The Xiajinbao gold deposit is located in the Yanshan District of the North China Craton. The geology of the Xiajinbao deposit consists of Archean gneiss, Proterozoic sedimentary rocks, granite porphyry, quartz porphyry dikes, and diorite dikes. The diorite contains a large quantity of magnetite. The granite porphyry and quartz porphyry dikes, which contain trace amounts of ilmenite without magnetite, have high aluminum saturation index values and a low magnetic susceptibility. These rocks are classified as ilmenite series granitic rocks. The pyrite quartz veins (Stage I), pyrite-chalcopyrite quartz veins (Stage II), and sphalerite-galena quartz veins (Stage III) of the Xiajinbao deposit were hosted in the granite porphyry. Major gold mineralization was associated with the sphalerite-galena quartz veins. The homogenization temperature and salinity of the sphalerite-galena quartz veins are 220 °C in a mode and 9.1–13.8 wt%, respectively. The sulfur fugacity of the Xiajinbao deposit ranged 10^{−7.8}–10^{−11} and 10^{−9.5}–10^{−14.3} atm for the pyrite-chalcopyrite quartz veins and sphalerite-galena quartz veins, respectively. The formation environment of the sphalerite-galena quartz veins of the Xiajinbao deposit is similar to that of zoned base metal veins associated with oxidized magma. The oxygen and hydrogen isotope ratios of the hydrothermal solution in equilibrium with quartz from the sphalerite-galena quartz veins range from +3.3 to +5.5 and −38 to −41‰, respectively, suggesting magmatic water. The range of sulfur isotope ratios of pyrite and galena from the sphalerite-galena quartz veins is from −0.2 to +3.6‰. These sulfur isotope ratios are in the range of ratios for ores formed by igneous activity associated with oxidized magma. Therefore, the gold mineralization of the Xiajinbao deposit is inferred to be the result of igneous activities from oxidized magmas, such as the diorite dikes, with the granite porphyry forming the host rock of the sphalerite-galena quartz veins of the Xiajinbao deposit.

Keywords: Xiajinbao, Gold, Fluid inclusion, Oxygen, Sulfur, Isotope, Oxidized magma
gested that the gold mineralization was related to igneous rocks in the late Mesozoic (Mao et al., 2005a, 2005b; Zhang et al., 2007; Zhang et al., 2013).

The gold deposits in the Paleozoic and Mesozoic intrusive rocks in the Yanshan District have been classified into three groups based on the host rocks: (1) gold deposits in felsic intrusions, including calc-alkaline granitic intrusions (e.g., the Yu’erya, Anjiayingzi, Linglong, and Jiaojia deposits) and breccia pipes (e.g., the Chenjiazhangzi and Qiyugou deposits), (2) gold deposits in alkaline intrusions (e.g., the Wulashan, Donghuofang, Dongping, Hougou, and Guilaizhuang deposits), and (3) gold deposits in ultramafic intrusions (e.g., the Jinjiahuang gold deposit) (Nie, 1998; Nie et al., 2004). The Xiajinbao deposit is one of the gold deposits hosted in alkaline intrusions in the northeastern part of the Yanshan District (Fig. 2). Gold mineralization in the Xiajinbao mining area mainly occurs in the granite porphyry, in addition to some quartz porphyry dikes and diorite dikes. Previous research in the Xiajinbao deposit concentrated on the chemical composition, age dating, and evolution of the granite porphyry. The age of granite porphyry, determined by the fission track method on apatite and zircon, is estimated to be 103–154 Ma (Yuan et al., 1999). The age for the granite porphyry and quartz porphyry dikes, determined by the zircon U-Pb method, is 158–163 Ma (Li et al., 2016; Zhang et al., 2016; Zou et al., 2016). Ilmenite–series granitic rocks and diorite are present in the Xiajinbao mining area.

Previous research in the Xiajinbao deposit thought that there was an intimate relationship between gold mineralization and the granite porphyry (Tang et al., 2009; Ye, 2013; Zhang et al., 2017); however, the role of diorite in mineralization has not been discussed. Therefore, there is no consensus on the relationship between magmatism and gold mineralization. Here the geological and geochemical characteristics of the Xiajinbao gold deposit are described to clarify this relationship. The results of this study are useful for understanding the ore-forming process of gold mineralization in the North China Craton and providing insights for future exploration.

GEOLOGICAL SETTING

Regional geology

The North China Craton comprises eastern and western blocks that were amalgamated by a major collisional event between 1.8 and 1.9 Ga (Zhao et al., 2001; Yang et al., 2014). The geology of the Yanshan District, including the Xiajinbao deposit, is composed of Archean base-
ment rocks overlain by Mesoproterozoic, Neoproterozoic, Phanerozoic, and Cenozoic strata in ascending order (Fig. 2). The Archean basement consists of tonalite–trondhjemite–granodiorite gneisses, migmatite, amphibolite, ultramafic rocks, and banded iron formations (Zhao et al., 2001; Santosh, 2010; Santosh et al., 2013; Zhai, 2014). The Mesoproterozoic and Neoproterozoic strata consist of medium to coarse–grained quartz sandstone, shale, dolomite, and quartzite in ascending order. The Phanerozoic strata are mainly Mesozoic sedimentary and volcanic rocks. There is an unconformity between the Mesoproterozoic and Mesozoic strata. The Mesozoic strata consist predominantly of Jurassic volcanic rocks, such as andesitic lava and breccia. The Cenozoic strata are mainly Quaternary sediments. The Mesozoic intrusive rocks in the Yanshan District consist of large amounts of felsic intrusive rocks with subordinate intermediate and mafic intrusive rocks. The number of intrusive rocks decreases from the northeast to the central part of the district. The age of intrusive rocks also becomes younger from the northeast to the central part of the district.

The gold deposits in the Yanshan District are distributed along the east–west trending faults (Fig. 2) that are subparallel to the contact between the North China Craton and Central Asian Orogen (Liu et al., 2016). Gold deposits in the district mainly occur in two different host rocks: Archean metamorphic rocks and granitoids. The Xiajinbao deposit is one of the gold deposits hosted in granitic and porphyritic rocks.

Geology of the Xiajinbao mining area

The geology of the Xiajinbao mining area consists of Archean gneiss, Proterozoic sandstone, dolomitic limestone, dolomite, Quaternary sediments, Jurassic granite porphyry, quartz porphyry dikes, and diorite dikes. Archean gneiss and Proterozoic sandstone and shale are present in the northeastern and northern parts of the Xiajinbao mining area, respectively (Fig. 3). Proterozoic limestone and dolomite are distributed across the southern and eastern parts of the mining area (Figs. 3 and 4e). Quaternary sediments cover the central part of the mining area. Intrusive rocks in the Xiajinbao mining area are granite porphyry, quartz porphyries, and diorite dikes. The granite porphyry is the host rock of the Xiajinbao gold deposit, which is distributed across the central part of the mining area (Fig. 3). Some diorite dikes and quartz porphyry dikes are present in the granite porphyry and Proterozoic sedimentary rocks, composed of alternating shale and sandstone in the southern part of the granite porphyry (Figs. 4a, 4b, and 4d). Some diorite dikes cut the granite porphyry in the Xiajinbao deposit (Fig. 5). The quartz porphyry also cuts the diorite in the Xiajinbao mining area (Fig. 4b). Based on the observed cross-cutting patterns, the sequence of emplacement of igneous rocks in the Xiajinbao mining area is inferred to begin with the granite porphyry, followed by the diorite dikes, and then the quartz porphyry dikes.

A large-scale fault (F4), having a strike and dip ranging 40°–70°NE and 40°–65°SE, respectively, is present in the Xiajinbao mining area. The F4 fault is inferred to have formed after the emplacement of the granite porphyry, because there was no mineralization in the fault zone.

Characteristic features of igneous rocks in the Xiajinbao deposit

The granite porphyry is reddish in color and has a porphyritic texture. The granite porphyry consists of quartz, orthoclase, plagioclase, and biotite phenocrysts, with small amounts of zircon, rutile, and apatite (Fig. 6a). The opaque minerals are mainly pyrite and small amounts of ilmenite. The quartz and plagioclase phenocrysts have a subhedral shape. The matrix of the granite porphyry is composed of fine–grained quartz and feldspar. A small amount of illite replacing plagioclase is present as altered minerals in the granite porphyry (Fig. 6b). The strongly altered granite porphyry near mineralized veins is characterized by a mineral assemblage of quartz and illite.

A quartz porphyry dike, having a gray color, consists of a large amount of quartz with small amounts of plagioclase and orthoclase (Figs. 6c and 6d). The matrix of the quartz porphyry consists of cryptocrystalline felsic minerals.
The outcropping and underground diorite dikes of the Xiajinbao deposit are dark gray in color. The diorite shows a massive structure and a hypidiomorphic granular texture (Fig. 6e). The mineral assemblage of diorite consists of large amounts of plagioclase and amphibole with small amounts of biotite and apatite (Fig. 6f). The size of the plagioclase in the underground diorite dike is smaller than that in the outcropping dike (Fig. 6g). The opaque minerals in the diorite dike are magnetite, hematite, and pyrite. Some magnetite veins occur in the outcropping diorite dike (Fig. 4c). The alteration minerals chlorite, dolomite, and illite are present in some parts of the diorite dike. The underground and outcropping diorite dikes are both strongly altered (Fig. 6h).

**Figure 4.** (a) Cross-cutting relationships between the diorite dikes and quartz porphyry dikes. (b) The mode of occurrence of the quartz porphyry dike cutting the diorite dike. (c) The presence of a magnetite vein in the diorite dike. (d) Quartz porphyry dike in altered Proterozoic sandstone and shale. (e) Proterozoic sandstone and shale.

**Figure 5.** Diagram showing the A-B cross-section in Figure 3 of the Xiajinbao gold deposit. The gold assay data are also shown. The units of the assay data are ppm.

The outcropping and underground diorite dikes of the Xiajinbao deposit are dark gray in color. The diorite shows a massive structure and a hypidiomorphic granular texture (Fig. 6e). The mineral assemblage of diorite consists of large amounts of plagioclase and amphibole with small amounts of biotite and apatite (Fig. 6f). The size of the plagioclase in the underground diorite dike is smaller than that in the outcropping dike (Fig. 6g). The opaque minerals in the diorite dike are magnetite, hematite, and pyrite. Some magnetite veins occur in the outcropping diorite dike (Fig. 4c). The alteration minerals chlorite, dolomite, and illite are present in some parts of the diorite dike. The underground and outcropping diorite dikes are both strongly altered (Fig. 6h).

**ORE DEPOSIT**

**Mode of occurrence of orebodies**

The orebodies of the Xiajinbao deposit mainly occur in an 800 × 300 m area within the granite porphyry (Fig. 3), and are composed of aggregates of a large number of veinlets and veins having widths ranging from 1 mm to 5 cm. These veins and veinlets are divided into three types: pyrite quartz veins, pyrite–chalcopyrite quartz veins, and sphalerite–galena quartz veins (Figs. 7a, 7d, and 7g). The general strikes and dips of these veins are 30°–40°NE–SW and 40°–80°NW, respectively. The pyrite–chalcopyrite quartz veins are dominant in the deep part of the orebodies, whereas the sphalerite–galena quartz veins are dominant in the shallow part of the orebodies. The thickness of the bonanza-grade orebodies, composed of the aggregates of these veins, ranges 1–15 m. The average Au grade in the whole Xiajinbao deposit is 1.9 g/t, and it is 7.5 g/t for the bonanza-grade orebodies. The part having the highest Au grade (222 ppm) is in a sphalerite–galena quartz vein–dominant area in the shallow part of the Xiajinbao deposit.
Diorite dikes are also present near the Au-rich part in the underground workings of the Xiajinbao deposit (Fig. 5). The distribution of diorite dikes appears to correspond to the distribution of these Au-rich areas in the orebodies.

Mineral assemblages and paragenetic sequence of pyrite quartz veins, pyrite-chalcopyrite quartz veins, and sphalerite-galena quartz veins

Pyrite quartz veins, pyrite-chalcopyrite quartz veins, and sphalerite-galena quartz veins were observed in the Xiajinbao deposit (Figs. 7a, 7d, and 7g). The sphalerite-galena quartz veins contain a large amount of gold. The pyrite quartz veins are cut by the pyrite-chalcopyrite quartz veins. The sphalerite-galena quartz veins cut both the pyrite quartz veins and pyrite-chalcopyrite quartz veins (Fig. 7d). Based on these observed cross-cutting patterns, the sequence of mineralization is inferred to have first occurred in the pyrite quartz veins (Stage I), then the pyrite-chalcopyrite quartz veins (Stage II), and finally the
sphalerite–galena quartz veins (Stage III). The mineral assemblages of these veins are summarized in Figure 8.

The pyrite quartz veins are composed of large amounts of quartz and pyrite with small amounts of chalcopyrite and electrum (Fig. 7a). The pyrite crystals mostly have a subhedral shape, with diameters ranging 0.1–2 mm (Fig. 7b). The chalcopyrite, which possesses an anhedral shape, infills the fracture of the subhedral pyrite (Fig. 7c). The electrum in this stage was observed in the interstices between the quartz and pyrite crystals (Fig. 7b).

The pyrite–chalcopyrite quartz veins are divided into two substages. Large amounts of quartz, pyrite, and chalcopyrite formed in earlier substage of stage II. Small amounts of chalcopyrite, galena, and electrum formed in the pyrite fractures in the latter substage of stage II. Electrum is also present in pyrite fractures and bismuth are also observed in the pyrite fractures in this latter substage, as well as in the sphalerite–galena quartz vein. It is thus inferred that this mineral assemblage formed during mineralization in stage III.

The sphalerite–galena quartz veins are divided into three substages. Large amounts of quartz, chalcopyrite, galena, and sphalerite, as well as small amounts of pyrite and electrum, formed in the earliest substage of stage III. Large amounts of quartz, galena, and sphalerite, with small amounts of pyrite and chalcopyrite, formed in middle substage of stage III. Small amounts of quartz, chal-

Figure 7. Pictures of a pyrite quartz vein, a pyrite–chalcopyrite quartz vein, and a sphalerite–galena quartz vein, and photomicrographs of the ore minerals in these veins. (a) Pyrite quartz vein. (b) Electrum, between the pyrite crystals. (c) Network of chalcopyrite in pyrite crystals in a pyrite quartz vein. (d) Mode occurrence showing a pyrite–chalcopyrite quartz vein that is cut by a sphalerite–galena quartz vein. (e) A galena–chalcopyrite–tetrahedrite–native bismuth veinlet in a pyrite crystal of pyrite–chalcopyrite quartz vein. Electrum is also present in pyrite. (f) Galena, tetrahedrite, and electrum veinlets in a pyrite crystal in a pyrite–chalcopyrite quartz vein. (g) Sphalerite–galena quartz vein. (h) Photomicrograph showing the mode of occurrence of a chalcopyrite veinlet cutting a pyrite crystal, sphalerite, and galena crystals in a sphalerite–galena quartz vein. (i) Mode occurrence of electrum in a sphalerite–galena quartz vein. Abbreviations: Py: pyrite; El: electrum; Cp: chalcopyrite; Sph: sphalerite; Gn: galena; Tet: tetrahedrite; Bi: native bismuth.
copyrite, galena, native bismuth, tetrahedrite, and electrum precipitated into the fractures of other minerals in the latter substage of stage III. Two types of electrum were observed in this stage. One type of electrum occurs with galena, chalcopyrite, native bismuth, and tetrahedrite as veinlets in the pyrite crystals in the latter substage of stage III (Fig. 7f), whereas the other type of electrum occurs as isolated anhedral grains in the sphalerite crystals in the earliest substage of stage III (Fig. 7i). The electrum formed in the latter substage has a lighter yellow color than that formed in the earliest substage.

Wall–rock alteration

Wall–rock alteration of the Xiajinbao deposit is characterized by quartz–illite–pyrite alteration with the subordinate presence of fluorite and carbonate minerals. Quartz–illite–pyrite alteration prevails from the deep to shallow parts of the orebody of the Xiajinbao deposit. The amount of illite in the altered rocks of the sphalerite–galena quartz veins is larger than the amount of illite in the altered rocks of the pyrite quartz veins and pyrite–chalcopyrite quartz veins. The color of the altered granite porphyry near the sphalerite–galena quartz veins changes from reddish to green according to the increasing degree of alteration. Silicification was also observed along these veins.

ANALYTICAL METHODS

Chemical analyses of the major elements in the granite porphyry, quartz porphyry, and diorite were carried out using a Panalytical Axios Max X-ray fluorescence analyzer at the Analytical Laboratory of Beijing Research Institute of Uranium Geology. The detection limit of the chemical analysis for major elements was 0.1 wt%.

The sphalerite chemical compositions were analyzed with the JEOL JXA–733 Electron Probe Microanalyzer (EPMA) at Akita University. The elements analyzed were Zn, Fe, Mn, Cd, Cu, In, Ga, and S. The acceleration voltage was 20 kV, the beam current was 20 nA, and the diameter of the electron beam was 3–5 µm. The characteristic X-rays for concentration measurements were ZnKα, FeKα, MnKα, CdLα, CuKα, InLα, GaKα, and SKα. The standards for these elements were Zn metal, CuFeS2, MnS, GaAs, Cd, and InSb, respectively. All of the data were corrected with a ZAF–metal matrix correction program (Sweatman and Long, 1969). The detection limit was 0.1 wt%.

The homogenization temperature and salinity of the fluid inclusions were measured using the Linkam THMSG600 cooling–heating stage at the Key Laboratory of Nonferrous Metal Metallogenic Prediction of Chinese Ministry of Education of Central South University. The measurement temperature ranges were from room temperature to 600 °C for heating and from room temperature to −50 °C for cooling. The analytical precision of temperature measurements for the heating and cooling experiments were ±1 and ±0.1 °C, respectively. Salinity was calculated on the basis of the final ice melting temperature and solubility of NaCl in water (Bodnar, 1993).

The sulfur isotope ratios of pyrite and galena that coexist in the sphalerite–galena quartz veins were determined using a MAT–251 mass spectrometer at the Analytical Laboratory of Beijing Research Institute of Uranium Geology. The δ34S values are given relative to the Canyon Diablo Troilite isotope ratio standards. The analytical precision was ±0.2‰.

Oxygen isotope ratios of the quartz and bulk samples of the altered rocks and hydrogen isotopic ratios of illite separated from the altered rocks were determined using a mass spectrometer (Iso-Prime) at Akita University. The method used for analysis of the oxygen isotope ratios was the F2–technique (Kita and Matsubaya, 1983). The δ18O values are given relative to Standard Mean
Ocean Water (SMOW). The analytical precision of oxygen isotope ratios was ±0.2‰. The illite was separated from the altered rocks by elutriation for determination of the hydrogen isotope ratios. The purity of illite was confirmed by X-ray diffraction analysis. All of the samples were composed of a large quantity of illite and a small amount of quartz. Those samples were used to measure the hydrogen isotope ratios of illite, following the method described by Marumo et al. (1995). The analytical precision of the hydrogen isotope ratios was ±1‰.

### CHARACTERISTIC FEATURES OF THE IGNEOUS ROCKS

The chemical compositions of the igneous rocks of the Xiajinbao deposit are shown in Table 1. The SiO₂ and TiO₂ concentrations of the granite porphyry are around 70 and 0.2 wt%, respectively (Fig. 9). In the granite porphyry, concentration ranges are: the Al₂O₃, 13.9-14.3 wt%; the TiO₂, 1.9-2.2 wt%; and the Na₂O and K₂O, 2.2-2.5 wt%, respectively.

In the diorite dikes, concentration ranges are: the SiO₂ and Al₂O₃, 49.1-51.1 and 13.7-16.3 wt%, respectively (Fig. 9); the TiO₂, 1.0-3.1 wt%; the Al₂O₃, 13.7-14.3 wt%; the Na₂O and K₂O, 2.2-2.5 wt%, respectively. On a Harker diagram (Fig. 9), the K₂O concentration of the granite porphyry is higher than that of the quartz porphyry dikes, while the Na₂O concentration of the granite porphyry is lower than that of the quartz porphyry dikes.

To confirm the degrees of alteration for the granite porphyry, diorite, and quartz porphyry dikes, the alteration index (AI) and chlorite-carbonate-pyrite index (CCPI) (Ishikawa et al., 1976; Large et al., 2001) were determined (Fig. 10). The AI and CCPI values of the granite porphyry range 52.7-66.1 and 23.4-26.9, respectively. Some of the granite porphyry has slightly higher AI values due to alteration. However, most of the granite porphyry values plot in the least altered box. The AI and CCPI values of the outcropping diorite dikes are 39.2 and 72.3, respectively, while the AI and CCPI values of

| Sample  | Z153 | Z144-2 | Z176-2 | Z190-1 | Z188 | Z175-3 | 132-2 | Z45 | 132-1 |
|---------|------|--------|--------|--------|------|--------|-------|-----|-------|
| wt%     |      |        |        |        |      |        |       |     |       |
| SiO₂    | 71.0 | 70.4   | 70.1   | 71.0   | 70.8 | 70.9   | 75.8  | 51.1| 49.1  |
| TiO₂    | 0.2  | 0.2    | 0.2    | 0.2    | 0.2  | 0.2    | 0.1   | 1.0 | 3.1   |
| Al₂O₃   | 14.1 | 14.3   | 14.3   | 14.1   | 14.3 | 13.9   | 13.1  | 16.3| 13.7  |
| T-Fe₂O₅ | 1.9  | 1.9    | 2.2    | 2.0    | 2.2  | 2.5    | 0.8   | 7.3 | 11.5  |
| MnO     | 0.04 | 0.04   | 0.03   | 0.04   | 0.04 | 0.03   | 0.03  | 0.14| 0.16  |
| MgO     | 0.4  | 0.5    | 0.5    | 0.5    | 0.5  | 0.5    | 0.2   | 2.6 | 3.4   |
| CaO     | 1.6  | 1.4    | 1.4    | 1.1    | 1.1  | 0.9    | 0.1   | 4.2 | 5.7   |
| Na₂O    | 2.9  | 2.8    | 2.7    | 2.6    | 2.3  | 2.4    | 3.9   | 4.1 | 3.3   |
| K₂O     | 4.6  | 4.8    | 5.0    | 4.9    | 5.0  | 6.1    | 4.4   | 2.9 | 2.4   |
| P₂O₅    | 0.07 | 0.08   | 0.08   | 0.08   | 0.08 | 0.08   | 0.01  | 0.42| 0.38  |
| LOI     | 2.2  | 2.4    | 2.5    | 2.3    | 2.5  | 2.5    | 1.2   | 9.8 | 6.4   |
| Total   | 98.9 | 98.9   | 98.9   | 98.9   | 99.0 | 100.0  | 99.6  | 99.9| 99.4  |
| ppm     |      |        |        |        |      |        |       |     |       |
| Zr      | 132  | 198    | 154    | 202    | 152  | 149    | 61    | 152 | 178   |
| Al      | 52.7 | 55.2   | 56.7   | 59.5   | 61.5 | 66.1   | 53.4  | 39.4| 39.2  |
| CCPI    | 23.4 | 24.2   | 25.4   | 25.0   | 26.9 | 25.6   | 10.4  | 58.5| 72.3  |
| ASI     | 1.12 | 1.15   | 1.15   | 1.22   | 1.28 | 1.13   | 1.15  | 0.93| 0.75  |

AI, 100(K₂O + MgO)/(K₂O + MgO + Na₂O + CaO).
CCPI, 100(MgO + FeO)/(MgO + FeO + Na₂O + K₂O).
ASI, Al₂O₃/(CaO + Na₂O + K₂O).
Alteration index (AI), Ishikawa et al. (1976); Chlorite-carbonate-pyrite index (CCPI), Large et al. (2001); Aluminum saturation index (ASI).
the underground diorite dike are 39.4 and 58.5, respectively. The AI and CCPI values of the quartz porphyry dike are 53.4 and 10.4, respectively. The CCPI value of the quartz porphyry dike plots outside the box of felsic rocks, because the quartz porphyry dike is characterized by low T–Fe₂O₃ and MgO concentrations (Fig. 10). Based on the mineralogical characteristics, the diorite dikes, granite porphyry, and quartz porphyry dikes contain some clay minerals. The granite porphyry and quartz porphyry dikes could thus be classified as weakly altered rocks.

The aluminum saturation index (ASI) values for the granite porphyry and quartz porphyry dikes were also examined. The ASI value of the granite porphyry ranges from 1.12 to 1.28, and that of the quartz porphyry is 1.15, with both units possessing similar high ASI values. In addition, the magnetic susceptibility of the granite porphyry, quartz porphyry, and diorite were measured in the field. The granite porphyry and quartz porphyry dikes have a low magnetic susceptibility, at less than 0.1 × 10⁻³ SI unit, while the diorite has a magnetic susceptibility (15–60) × 10⁻³ SI unit. The high ASI values, low magnetic susceptibility, and presence of ilmenite in both the granite porphyry and quartz porphyry dikes suggest that the granite porphyry and quartz porphyry dikes in the Xiajinbao deposit are ilmenite-series granitic rocks. The higher magnetic susceptibility and the occurrence of magnetite in the diorite dikes suggest that the diorite dikes were formed by oxidized magma.

Zr and Ti are immobile elements during hydrothermal alteration and metamorphism. The ratios of these immobile elements can thus be utilized to discriminate the type of igneous rocks (MacLean and Barrett, 1993). The Zr/TiO₂ ratios of the igneous rocks in the Xiajinbao mining area are shown in Figure 11. The Zr/TiO₂ ratios of the granite porphyry and quartz porphyry dikes are both around 770 (Fig. 11). However, the Zr/TiO₂ ratios of the underground and outcropping diorite dikes are approximately 150 and 60, respectively, and they are different from the ratios of the granite porphyry and quartz porphyry dikes. These ratios indicate that similar magmas formed the granite porphyry and quartz porphyry dikes, whereas different magmas formed the diorite dikes. The igneous activities that shaped the Xiajinbao mining area are thus inferred to be bimodal.

**CHEMICAL COMPOSITION OF SPHALERITE**

The sphalerite concentrations of the pyrite–chalcopyrite quartz veins and sphalerite–galena quartz veins throughout the Xiajinbao deposit were analyzed by EPMA. The Fe, Cd, Cu, and Ga concentrations of sphalerite in the pyrite–chalcopyrite quartz veins are 0.3–0.6, 0.9, 0.2–
0.5, and below the detection limit of 0.1 wt%, respectively. The Fe, Cd, Cu, Mn, and Ga concentrations of the sphalerite–galena quartz veins are 0.4–8.4, 0.6–1.6, below the detection limit of 0.4, below detection the limit of 0.2, and below the detection limit of 0.1 wt%, respectively. The In concentration in all samples was below the detection limit (Table 2).

**Table 2. Chemical compositions of sphalerite in the pyrite-chalcopyrite quartz veins and sphalerite-galena quartz veins of the Xiajinbao deposit**

| No. | 1     | 2     | 3     | 4     | 5     | 6     | 7     | 8     | 10    | 11    | 12    | 13    | 14    | 15    | 16    | 17    | 18    |
|-----|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|
| Zn  | 66.9  | 67.1  | 66.4  | 64.9  | 65.2  | 64.4  | 66.3  | 66.5  | 65.5  | 63.4  | 63.3  | 65.9  | 64.6  | 66.7  | 65.7  | 58.5  | 59.2  |
| Fe  | 0.3   | 0.3   | 0.6   | 1.6   | 1.7   | 3.1   | 0.8   | 0.4   | 1.8   | 3.5   | 3.2   | 1.2   | 3.0   | 0.5   | 1.0   | 8.4   | 7.4   |
| Mn  | -     | -     | -     | -     | -     | -     | -     | -     | -     | -     | -     | -     | -     | -     | -     | -     | -     |
| Cd  | 0.9   | 0.9   | 0.9   | 0.6   | 0.6   | 1.0   | 1.6   | 1.6   | 0.7   | 0.9   | 1.0   | 0.9   | 0.7   | 1.0   | 0.9   | 0.8   | 0.9   |
| Cu  | 0.3   | 0.2   | 0.5   | 0.3   | 0.2   | -     | 0.2   | -     | 0.2   | -     | 0.1   | 0.4   | -     | 0.1   | 0.3   | -     | -     |
| In  | -     | -     | -     | -     | -     | -     | -     | -     | -     | -     | -     | -     | -     | -     | -     | -     | -     |
| Ga  | -     | 0.1   | 0.1   | -     | -     | -     | -     | -     | 0.1   | 0.1   | -     | -     | 0.1   | 0.1   | -     | 0.1   | -     |
| S   | 32.8  | 33.1  | 33.0  | 33.1  | 33.3  | 33.5  | 33.1  | 33.1  | 33.3  | 33.0  | 33.4  | 33.6  | 33.2  | 33.4  | 33.3  | 33.7  | 33.8  |
| Total | 101.1 | 101.7 | 101.6 | 100.5 | 101.1 | 101.9 | 101.9 | 101.7 | 101.4 | 101.1 | 101.3 | 101.6 | 101.6 | 101.9 | 101.3 | 101.8 | 101.7 |

- Below detection limit. Nos. 1 to 3 are sphalerite data from the pyrite-chalcopyrite quartz veins. Nos. 4 to 18 are sphalerite data from the sphalerite-galena quartz veins.

### Estimation of the formation temperature

The fluid inclusions in the quartz grains of the pyrite quartz veins (Stage I), pyrite-chalcopyrite quartz veins (Stage II), and sphalerite-galena quartz veins (Stage III) were investigated (Table 3). Microscopic observations indicated that most of the fluid inclusions had irregular shapes, such as ellipsoidal shapes. The sizes of the fluid inclusions ranged from 4 to 10 μm. All of the fluid inclusions in these samples were liquid-dominant vapor-liquid two-phase fluid inclusions that were homogenized into the liquid phase when the fluid inclusions were heated. These facts suggest that there was no boiling during the mineralization.

The homogenization temperatures of the fluid inclusions in the pyrite quartz veins range 225–300 °C with a mode of homogenization temperature at 265 °C. The salinities of the fluid inclusions in the pyrite quartz veins range 6.0–11.4 wt% NaCl equivalent (Fig. 12). The homogenization temperatures of the fluid inclusions in the pyrite-chalcopyrite veins range 215–290 °C with a mode of homogenization temperature at 250 °C. The salinities

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**Figure 11. TiO2–Zr diagram of the igneous rocks in the Xiajinbao deposit.**
of the fluid inclusions in the pyrite–chalcopyrite quartz veins range 6.2–12.6 wt% NaCl equivalent. The homogenization temperatures of the fluid inclusions in the sphalerite–galena quartz veins range 175–255 °C with a mode of homogenization temperature at 220 °C. The salinities of the fluid inclusions in the sphalerite–galena quartz veins range 9.1–13.8 wt% NaCl equivalent. The homogenization temperatures of these veins decreased from the pyrite quartz veins (Stage I) to the galena–sphalerite quartz veins (Stage III) according to the progress of mineralization, whereas the salinities of the fluid inclusions increased from Stage I to Stage III (Fig. 12).

The common mineral assemblage of the alteration minerals of the Xiajinbao deposit is quartz–illite–pyrite. Based on the presence of illite in the Xiajinbao deposit and the stability of illite above 200 °C (Henley et al., 1984), the formation temperature of the Xiajinbao deposit was estimated to be above 200 °C. This temperature is in agreement with the temperature of the pyrite quartz veins, pyrite–chalcopyrite veins, and sphalerite–galena quartz veins determined by the fluid inclusion study.

The δ34S values of pyrite and galena from the sphalerite–galena quartz veins range from +3.3 to +3.6‰ and −0.2 to +0.8‰ (Table 4), respectively. These δ34S values are similar to the δ34S values of magmatic sulfur. The formation temperature of sphalerite–galena quartz veins could also be estimated on the basis of the δ34S values of pyrite and galena that coexist in the sphalerite–galena quartz veins and the fractionation factor between pyrite and sphalerite (Rye, 1974). The formation temperatures of sphalerite–galena quartz veins were estimated to be around 235 °C (Table 4). The formation temperature that

| Sample | Type | Homogenization temperature (°C) | Final ice melting temperature (°C) | Salinity (NaCl wt% eq.) |
|--------|------|--------------------------------|----------------------------------|------------------------|
| Z31 Pyrite quartz vein | Quartz | 273 | -3.4 | 6.6 |
| | | 224 | -4.5 | 8.9 |
| | | 236 | -5.6 | 11.4 |
| | | 301 | -4.2 | 8.3 |
| Z95 Pyrite quartz vein | Quartz | 273 | -4.7 | 9.4 |
| | | 261 | -3.8 | 7.4 |
| | | 259 | -4.1 | 8.1 |
| | | 281 | -3.5 | 6.8 |
| | | 290 | -4.3 | 8.5 |
| | | 240 | -3.1 | 5.9 |
| | | 260 | -3.6 | 7.0 |
| | | 252 | -4.8 | 9.6 |
| | | 246 | -5.2 | 10.5 |
| | | 267 | -4.6 | 9.1 |
| Z47 Pyrite-chalcopyrite quartz vein | Quartz | 223 | -4.7 | 9.4 |
| | | 247 | -5.5 | 11.2 |
| | | 253 | -3.2 | 6.2 |
| | | 286 | -3.8 | 7.4 |
| | | 269 | -5.3 | 10.7 |
| | | 216 | -5.7 | 11.6 |
| | | 246 | -6.1 | 12.6 |
| | | 219 | -4.6 | 9.1 |
| | | 231 | -4.8 | 9.6 |
| | | 259 | -5.4 | 10.9 |
| Z158 Sphalerite-galena quartz vein | Quartz | 224 | -5.2 | 10.5 |
| | | 220 | -5.4 | 10.9 |
| | | 256 | -5.6 | 11.4 |
| | | 187 | -4.6 | 9.1 |
| | | 173 | -4.8 | 9.6 |
| | | 216 | -5.4 | 10.9 |
| | | 204 | -6.2 | 12.8 |
| | | 189 | -5.3 | 10.7 |
| | | 246 | -6.6 | 13.8 |

Table 3. Microthermometric data for the fluid inclusions in quartz from the pyrite quartz veins, pyrite–chalcopyrite quartz veins, and sphalerite–galena quartz veins of the Xiajinbao deposit
was estimated by the sulfur isotope ratios is in agreement with the formation temperatures estimated by fluid inclusions and the presence of illite as alteration minerals. Integration of these results suggests that the formation temperature of the gold–bearing quartz veins of the Xiajinbao deposit was around 220–265 °C.

Estimation of sulfur fugacity

The sulfur fugacity during the formation of the sphalerite–galena quartz veins was estimated on the basis of the FeS concentration in sphalerite and the thermodynamic data (Barton and Toulmin, 1966). The Fe concentration of sphalerite in the sphalerite–galena quartz veins of the Xiajinbao deposit ranges 0.3–7.1 at% and 0.6–14.2 FeS mol% (Fig. 13). Based on the formation temperatures estimated by the homogenization temperature of the fluid inclusions and the chemical composition of sphalerite, the sulfur fugacity of the Xiajinbao deposit ranges from 10−7.8–10−11 and 10−9.5–10−14.3 atm for the pyrite–chalcopyrite quartz veins and the sphalerite–galena quartz veins, respectively. (Fig. 13). The sulfur fugacity of the pyrite–chalcopyrite quartz veins is slightly higher than that of the sphalerite–galena quartz veins. The mineral assemblage changes from the pyrite–chalcopyrite dominant stage (stage II) to the sphalerite–galena dominant stage (stage III). The variation in formation environments from stages II to III is similar to the evolution of formation environments from Zn–Cu mineralization to a Pb–Zn–Ag zoned base metal vein derived from oxidized andesitic magma (Einaudi et al., 2003).

Oxygen and hydrogen isotope ratios

The δ18O values of quartz from the pyrite quartz veins, pyrite–chalcopyrite quartz veins, and sphalerite–galena quartz veins range from +13.7 to +14.3, +13.6 to +13.7, and +14.5 to +16.7‰, respectively (Table 5). The δ18O value of the hydrothermal solution forming the gold–bearing quartz veins was estimated, with the assumption of an isotopic equilibrium between quartz and H2O (Matsuhisa et al., 1979), from the formation temperature and δ18O value of quartz. Based on the homogenization temperature of the fluid inclusions, the δ18O values of the hydrothermal solution in equilibrium with quartz in the pyrite quartz veins, pyrite–chalcopyrite quartz veins, and sphalerite–galena quartz veins were estimated under temperatures of 265, 250, and 220 °C, respectively. The δ18O values of the hydrothermal solution ranged from +4.9 to +5.5, +4.2 to +4.3, and +3.3 to +5.5‰, respectively.

The δD values of the hydrothermal solution were derived from the formation temperature and the assumption of an isotopic equilibrium between illite and H2O from the fractionation factor between illite and H2O (Sheppard and Gilg, 1996). The δD values of the hydrothermal solution in equilibrium with illite near the pyrite quartz veins, pyrite–chalcopyrite quartz veins, and sphalerite–galena quartz veins ranged from −39 to −42, −41 to −47, and −38 to

Table 4. Sulfur isotope ratios of the sulfide minerals and estimated formation temperatures of the Xiajinbao deposit

| Sample | \( \delta^{34}S_{SV-CDT} \) (%o) | \( \delta^{34}S_{SV-CDT} \) (%o) | Temperature (°C) | Remarks |
|--------|-----------------|-----------------|-----------------|--------|
| S1     | 3.5             | −0.1            | 235             |        |
| S2     | 3.4             | −0.2            | 235             |        |
| S3     | 3.3             | 0.8             | 337             |        |
| S4     | 3.6             | −0.2            | 222             |        |
| PD9    | 4.6             | 1.6             |                 | This study |
| PD20   | 7.4             | 0.4             |                 |        |
| PD2-Py2| 4.3             |                 |                 |        |
| PD2-Ga-Py | 3.2           |                 |                 |        |
| PD2-Py1| 2.4             |                 |                 |        |
| LTI(882)04 | 1.9          |                 |                 |        |
| PD516-2| 3.0             |                 |                 |        |
| PD516-4| 2.2             |                 |                 |        |
| PD516-7| 2.9             |                 |                 |        |
| PD516-9| 2.0             | 0.9             |                 |        |
| PD516-11| 3.1            |                 |                 |        |
| LTI(882)05 | −2.1         |                 |                 |        |

(1) Yang (1983). (2) Hebei Bureau of Geology and Mineral Resources Exploration (1992).
−43‰, respectively. These δ\(^{18}\)O and δD values of the hydrothermal solution forming the pyrite quartz veins and pyrite–chalcopyrite quartz veins plot near the area of felsic magmatic water and primary magmatic water (Fig. 14). The δ\(^{18}\)O values of the sphalerite–galena quartz veins are similar to those of the hydrothermal solution forming the pyrite quartz veins and pyrite–chalcopyrite quartz veins, and are also close to the δ\(^{18}\)O value of magmatic water. These values suggest that the hydrothermal solution that promoted the gold mineralization of the Xijinbao deposit was derived from magmatic water, regardless of whether the magmatic water was derived from oxidized magma or not.

The salinity of the fluid inclusions in the gold-bearing sphalerite–galena quartz veins is higher than the salinities of the fluid inclusions in the pyrite quartz veins and pyrite–chalcopyrite quartz veins (Fig. 12). Based on these data, there is a possibility that the contribution of the magmatic components increased when the sphalerite–galena quartz veins were formed.

The initial water related to the formation of the pyrite quartz veins, pyrite–chalcopyrite quartz veins, and sphalerite–galena quartz veins of the Xijinbao deposit was also estimated on the basis of the final δ\(^{18}\)O values of the altered rocks after the oxygen isotopic exchange, as a result of water–rock interactions (Taylor, 1971). The δ\(^{18}\)O values of the altered rocks near the pyrite quartz veins, pyrite–chalcopyrite quartz veins, and sphalerite–galena quartz veins range from +11.8 to +13.1, +11.3 to +11.4, and +11.4 to +12.1‰, respectively (Table 5), yielding similar values. Assuming the water–rock interactions of the felsic igneous rocks (having an initial δ\(^{18}\)O value of +8‰) with magmatic water, meteoric water, and seawater, the final δ\(^{18}\)O value of rock after the water–rock interactions can be estimated by the following equation (Taylor, 1971):

\[
\delta^{18}\text{O}_f = \frac{\delta^{18}\text{O}_i + 2W/R \times (\delta^{18}\text{O}_w + 1000 \ln \alpha_{rw})}{1 + 2W/R},
\]

where W/R is the water/rock ratio, expressed as proportion of oxygen atoms, \(r\) and \(w\) refer to rock and water, respectively, \(i\) and \(f\) refer to the initial and final states, respectively. 1000 ln \(\alpha_{rw}\) is the fractionation factor between the rock and water (Matsuhisa et al., 1979). Based on a com-

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**Table 5.** Hydrogen isotope ratios of illite and oxygen isotope ratios of quartz and the whole rock of the Xijinbao deposit

| Sample (%) | Depth (m) | Type of vein | δD\(_{\text{H}_{2}O, \text{SMOW}}\) | δ\(^{18}\)O\(_{\text{quartz,SMOW}}\) | δD\(_{\text{H}_{2}O, \text{SMOW}}\) | δ\(^{18}\)O\(_{\text{H}_{2}O, \text{SMOW}}\) | δ\(^{18}\)O\(_{\text{rock,SMOW}}\) |
|------------|----------|--------------|----------------|----------------|----------------|----------------|----------------|
| Z49        | 373.5    | Py-Qz        | -64            | +14.3          | -39            | +5.5           | +13.1          |
| Z61        | 462.7    | Py-Qz        | -67            | +13.7          | -42            | +4.9           | +11.8          |
| Z46        | 360.2    | Py-Cp Qz     | -72            | +13.6          | -47            | +4.2           | +11.4          |
| Z51        | 402.0    | Py-Cp Qz     | -66            | +13.7          | -41            | +4.3           | +11.3          |
| Z99        | 108.8    | Sph-Gn Qz    | -65            | +14.7          | -40            | +3.5           | +12.1          |
| Z15        | 164.0    | Sph-Gn Qz    | -66            | +14.5          | -41            | +3.3           | +12.5          |
| Z27        | 210.8    | Sph-Gn Qz    | -68            | +16.7          | -43            | +5.5           | +11.4          |
| Z71        | 531.5    | Sph-Gn Qz    | -63            | +16.1          | -38            | +4.9           | +11.4          |
parison of the oxygen isotope ratios of whole rock and the minerals present in plutonic rocks, the oxygen isotope ratio of whole rock is similar to that of plagioclase with 30% An. Therefore, the fractionation factor between whole rock and water can be substituted by the fractionation factor between plagioclase (30% An) and water (O’Neil and Taylor, 1967). In the case of water–rock interactions between felsic igneous rocks and seawater (Fig. 15b) or meteoric water (Fig. 15c) and at temperatures ranging 200–300 °C, the δ18O values of the altered rock of those veins should be around +9‰ for seawater and −1‰ for meteoric water. These values are lower than the δ18O values of the altered rocks of the Xiajinbao deposit. However, in the case of the water–rock interaction between felsic igneous rocks and magmatic water, the δ18O values of final rock are estimated to be from +11.5 to +12.5‰. The estimated δ18O values are similar to those of the altered rocks of the Xiajinbao deposit. The calculations suggest that the alterations of the Xiajinbao deposit were formed by the interaction between magmatic water and felsic rock under a water–rock mass ratio above one and at temperatures in the 200–300 °C range (Fig. 15a).

Relation between the oxidation state of magma and gold mineralization of the Xiajinbao deposit

Previous research on the sulfur isotope ratios of composite ores and related igneous rocks by Sato et al. (1992) suggests that there is an intimate relation between the δ34S values of the composite ores of ore deposits and those of the related igneous rocks. The ranges of the δ34S values for the magnetite–series granitoids and those for the composite ores of the mineral deposits associated with the magnetite–series granitoids are from −4 to +7‰ and 0 to +10‰, respectively, whereas those for the ilmenite–series granitoids and sulfide minerals of the mineral
deposits associated with the ilmenite-series granite porphyry dikes and the magnetite-bearing diorite dikes. The distribution of bonanza-grade gold mineralization of the Xiajinbao deposit is present near the diorite dikes in the granite porphyry.

(2) The formation temperature of the gold-bearing sulfide quartz veins ranges 220–260 °C. The formation temperature and log$f_2$ of the sphalerite–galena quartz veins are similar to the formation environment of the intermediate sulfidation condition of a Pb–Zn–Ag zoned base metal vein associated with oxidized magma. The $\delta^{18}O$ and $\delta D$ values of the ore-forming fluid for gold mineralization of the sphalerite–galena quartz veins show a signature of magmatic water. The sulfur isotope ratios of the sulfide minerals of the Xiajinbao deposit also suggest that the gold mineralization in the Xiajinbao deposit is related to oxidized magma. These facts suggest that gold mineralization of the Xiajinbao deposit was formed by oxidized igneous activity.

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