrees of mixing between mantle and crustal helium somewhere along the flow path.

The $^{40}$Ar/$^{36}$Ar ratios range from 309 to 632 for the Dongping fluid inclusions and are all only slightly higher than that of atmospheric argon. In contrast to the helium, it is impossible to distinguish radiogenic (crustal) argon from mantle argon, as both are characterized by high $^{40}$Ar/$^{36}$Ar ratios (Xu et al., 1995). The $^{40}$Ar/$^{36}$Ar ratios, therefore, are not necessarily diagnostic of a mantle contribution. As stated previously, perhaps significant adsorbed argon is present. The $^{40}$Ar/$^{36}$Ar ratio of samples from orebody 70 is higher than that of samples from orebody 1 (Table 4), and it is plausible that some potassium within the K-feldspar in orebody 70 has decayed to add excess $^{40}$Ar into some fluid inclusions. Alternatively, the relatively low $^{40}$Ar/$^{36}$Ar ratios could indicate variable interaction with air-saturated water, which is not inconsistent with some of the measured light $\delta^{18}$O measurements on feldspar from this deposit (e.g., Hart et al., 2002)

Discussion and Conclusions

Summary of PTX of ore-forming fluids

Observed features suggesting CO₂ and H₂O immiscibility at the Dongping deposit are in agreement with the criteria of Roedder (1984) and Zhang and Chen (1993). Inclusions with different degrees of fill of CO₂ and H₂O appear to be coeval, and fluid inclusions show a wide range in phase ratios extending to almost pure CO₂ and H₂O inclusions. In addition, the modes of homogenization of CO₂- and H₂O-type inclusions are different (the former homogenizing to the CO₂ phase and the latter to the H₂O phase), but the homogenization temperatures of both are similar. Finally, the pressures of CO₂- and H₂O-type inclusions were identical during their total homogenization.

Because CO₂–H₂O immiscibility was characteristic of ore deposition at the Dongping deposit, the fluid inclusion homogenization temperatures reflect the true trapping temperatures of the ore-forming fluids. The CO₂-type inclusions homogenized to a vapor at temperatures of 255° to 372°C and to a liquid at 250° to 365°C. Similarly, the homogenization temperatures of the H₂O-type inclusions are nearly identical, ranging from 250° to 380°C. The calculated pressures at these temperatures for the CO₂-type inclusions are 0.6 to 1 kbar and for the H₂O-type inclusions are 0.4 to 1 kbar. Ore fluids have low salinities (5–7 wt % NaCl equiv), with little noticeable difference between the CO₂- and H₂O-rich unmixed fluids (Fig. 10). With decreasing temperatures of the hydrothermal fluids, from 380° to as low as 250°C, the degree of unmixing increases until almost pure CO₂ and H₂O

Table 4. He and Ar Isotope Data of Pyrite Separates from the Ore-Bearing Veins and K-Feldspar-Rich Granite Whole-Rock Samples from near the Dongping Gold Deposit

<table>
<thead>
<tr>
<th>Sample no.</th>
<th>Minerals or rocks</th>
<th>$^{3}$He/$^{4}$He × 10⁻⁶</th>
<th>$^{4}$He × 10⁻⁸ (cm³ STP/g)</th>
<th>R/Ra</th>
<th>$^{40}$Ar/$^{36}$Ar × 10⁻⁶ (cm³ STP/g)</th>
<th>$^{40}$Ar/$^{36}$Ar</th>
</tr>
</thead>
<tbody>
<tr>
<td>DP1-1</td>
<td>Pyrite</td>
<td>7.2 ± 0.5</td>
<td>21.5</td>
<td>5.2</td>
<td>20.3</td>
<td>309 ± 1</td>
</tr>
<tr>
<td>DP1-2</td>
<td>Pyrite</td>
<td>3.0 ± 0.4</td>
<td>4.0</td>
<td>2.1</td>
<td>6.9</td>
<td>375 ± 2</td>
</tr>
<tr>
<td>DP1-3</td>
<td>Pyrite</td>
<td>2.9 ± 0.8</td>
<td>1.5</td>
<td>2.1</td>
<td>7.5</td>
<td>476 ± 7</td>
</tr>
<tr>
<td>DP1-4</td>
<td>Pyrite</td>
<td>5.8 ± 1.7</td>
<td>0.7</td>
<td>4.2</td>
<td>5.2</td>
<td>483 ± 9</td>
</tr>
<tr>
<td>DP1-7-1</td>
<td>Pyrite</td>
<td>6.9 ± 1.6</td>
<td>0.5</td>
<td>5.0</td>
<td>5.5</td>
<td>431 ± 6</td>
</tr>
<tr>
<td>DP7-4</td>
<td>Pyrite</td>
<td>1.1 ± 0.4</td>
<td>4.6</td>
<td>0.8</td>
<td>1.9</td>
<td>525 ± 7</td>
</tr>
<tr>
<td>DP7-6</td>
<td>Pyrite</td>
<td>0.4 ± 0.1</td>
<td>14.9</td>
<td>0.3</td>
<td>2.3</td>
<td>632 ± 39</td>
</tr>
</tbody>
</table>

Note: 2σ errors referred to the last digits are given for $^{3}$He/$^{4}$He and $^{40}$Ar/$^{36}$Ar.
inclusions are trapped. The few highly saline inclusions, which contain the daughter salts, perhaps represent a second trapped hydrothermal fluid generation. As these inclusions are relatively rare, their relationship to the gold ores and the main generation of heterogeneously trapped fluids is uncertain. It can not be ruled out, however, that such fluids were generated by phase separation and concentration of the salts in the aqueous phase (e.g., Robert and Kelly, 1987). The combined fluid inclusion microthermometric and laser Raman spectroscopy analyses indicate that the composition of fluids forming the ores of the Dongping deposit was mainly CO\textsubscript{2} and H\textsubscript{2}O with lesser, but significant, H\textsubscript{2}S and N\textsubscript{2}. The fluids have a low salinity (5–7 wt % NaCl equiv) and, in general, are typical of those reported for orogenic gold systems (e.g., Goldfarb et al., 1998; Groves et al., 1998) and many intrusion-related gold deposits (e.g., Thompson and Newberry, 2000). The total CO\textsubscript{2} content in the fluids is 4 to 95 but mainly 20 to 60 mole percent. The total density of the fluids mainly ranges from 0.482 to 0.794 g/cm\textsuperscript{3}. The bulk parent fluid composition is not known but was probably somewhere near 50 percent H\textsubscript{2}O and 50 percent CO\textsubscript{2} ± H\textsubscript{2}S + N\textsubscript{2}.

Sources of ore fluids

The fluid inclusion data from the Dongping deposit are not definitive of any distinct fluid source. Ore fluids of low salinity and enriched in CO\textsubscript{2} have been defined by some workers as mainly magmatic (e.g., Higgins and Kerrich, 1982; Burrows et al., 1986; Burrows and Spooner, 1987; Cameron and Hattori, 1987; Wake and Taylor, 1988; Jiang et al., 1999) and by others as metamorphic (e.g., Kerrich and Fyfe, 1981; Goldfarb et al., 1988; Kerrich, 1991). Schmidt et al. (1997) went so far as to call the highly carbonic fluids trapped in inclusions in veins of the Ashanti gold fields, Ghana, a new fluid type of unclear origin. Sulfur, oxygen, and carbon isotope data from the Dongping deposit also do not discriminate between either a magmatic or metamorphic fluid source (Song and Zhao, 1996; Hart et al., 2002). Newton et al. (1980), Groves et al. (1988), and Mao et al. (2000, 2002a) additionally suggested such low-salinity, CO\textsubscript{2}-rich compositions to be consistent with fluids derived from mantle degassing or, at least, indicative of some participation of mantle fluids in the mineralization process.

Geologic constraints cannot rule out either magmatic, metamorphic, or mantle sources for the fluid components. In the North China craton, regional metamorphism of basement rocks took place in the Late Archean or Early Proterozoic (Lu et al., 1996). Thus, much of the regional metamorphism is obviously far older than the ca. 150 Ma mineralization. Although the Dongping deposit is hosted by the 390 Ma alkalic magmatic complex, there is a time gap of more than 200 m.y. between the alkaline magmatism and the gold mineralization. However, the mineralization does overlap with the Late Jurassic-Early Cretaceous magmatism and with metamorphism of younger rock units within the Yanshan belt. Both processes could have theoretically sourced fluids that ultimately were focused into the Dongping area.

The Dongping deposit overlies the 100-km-wide, north-south gravity lineament, a feature that divides the North China craton and parallels the Tan-Lu fault system, a San Andreas-like continental margin strike-slip structure that is located 400 km to the east. The regional significance of the north-south gravity lineament has been stressed by Griffin et al. (1998). It separates a region of thinner crust, shallow asthenosphere, and high heat flow to the east from a part of the craton to the west that still maintains its thick, cool Archean lithospheric keel. Griffin et al. (1998) suggest that the Jurassic-Cretaceous

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**Fig. 10.** Influence of pressure on the CO\textsubscript{2}- and H\textsubscript{2}O-rich two-phase immiscible region in the NaCl-CO\textsubscript{2}-H\textsubscript{2}O system with a salinity of 6 wt percent NaCl equiv (after Bowers and Helgeson, 1983).
oblique subduction of the Kula plate beneath eastern Asia initiated rifting and subsequent dispersion of most of the old lithospheric mantle beneath the eastern part of the craton. The upwelling of fertile young mantle to depths of as shallow as 50 km certainly would represent some of the most favorable conditions under which mantle fluids could migrate into shallow crustal regions. Therefore, a mantle source is also very possible for the high fluid flux in the Dongping area (e.g., Deng et al., 1999).

The noble gas data allow us to further evaluate these three possible fluid sources. As shown by the results of helium isotope analyses (Table 4), the ore fluids in the Dongping gold deposit must contain a mantle component, as their $^{3}$He/$^{4}$He ratio is about ten to one hundred times higher than that of crustal sources. Assuming pure mantle helium to have a value of 8 Ra (Tolstikhin, 1978), fluid inclusions in pyrite from the Dongping orebody 1 would contain approximately 26 to 65 percent mantle helium, whereas the proportion in the more disseminated ores of orebody 70 would be 4 to 10 percent. However, as suggested by Kendrick et al. (2001), these probably are minimum estimates for the original fluids because crustal contamination and magmatic aging may lower the Ra value for mantle melts. Most importantly, there is no way to rule out the presence of significant amounts of mantle helium in the ore-forming fluids, although O, H, and S stable isotopes also imply important crustal fluid components (Zhang and Mao, 1995; Song and Zhao, 1996; Nie et al., 1998). In fact, the stable isotopes alone show no evidence of mantle contributions.

The new fluid inclusion and rare gas data from our study are similar to those from other deposits along the margins of the North China craton. The high $^{4}$He/$^{3}$He ratios are analogous to those (3–7.2 Ra) in carbon dioxide gas reservoirs in a subsiding basin developed along the Tan-Lu fault in eastern China (Xu et al., 1995) and are similar to those (3.5–9.8 Ra) from the 120 Ma Wangu orogenic gold deposit adjacent to the Tan-Lu fault (Mao et al., 2002a). Analyses of inclusion fluids in pyrite from auriferous and barren quartz veins in the Wangu deposit (Mao et al., 2002a) are separated into two rare gas groups, which are interpreted as having been dominantly sourced in the mantle and crust, respectively (Fig. 11). A similar distinction between the helium ratios for pyrite from the auriferous veins versus the K-feldspar–altered wall rocks at Dongping is interpreted to suggest the same sources, with the lower value for the latter crustal group perhaps indicating some mixing with meteoric water. Such would be consistent with the light $\delta^{18}O$ values (e.g., ~1.7 to +2.2‰) from ore-related K-feldspar (Hart et al., 2002).

Fluid inclusion PTX characteristics of the Dongping deposit generally overlap those of other Late Jurassic–Early Cretaceous gold deposits elsewhere along the northern margin of the craton (Hart et al., 2002), within the Jiaodong peninsula on the eastern margin (Qin et al., 2002) and along the southern margin in the Xioaqinling gold belt (Xu et al., 1998). The Dongping deposit, with trapping pressures estimated between 0.6 to 1.0 kbar, may have formed at shallower crustal levels than many of the other deposits. Also, the relatively low methane concentrations in the fluid inclusions from the granitoid-hosted Dongping deposit contrasts with much higher concentrations where the gold ores are hosted in Precambrian basement rocks (e.g., Xiaojinling belt of Xu et al., 1998; Mao et al., 2002b). This may reflect either a redox influence on the ore fluids by the metasedimentary country rocks and/or local addition of methane from these rocks.

Proposed mineralization model

Bonham (1986) proposed that telluride-rich gold deposits are essentially epithermal deposits. Some gold-bearing telluride epithermal deposits, such as Cripple Creek, Colorado (Thompson et al., 1985; Bonham, 1986), Emperor, Fiji (Ahmad et al., 1987), Macassa, Abitibi belt, Canada (Tsefaye, 1992), and Golden Sunlight, Montana (Spy et al., 1996), are genetically related to alkaline complexes; others, such as the Major's Creek deposit, Australia (Wake and Taylor, 1985) and some deposits in the Kuril island arc, Russia (So et al., 1995), are related to shallow granites and their associated volcanism. Fluid inclusion temperature and pressure estimates at Dongping are not, however, typical of ore deposits formed in an epithermal environment. Rather, we agree with the classification of Sillitoe (2002) that the Dongping deposit is similar in features to many deeper orogenic and intrusion-related gold deposits. We disagree, however, with any genetic association to alkaline rocks (e.g., Song, 1990, 1991; Song and Zhao, 1996), as geochronology now clearly indicates a ca. 250 m.y. time gap between alkaline magmatism and gold deposition.

The requirement of a mantle source area for at least a part of the ore fluid at Dongping indicates that further refinement of the ore genesis model for gold deposits throughout the North China craton and elsewhere is still necessary. Strict application of the metamorphic devolatilization model that
characterizes orogenic gold deposits or the magmatic exsolution model for intrusion-related gold systems does not explain the mantle-related helium at Dongping. The rare gas data from this study require some degree of mantle input to the hydrothermal system. Ore-related δ13C values of –4.3 to –4.0 per mil (Nie, 1998) are also consistent with mantle-derived carbon, although these latter data may not be uniquely diagnostic (Kysers and Kerrich, 1990). Lead isotope data of Nie (1998) also suggest a mixing of mantle and crustal lead in pyrite and galena from the Dongping deposit. Whereas the crustal component is likely Archean granulite, the source of the mantle lead could be either the host Paleozoic alkaline igneous rocks or the Mesozoic mantle. Lead isotope data for hydrothermal K-feldspar (Nie, 1998) are much more consistent with a mainly mantle and/or lower crust source, and we suggest that it is possible that the significant volume of added potassium may also be derived from a deeply sourced fluid.

The Dongping data warrant a reconsideration of the granulitization model for lode gold genesis, as was described by Colvin et al. (1988) and Cameron (1988). The upwelling of shallow mantle beneath the North China craton, at approximately the ca. 150 Ma time of gold formation, is consistent with such a model that releases heat and volatiles into the lower crust from ponded melts in the underlying mantle. Lithophile element enrichment within the fluid phase, as supported by the extensive potassium metasomatism at Dongping, would also be expected during such granulitization. The actual melting of the shallow Jura-Cretaceous mantle would have been triggered by devolatilization of the subducting Kula Pacific margin slab beneath the thinned eastern Asian crust. Localization of the Dongping deposit along the margin of the area of lithospheric thinning suggests that released mantle volatiles were focused into areas of contrasting thickness in the subcontinental lithosphere and then into the poorly mapped regional features at shallower crustal levels. Future work on the North China lode gold deposits needs to address the significance of the mantle rare gases; do they indicate significant fluid contributions from the mantle or just a minor mantle influx into a mainly crustally derived hydrothermal fluid?

Metamorphic and/or magmatic fluid components, perhaps products of the N-S-directed Yanshan deformation, were added into the hydrothermal system along this upward flow path, as required by the stable and other radiogenic isotope ratios. A metamorphic contribution is preferable because a CO2-rich carbonic-aqueous fluid will only exsolve from a magma at pressures above about 3 kbars (Eggl er and Kadik, 1979), and the Dongping deposit formed under much lower pressures. The rare, highly saline fluid inclusions of unknown significance, however, may indicate the influx of a second hydrothermal pulse, which is of magmatic origin but unrelated to the main ore-forming system.

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