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Geology and Geochemistry of the Dongping Gold Telluride Deposit, Heibei Province, North China

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Abstract

The Dongping gold telluride deposit, located on the northern margin of the China-Korea platform, is one of the largest gold deposits in China. The deposit is spatially and temporally associated with the Shuiquangou complex (comprising calc-alkaline, weakly alkaline, and alkaline rocks) and genetically related to the weakly alkaline rocks.

The Dongping deposit is a system of quartz-filled fractures and K-feldspar-altered rocks. The ore veins consist mainly of quartz, K-feldspar with minor pyrite, tellurides, gold, sphalerite, chalcopyrite, galena, specularite, calcite, sericite, and barite. Mineralization in the deposit can be divided into five stages, with gold mineralization occurring in two major stages—in stage II as native gold (Au) in specularite-quartz veins, and in stage III as native gold and gold tellurides associated with base-metal sulfides and disseminated pyrite in altered K-feldspar rocks. The proportion of gold that is native is more than 90%. Pyrite is the most common sulfide mineral and can be recognized as three generations. All pyrites of the three generations have a cubic shape, which may suggest a low fO2-forming environment.

Introduction

The Dongping gold telluride deposit, discovered in 1985, is located in Chongli County, northern Hebei Province, about 160 km northwest of Beijing. The exploration program for the No. 1 and No. 2 ore veins developed a reserve of 35 metric tons of gold and was conducted in 1986 by the No. 8 Party of the Police Gold Army; other veins now are being explored. At the time of its discovery and early exploration, Dongping proved to be a new type of deposit for China; subsequently, more than 10 gold deposits similar to the Dongping—especially the Hegou, Zhongshangou, and Huangtuliang—were discovered in the Shuiquangou complex, with gold reserves totalling about 100 metric tons. The Dongping gold deposit proper is characterized by enrichment in tellurides. With respect to its genesis, some researchers postulated a relationship to alkaline rocks and regarded Dongping as an alkaline-rock-related gold deposit (Bonham, 1984, 1988; Head et al., 1987). After summarizing a large amount of information reported from the gold telluride and tellurium deposits of the world, Zhang and Li (1994) proposed that the Dongping deposit is related to a high-tellurium geochemical field and has no relationship to any special lithologic characteristic.

In this paper, we review the geology and geochemistry of the Dongping gold telluride deposit in detail. We present a discussion both of the ore-forming physico-chemical conditions and of the possible relationship between the alkaline rocks and the deposit.

Regional Geology

The Dongping gold deposit is located in an uplifted region of the basement, south of the Shangyi-Chongli-Chicheng deep fault, which marks the boundary between the Inner Mongolian axis and the Yanshanian subsidence belt. The regional stratigraphic sequence is an Archean amphibolite facies of metamorphic rocks of the Shanggan Group, overlain by Proterozoic marine sedimentary systems of the Changcheng and Jixian formations, and then...
DONGPING GOLD TELLURIDE DEPOSIT, CHINA

FIG. 1. Geological sketch map of the Shuiquangou complex and the distribution of gold deposits.

The Shuiquangou Complex

The Shuiquangou complex outcrops across a subrectangular 400-km² area, striking approximately E-W (Fig. 1). The western and middle parts of the Shuiquangou complex formerly were covered by an Upper Jurassic sequence of volcanic-sedimentary rocks. The complex is intruded by several Yanshanian-age granite plutons (180 to 160 Ma) so it can be inferred that the emplacement of the Shuiquangou complex occurred earlier than did the Yanshanian period. Radiometric K-Ar, Ar-Ar, and Rb-Sr isochron ages of the Shuiquangou complex have a wide spread, from 150 to 309 Ma. The youngest ages probably are the result of alteration.

On the basis of intrusive sequences, mineral paragenetic associations, and the characteristics of accessory minerals, Zhang (1995a, 1995b, 1995c) divided the Shuiquangou complex into four associations—a pyroxene diorite association, an amphibole-monzonite association, a syenite association, and an alkali-feldspar/syenite association. Each association consists of several types of rocks, although there are no clear boundaries between the different associations. The main characteristics of the four associations are listed in Table 1.

There are many enclaves in the complex, especially in the contact zone between it and the surrounding country rocks. Almost all of the enclaves are xenoliths of the country rocks that now are rich in K-feldspar and poor in amphibole and biotite compared to the original rocks.

Zhang (1995a, 1995b, 1995c) used the Rittman Index (σ) and the total alkali-silica diagram of Mutschler (1991) to separate the Shuiquangou complex into three series—a calc-alkaline, a weakly alkaline, and an alkaline series. Pyroxene diorite, quartz alkali-feldspar
TABLE 1. Main Characteristics of the Four Associations of the Shuiquangou Complex

<table>
<thead>
<tr>
<th>Association</th>
<th>Types of rocks</th>
<th>Mineralogy</th>
<th>Texture and structure</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pyroxene diorite</td>
<td>Biotite-pyroxene, diorite, and monzodiorite</td>
<td>Biotite, augite, andesine, oligoclase, apatite, magnetite</td>
<td>Subhedral grained texture; massive structure</td>
</tr>
<tr>
<td>Amphibole-monzonite</td>
<td>Amphibole-monzonite, pyroxene amphibole monzonite,</td>
<td>Perthite, microcline, oligoclase, amphibole, pyroxene, apatite, titanite, magnetite</td>
<td>Subhedral grained texture, perthitic texture; massive structure</td>
</tr>
<tr>
<td>Syenite</td>
<td>Syenite, aegirine-augite syenite, alkali-feldspar syenite, quartz alkali-feldspar syenite</td>
<td>Perthite, microcline, albite, oligoclase, aegirine, biotite, quartz, garnet, titanite, zircon, magnetite</td>
<td>Subhedral texture edulcoration border, poikilitic, perthitic, resorption texture; massive structure</td>
</tr>
<tr>
<td>Alkali-feldspar syenite</td>
<td>Syenite, alkali-feldspar syenite, quartz alkali-feldspar syenite</td>
<td>Perthite, microcline, albite, oligoclase, amphibole, quartz, garnet, titanite, zircon, allanite, bastnaesite, magnetite, ilmenite</td>
<td>Subhedral grained, hetero-granular, perthitic, poikilitic texture; massive structure</td>
</tr>
</tbody>
</table>

Syenite, and alkali-feldspar granite belong to the calc-alkaline series; amphibole-monzonite, syenite, aegirine-augite syenite, and alkali-feldspar syenite belong to the weakly alkaline or to the alkaline series. Amphibole monzonite, syenite, and alkali-feldspar syenite are the host rocks for the gold deposits in this area. Based on the study of rare-earth elements (REE), trace elements, and stable isotopes, Zhang (1995a, 1995b, 1995c) suggested that the primitive magma was derived from a mixed source of crust and mantle.

Deposit Geology

The Dongping gold deposit is located on the southern side of the contact zone of the Shuiquangou complex (Fig. 2). The host rocks for gold mineralization in the deposit are mainly syenite and alkali-feldspar syenite. Many mineralized veins have been discovered in the mine, but only five attain economic concentrations. These ore veins, developed in a group of en echelon, tensile-shear faults, vary from several hundred to more than one thousand meters in length. A single orebody is tens to more than 100 m in length, 0.24 to 5.10 m in width, and several to more than 400 m deep. The main ore veins strike from 0° to 35° NE, and dip NW at 30° to 55°. Each ore vein consists of several gold orebodies. A single gold orebody generally includes the auriferous quartz vein and the K-feldspar-altered rocks around the vein, but some orebodies contain only K-feldspar-altered rocks. The boundaries between the country rocks and gold orebodies is obscure and can be determined only by chemical analysis.

The No. 1 vein, consisting of one main vein and several branch veins, is the largest in the Dongping gold deposit and has a reserve of 22.5 metric tons of gold. The main vein strikes NNE to N-S, and some of the branch veins strike NW. The shape of the No. 1 vein becomes more complex from the surface downward (Fig. 3). At shallow depths, the orebodies consist of auriferous vein quartz and the related K-feldspar-altered rocks, and at deeper levels they contain auriferous vein quartz, branch veins, and K-feldspar-altered rocks. The shape of the No. 1 vein at depth is complex and feather-like. The No. 70 vein, which differs from the No. 1 ore vein in terms of the shape of the orebodies, contains discontinuous auriferous vein quartz. The orebodies usually contain K-feldspar-altered rocks over a width of about 10 meters.

Five hydrothermal stages can be recognized from mineral associations and crosscutting
relationships: Stage I is white quartz-K-feldspar-pyrite. The white quartz is coarse grained with a greasy luster; the K-feldspar is red and coarse grained and occurs along both sides of the quartz vein; and the pyrite is disseminated as euhedral crystals. The gold grade of Stage I is quite low.

Stage-II veins are intersected by the veins of Stage I, and consist of gold, specularite, and quartz; the specularite aggregates have a sheared or radial structure. Stage II is an important stage in gold mineralization. Native gold is about 1 mm in diameter and fills in fissures in quartz and specularite.

Stage III—a gold, polymetallic sulfide, and quartz stage—is the most important stage of gold mineralization. The quartz is grayish white, fine-grained, and semitransparent; pyrite occurs in veinlets and also in isolated masses; and telluride minerals, including calaverite, tellurium, altaite, and tellurobismuthite, are developed in Stage III. The most common telluride is calaverite. Some tellurides are associated with galena and some occur as fillings in fissures of quartz.

Stage IV consists of a gray chalcedony-like quartz and pyrite. The quartz is gray, dense, and chalcedonic on both sides of the main mineralization vein of Stage III. Its width ranges from 5 cm to 20 cm and yellow bands can frequently be seen in it. Most of the quartz veins of this stage are found deep within the No. 70 ore vein. The gold grade in this stage is low.

Stage V is a carbonate and barite stage. The carbonate-barite veins are fissure fillings and have irregular shapes, intersecting all the early quartz veins. However, they do not contain gold.

**Mineralogy**

Quartz and K-feldspar are the most abundant gangue minerals in the ores, with quartz occurring in all hydrothermal stages. Native gold occurs as grains as large as 2 mm in diameter and generally is accompanied by calaverite, pyrite, specularite, chalcopyrite, and galena. There are three gold settings—fissure fillings, gold enclosed in other minerals, and gold trapped in crystals. The ratio of these three occurrences is 43:21:36, respectively. Gold grains at shallow depths often are larger than those at deeper levels. The tellurides at shallow levels also are more abundant than those at depth, and the Te/(Au+Ag) ratio in the ores...
declines from the surface downward. Electron microprobe analyses indicate that the gold percentage in native gold ranges from 93.4% to 97.3%.

Sulfides are the major metallic minerals in the ores, but their amount is generally less than 5% by volume. Pyrite is the most abundant sulfide. Chemical analysis of pyrite indicates that the Te content ranges from 21 to 1100 ppm, with an average of 408 ppm.

**Wall-Rock Alteration**

The most important feature of wall-rock alteration in the Dongping gold deposit is an intense K-feldspathization, which occurs widely in the form of a distinctive red rock. Besides feldspathization, silica, pyrite, sericite, albite, carbonate, and barite alterations occur. The K-feldspathization and pyrite and silica alterations are closely associated with gold mineralization.

K-feldspathization can be divided into three stages. In the first, K-feldspar is coarse-grained and is accompanied by gold mineralization; this coarse-grained alteration is developed extensively in the mine. The second stage is bronze-colored and occurs along both sides of the quartz veins. Second-stage alteration actually forms an ore, because it contains 3.29 g/t of gold on average. The third stage consists of a reddish-pink and medium fine-grained feldspar that occurs outside of the second-stage K-feldspathization. The third stage alteration also has produced an ore, in which the gold grade can reach up to 49 g/t. The second and third stages generally extend 2 to 3 m and sometimes up to 10 m outside the vein.

Changes in major elements in the alteration zones are presented in Table 2 and Figure 4. The following conclusions can be derived from Figure 4:

1. There is an increase of SiO$_2$ from the primary rocks through a weak alteration zone, an intense alteration zone, and then the quartz vein. This implies that SiO$_2$ was introduced into the country rocks from the hydrothermal ore-forming fluid.

2. Al and Na change in parallel, increasing in the weak alteration zone and decreasing in the intense alteration zone; this implies that the changes can be related to the formation of albite:

$$\text{CaAl}_2\text{Si}_2\text{O}_8 + 4\text{SiO}_2 + 2\text{Na}^+ = 2\text{NaAlSi}_3\text{O}_8 + \text{Ca}^{2+}. \quad (1)$$

3. Fe$^{3+}$, Fe$^{2+}$, Ca$^{2+}$, and Mg$^{2+}$ decrease in the various alteration zones, especially in the weakly altered zone, indicating that these ions were probably removed from the primary rocks and that they supplied the material base for the later formation of pyrite and carbonate minerals.

### Table 2. Changes in Chemical Compositions in the Varied Alteration Zones

<table>
<thead>
<tr>
<th>Major oxides</th>
<th>Syenite</th>
<th>Weak alteration zone</th>
<th>Intense alteration zone</th>
<th>Quartz vein</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>A</td>
<td>B</td>
<td>C</td>
<td>A</td>
</tr>
<tr>
<td>SiO$_2$</td>
<td>63.78</td>
<td>588</td>
<td>65.77</td>
<td>592</td>
</tr>
<tr>
<td></td>
<td>+4</td>
<td></td>
<td>+16</td>
<td>+29</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>+44</td>
<td></td>
</tr>
<tr>
<td>TeO$_2$</td>
<td>0.18</td>
<td>2</td>
<td>0.05</td>
<td>0.3</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>+1.7</td>
<td></td>
</tr>
<tr>
<td>TeO$_2$</td>
<td>16.88</td>
<td>183</td>
<td>18.86</td>
<td>199</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>+16</td>
<td></td>
</tr>
<tr>
<td>Fe$_2$O$_3$</td>
<td>2.44</td>
<td>17</td>
<td>0.69</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>-12</td>
<td></td>
</tr>
<tr>
<td>FeO</td>
<td>1.28</td>
<td>10</td>
<td>0.20</td>
<td>1.5</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>-3.8</td>
<td></td>
</tr>
<tr>
<td>MnO</td>
<td>0.05</td>
<td>0.6</td>
<td>0.03</td>
<td>0.2</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>-0.4</td>
<td></td>
</tr>
<tr>
<td>MgO</td>
<td>0.41</td>
<td>14</td>
<td>0.06</td>
<td>0.8</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>-13.2</td>
<td></td>
</tr>
<tr>
<td>CaO</td>
<td>1.47</td>
<td>14</td>
<td>0.50</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>-9</td>
<td></td>
</tr>
<tr>
<td>Na$_2$O</td>
<td>5.16</td>
<td>92</td>
<td>5.66</td>
<td>98.5</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>-6.5</td>
<td></td>
</tr>
<tr>
<td>K$_2$O</td>
<td>5.04</td>
<td>60</td>
<td>7.68</td>
<td>89</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>+29</td>
<td></td>
</tr>
<tr>
<td>P$_2$O$_5$</td>
<td>0.11</td>
<td>1</td>
<td>0.1</td>
<td>0.8</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>-0.2</td>
<td></td>
</tr>
<tr>
<td>H$_2$O</td>
<td>0.26</td>
<td>16</td>
<td>0.2</td>
<td>12</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>-4</td>
<td></td>
</tr>
</tbody>
</table>

Abbreviations: A = wt%; B = atomic numbers of element in the standard rock cell; C = difference from the primary rock; (+) and (-) represent "in" and "out" respectively.
Fluid-Inclusion Studies

Fluid inclusions were observed in the quartz of the quartz veins, calcite of the calcite-barite vein, sphalerite, and K-feldspar. Thirteen representative quartz samples and one calcite sample were chosen for measurement of homogenization temperatures. The shapes of inclusions varied from irregular, ellipsoidal, and spindle-like to polygonal; some inclusions even had negative crystal shapes. The inclusions ranged from 5 to 15 µm in diameter.

The inclusions can be classified into four types (or assemblages), based on the number and volume proportions of phases present:

1. Liquid inclusions, consisting almost entirely of a liquid phase, are rare in the Dongping deposit.
2. Vapor inclusions, with vapor phase ratios from 60% to 85%, occur mainly in the early metallogenic veins.
3. Two-phase inclusions (aqueous solutions and vapor) are the most common type, with phase ratios ranging from 10% to 25%.
4. Liquid CO$_2$ inclusions commonly occur in groups.

Homogenization temperatures ($T_h$) of the primary fluid inclusions in the Dongping deposit range from 120° to 350° C (there are no pressure corrections for the $T_h$ values). $T_h$ values of the quartz–K-feldspar–pyrite stage range from 315° to 350° C; the gold–polymetal sulfides–quartz stages range from 240° to 285° C; and the late carbonate–barite stage ranges from 120° to 185° C. The $T_h$ values of the other stages have not been measured, because their inclusions are too small.

Only large and clearly visible two-phase inclusions were used to determine fluid compositions, and in most cases heating runs were made immediately after freezing. The temperature of melting of the last crystals of ice have been converted to equivalent wt% NaCl using the equations provided in Potter et al. (1978). The salinities range from 1.5 to 10.0 equiv. wt% NaCl, averaging 5.33 equiv. wt% NaCl. There is
no apparent variation in salinity with time or depth.

The CO₂ concentration has been used to estimate the pressure during deposition. Pressures range from 500 to 690 bars. The depth of mineralization can be inferred to range from 1.67 to 2.30 km from the surface at a rate of pressure increase of 250 to 300 bar per kilometer.

The compositions of liquid- and vapor-phase fluid inclusions in the Dongping gold deposit are presented in Table 3. Compositions of other gold deposits in this area also are listed for comparison. K⁺ and Na⁺ are the major positive ions in all the gold deposits of the area, and Mg²⁺ and Ca²⁺ are minor. Cl⁻ is the dominant negative ion in the Dongping and the other gold telluride deposits in this area, whereas SO₄²⁻ instead of Cl⁻ is the dominant negative ion in the Jingjiazhuang sulfide-rich gold deposit, where tellurides are not present.

Based on the compositions of the fluids (Table 3), some physico-chemical parameters of the Dongping gold deposit can be calculated: pH = 6.08–7.49; Eh = 0.58–0.86; fO₂ = 10⁻３２–10⁻３０.５. The fO₂ values calculated from the

TABLE 3. Compositions of the Liquid and Vapor Phase of the Fluid Inclusions in the Investigated Area

<table>
<thead>
<tr>
<th>Mines</th>
<th>Description</th>
<th>Unit</th>
<th>Liquid phase</th>
<th>Vapor phase</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>K⁺</td>
<td>Na⁺</td>
<td>Ca²⁺</td>
</tr>
<tr>
<td>Dongping</td>
<td>Stage I, quartz</td>
<td>µg/g</td>
<td>1.04</td>
<td>1.86</td>
</tr>
<tr>
<td></td>
<td>Stage II, quartz</td>
<td>µg/g</td>
<td>0.09</td>
<td>0.38</td>
</tr>
<tr>
<td></td>
<td>Stage III, quartz</td>
<td>µg/g</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td></td>
<td>stage IV, quartz</td>
<td>µg/g</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>Zhong</td>
<td>Polymetallic sulfides—quartz</td>
<td>µg/g</td>
<td>0.75</td>
<td>0.00</td>
</tr>
<tr>
<td></td>
<td></td>
<td>mol%</td>
<td>1.20</td>
<td>1.20</td>
</tr>
<tr>
<td></td>
<td></td>
<td>µg/g</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td></td>
<td></td>
<td>µg/g</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td></td>
<td></td>
<td>µg/g</td>
<td>0.00</td>
<td>0.00</td>
</tr>
</tbody>
</table>

**Fig. 5.** S-isotope compositions of the Dongping gold deposit.
expression of Bortnikov et al. (1988) range from $10^{-8.50}$ to $10^{-5.43}$.

### Stable-Isotope Studies

The S- and Pb-isotope compositions of minerals have been studied and reported on in detail elsewhere (Jiang et al., 1992; Song, 1991; Song and Gou, 1992), and only a brief summary of the results is provided here. The $\delta^{34}S$ values of pyrites and galenas range from $-4.2$ to $-16.3$ per mil (Fig. 5). This is similar to the values of $-4.2$ to $-15.3$ per mil from the Emperor mine (Ahmad et al., 1987). The $\delta^{34}S_{\Sigma S}$ value, calculated from the model of Ohmoto (1972), is approximately zero, suggesting that S was derived from a deep source during the hydrothermal events.

The Pb-isotope compositions of galenas from the Dongping deposit, the Shuiquangou complex, and the Archean metamorphic rocks of the Shanggan Group are listed in Table 4. There is a close overlap between the Pb-isotope compositions of galena and the Shuiquangou complex, but no overlap between those of galena and the Archean metamorphic rocks of the Shanggan Group. The projected points of the galenas (Fig. 6) are located between the two lead lines of the mantle and the orogenic zone proposed by Zartman and Doe (1981). These projected points suggest two mixing sources for Pb. The H- and O-isotope compositions in both minerals and fluid inclusions are listed in Table 5. The $\delta^{18}O_{H_{2}O}$ values in quartz were calculated from the fractional equation of quartz-H$_2$O by Zhang (1989), $1000 \ln \alpha_{H_{2}O} = 3.306 \times 10^{6}/T^2 - 2.01$, and those in calcite from the fraction equation of calcite-water by Field (1985) — $\ln \alpha_{C_{H_{2}O}} = 2.78 \times 10^{6}/T^2 - 2.89$.

There are many granite-aplite dikes in the Dongping mine, believed to be the last products of the magma that shares the same source as the Shuiquangou complex. Their H- and O-isotope compositions may represent those of the primary magma water. The $\delta D$ values of inclusions in quartz range from $-74$ to $-101.3$; the $\delta^{18}O$ values for inclusions in quartz of Stage I fall within the range of 3.57 to 3.70 per mil. The $\delta^{18}O$ values of Stages III and IV range from 0.60 to 1.75 per mil and from $-0.56$ to $-0.26$ per mil, respectively; and the $\delta^{18}O$ value of the carbonate-barite stage is $-13.20$ per mil. Therefore, the $\delta^{18}O$ values for inclusions gradually decrease from the magma stage (granite-aplite) to the late hydrothermal stage (carbonate-barite). This implies that meteoric water emerged during the process of mineralization in the Dongping area. Compared with the $\delta^{18}O$ values of the primary rocks, the $\delta^{18}O$ values of the altered rocks are lower—7.0 per mil in the weakly altered rock and 3.9 per mil in the intensely altered rock ($\delta^{18}O$ values in primary rocks are...
FIG. 6. Lead-isotope compositions of galena from the Dongping and the Shuiquangou complex plotted on the lead structural model of Zartman and Doe (1981). The filled circles represent the galena from the Dongping and the open circles represent the Shuiquangou complex.

TABLE 5. H- and O-Isotope Compositions of the Dongping Deposit

<table>
<thead>
<tr>
<th>No.</th>
<th>Mineral</th>
<th>Description</th>
<th>T(°C)</th>
<th>δ¹⁸O (‰)</th>
<th>δD</th>
<th>δ¹⁸O₁₈O (‰)</th>
<th>δD₁₈O (‰)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Quartz</td>
<td>Granite-aplite</td>
<td>450</td>
<td>11.8</td>
<td>-74</td>
<td>8.19</td>
<td>-74</td>
</tr>
<tr>
<td>2</td>
<td>Quartz</td>
<td>Stage I</td>
<td>330</td>
<td>10.1</td>
<td></td>
<td>3.70</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>Quartz</td>
<td>Stage I</td>
<td>280</td>
<td>11.75</td>
<td>-91.2</td>
<td>3.57</td>
<td>-91.2</td>
</tr>
<tr>
<td>4</td>
<td>Quartz</td>
<td>Stage III</td>
<td>270</td>
<td>9.1</td>
<td></td>
<td>0.60</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>Quartz</td>
<td>Stage III</td>
<td>245</td>
<td>10.49</td>
<td>-89.3</td>
<td>0.88</td>
<td>-89.3</td>
</tr>
<tr>
<td>6</td>
<td>Quartz</td>
<td>Stage III</td>
<td>240</td>
<td>11.60</td>
<td>-81</td>
<td>1.75</td>
<td>-81</td>
</tr>
<tr>
<td>7</td>
<td>Quartz</td>
<td>Stage IV</td>
<td>230</td>
<td>11.50</td>
<td>-5.60</td>
<td>0.26</td>
<td>-101.3</td>
</tr>
<tr>
<td>8</td>
<td>Quartz</td>
<td>Stage IV</td>
<td>210</td>
<td>11.20</td>
<td>-101.3</td>
<td>0.26</td>
<td>-101.3</td>
</tr>
<tr>
<td>9</td>
<td>Calcite</td>
<td>Carbonate-barite</td>
<td>180</td>
<td>0.2</td>
<td>-97</td>
<td>-13.20</td>
<td>97</td>
</tr>
<tr>
<td>10</td>
<td></td>
<td>Weakly altered rock</td>
<td></td>
<td>7.0</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>11</td>
<td></td>
<td>Intensely altered rock</td>
<td></td>
<td>3.9</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

about 7.5 per mil)—a result of the effects of meteoric water.

The δ¹⁸O and δD values for mine ore fluids and granite-aplites are plotted in Figure 7. It is apparent that the fluid inclusions, except for the carbonate-barite state, fall within the two lines of magmatic water and meteoric water (the calculated positions for the two lines are shown in Fig. 7). This reflects the H- and O-isotope compositions of the fluid-inclusion water, except that the carbonate-barite stage cannot result from a single type of original water (magmatic or meteoric), but rather results from a mixture of the two. The H- and O-isotope compositions of the inclusions in the quartz of granite-aplite dikes falls in the magmatic-water range. The fluid inclusions of the carbonate-barite stage plot on the developed line of meteoric water. This implies that the water of the final hydrothermal stage consists of meteoric water (e.g., Craig, 1961).

Thermodynamic Analyses of Formation of Tellurides

For tellurides, ideal Te₂ gas is adopted as the most appropriate reference state, because Te₂ is the most abundant species over geologically significant ranges in temperature (Afifi et al., 1988a, 1988b). The stability of metal tellurides with respect to their corresponding native elements and sulfides is a function in part of the fugacity of tellurium in the system.

The formation of some common tellurides is bounded by the following reactions:
FIG. 7. δ¹⁸O vs δD diagram of the Dongping mine. The left and right curved lines represent the developed lines of meteoric water and magmatic water, respectively. The calculated conditions for the two developed lines from the equation of mass equilibrium at 250°C and various ratios of W/R are as follows: the initial δ¹⁸O and δD values of rocks are 7.5 per mil and -100 per mil, respectively, and the initial δ¹⁸O and δD values of magmatic water are 8.10 per mil and -75 per mil, respectively. The initial δ¹⁸O and δD values of meteoric water are -15.8 per mil and -101 per mil, respectively, and the rocks contain about 1% water. Abbreviations: MWL = meteoric water line (Craig, 1961); SMOW = standard mean ocean water; MW = magmatic water; MMW = metamorphic water.

\[
\begin{align*}
\text{Au} + \text{Te}_2\text{(gas)} & = \text{AgTe}_2, \quad (2) \\
4\text{Ag} + \text{Te}_2\text{(gas)} & = 2\text{Ag}_2\text{Te}, \quad (3) \\
2\text{PbS} + \text{Te}_2\text{(gas)} & = 2\text{PbTe} + \text{S}_2\text{(gas)}, \quad (4) \\
\text{Bi}_2\text{S}_3 + \frac{1}{2}\text{Te}_2\text{(gas)} & = \text{Bi}_2\text{Te}_3 + \frac{1}{2}\text{S}_2\text{(gas)}. \quad (5)
\end{align*}
\]

For the general reaction, we use the well-known relationship:

\[
\log K = \frac{-\Delta G^\circ}{2 \times 303 RT}. \quad (6)
\]

Thus, we can hold the relationship between \( f_{\text{Te}_2} \) and temperature at a constant \( f_{\text{S}_2} \) from free-energy data. This enables us to establish the \( f_{\text{Te}_2} \) temperature diagrams of Figure 8, which indicate that low temperature favors the formation of all tellurides at constant \( f_{\text{S}_2} \) and \( f_{\text{Te}_2} \). It can be used to explain the presence of most tellurides in epithermal deposits.

If the temperature is constant, we can establish the \( f_{\text{Te}_2} \) vs. \( f_{\text{S}_2} \) diagram in Figure 9. For \( \text{AuTe}_2 \), we convert eq. (6) at the constant temperature:

\[
\log f_{\text{Te}_2} = C.
\]

Because C is constant, \( \log f_{\text{Te}_2} \) does not change upon a change of \( f_{\text{S}_2} \). However, most of the tellurides coexisting with corresponding metal sulfides are controlled not only by temperature, but also by \( f_{\text{S}_2} \) (see eq. (4)).

For reaction (4), we can convert eq. (6) to derive the following expression:

\[
\frac{\log f_{\text{S}_2}/f_{\text{Te}_2}}{\log f_{\text{Te}_2}} = \frac{-\Delta G^\circ}{2 \times 303 RT} = -C_i
\]

\[
\log f_{\text{Te}_2} = \log f_{\text{S}_2} + C.
\]

For all tellurides with the corresponding metal sulfides, their expression will be:

\[
\log f_{\text{Te}_2} = a \log f_{\text{S}_2} + b. \quad (7)
\]

Eq. (7) indicates that the tellurides with the corresponding metal sulfides show inclined lines as various slopes on the \( f_{\text{Te}_2} - f_{\text{S}_2} \) diagram (Fig. 9). The formation of calaverite is controlled by the fugacity of Te₂ at constant temperature. The formation of the tellurides with
the corresponding sulfides is controlled by the ratio of $f_{Te}/f_{S_2}$ at constant temperature. A low $f_{S_2}$ favors the formation of tellurides at the same $f_{Te}$. It can be used to explain why most gold telluride deposits contain minor sulfides.

The compositions of fluid inclusions indicate that Cl$^-$ is the major negative ion. The gold may have been carried as a chloride complex, such as AuCl$_2$ or AuCl$_4$ (Seward, 1984; Hayashi and Ohmoto, 1991). Thus, the process of gold mobility can be expressed as:

$$K_1 = \frac{[\text{AuCl}_2]}{[^{\text{Cl}}^-]^2[^{\text{H}^+}]^{\frac{1}{2}}f_{O_2}^{\frac{1}{4}}},$$

$$K_2 = \frac{[\text{AuCl}_4]}{[^{\text{Cl}}^-]^4[^{\text{H}^+}]^{\frac{3}{2}}f_{O_2}^{\frac{9}{4}}},$$

then

$$\log [\text{AuCl}_2] = \log K_1 + 2\log [\text{Cl}^-] + \frac{1}{4}\log f_{O_2} - \text{pH},$$

$$\log [\text{AuCl}_4] = \log K_2 + 4\log [\text{Cl}^-] + \frac{9}{4}\log f_{O_2} - 3\text{pH},$$

and the total Au contents are carried as gold chloride complexes:

$$\log C_{Au} \approx \log K_1 + \log K_2 + 6\log [\text{Cl}^-] + \frac{1}{4}\log f_{O_2} - 4\text{pH}.$$
FIG. 9. The stabilities of some sulfides and tellurides as a function of the fugacities in the Dongping and two other ore deposits. Legend: 1 = Dongping; 2 = Emperor gold mine, Fiji; 3 = Boulder County, Colorado. Abbreviations: Bn = bornite; Cp = chalcopyrite; Po = pyrrhotite; Py = pyrite.

derived from the magmas, perhaps inheriting these features, has a high capacity for leaching gold from the wall rocks.

The Te contents in the pyrite of the Dongping deposit are quite high. It can be inferred that the Te may exist mainly in the sulfides. When the sulfides are dissolved in an acid solution, the Te in the rocks may be released:

\[
\text{Fe(S, Te)}_2 + 4\text{H}^+ = \text{Fe}^{2+} + \text{H}_2\text{S} + \text{H}_2\text{Te}. \tag{13}
\]

Reaction (13) indicates that an acid environment favors Te mobility. When the temperature decreases, Au-chloride complexes are replaced by Au-sulfur complexes, such as Au(HS)\textsubscript{2} and AuHS. However, the fugacity of sulfur in the fluid is very low, so that part of the gold may have been carried as an Au tellurium complex, such as Au(HTe)\textsubscript{2} because of the geochemical similarity between sulfur and tellurium.

Discussion and Conclusion

The Dongping gold telluride deposit is hosted by weakly alkaline rocks and is relatively rich in tellurides. Some researchers have classified the deposit as an Au-Te or alkaline type (Bonham, 1984, 1988; Cox and Bagby, 1986; Head et al., 1987). Other examples of this type may include Mount Kare and Porgera, Papua New Guinea (Richards et al., 1991; Richards and Kerrich, 1993; Richards and Ledlie, 1993); Emperor, Fiji (Anderson and Eaton, 1990); Tongyoung, Korea; Golden Sunlight, Zortman-Landusky, and Gies, Montana; and Cripple Creek (Thompson et al., 1985) and the telluride deposits of Boulder County, Colorado. The most distinctive features of these deposits are their association with alkaline intrusive complexes and the occurrence of Au-Ag tellurides, roscoelite, and fluorite. Some investigators have proposed a direct infusion of metals and other components into the hydrothermal system via magmatic fluids (e.g., Ahmad et al., 1987), whereas others have preferred to believe that remobilization from earlier stages of magmatic hydrothermal mineralization was responsible (Moyle et al., 1990; Richards and Kerrick, 1993; Richards and Ledlie, 1993). Compared with the gold deposits mentioned above, the Dongping gold deposit is lacking in roscoelite and fluorite. Many gold telluride deposits in China are not spatially and temporally related to alkaline rocks; examples include the Xiaoyin-
pan gold deposit hosted in the metamorphic rocks of the Shanggan Group and the gold telluride deposits in the Xionger-Xiaoqinling region (Ren et al., 1993). However, such Au-Te deposits share some common features: similar crystal forms of pyrite (mainly \{100\}), high percentage of gold in the form of native gold (more than 90%), relatively low sulfide contents in ores (generally less than 5 vol%), and high Cl\(^-\) concentration in the fluid. These features suggest similar ore-forming processes among Au-Te deposits.

The pyrites of all gold telluride deposits in this area, including those hosted in the Shuiquangou complex and Archean metamorphic rocks, have similar crystal forms—cube \{100\} with very minor representation of other crystal forms—whereas the crystal forms of pyrites of the sulfide-rich gold deposits, such as the Zhangquanzhuang and Jinjiazhuang, are \{100\}+\{210\}. The crystal form of pyrite can indicate the mineralization circumstances (Cheng et al., 1987): \{100\} is indicative of conditions of low \(S_2\) fugacity and high temperature gradients; \{210\} forms under conditions of intermediate temperatures, low temperature gradient, and high \(S_2\) fugacity; and \{111\} forms under conditions of low temperature, slight alkalinity, hypabyssal fluids, and high As contents. All of this implies that the pyrites of the Dongping and other gold telluride deposits hosted in the Shuiquangou complex were formed under conditions of low and high temperature gradients. The low in the fluid resulted in a high \(f_{Te}/f_{S_2}\) ratio, which would favor the formation of tellurides.

The ratio of native gold in total gold is high in the Dongping mine. Such a high ratio also is present in the other gold telluride deposits in China, such as Lishui (Jiangsu), Shanggong, and Beilin (Henan). This phenomenon can be explained by the following equation of Bortnikov et al. (1988):

\[
\log f_{Te} = \frac{1}{4\times576T} \left[-55592 + 22 \times 87T - 18 \times 296T \log N_{Ag} + 4(1 - N_{Ag})^2 \right] \left[5650 - 1600(1 - N_{Ag}) - 1 \times 375T \right],
\]

(14)

where \(N_{Ag}\) is the atomic percentage of silver in the Au-Ag solid solution and \(T\) is the temperature in degrees Kelvin. Eq. (14) is valid at temperatures above 478.15K. Under conditions of constant temperature, if we exert the derivative on both sides of eq. (14) we can conclude that:

\[
(\log f_{Te}) < 0.
\]

From eq. (15) we can conclude that \(\log f_{Te}\) decreases with the \(N_{Ag}\); that is, \(f_{Te}\) is positively correlated with the gold percentage of native gold. Therefore, the percentage of native gold in the gold telluride deposits with high \(f_{Te}\) is high.

Cl\(^-\) is the major negative ion in the mineralization fluid in the Dongping gold telluride deposit. It has been reported that Cl\(^-\) also is the dominant negative ion in many other gold telluride deposits, e.g., Emperor, Fiji (Ahmad et al., 1987); Porgera, Papua New Guinea (Richards and Kerrich, 1993); Lishui, Jiangshu province (Wang et al., 1985); and Shanggong, Henai province (Ren et al., 1993). The high Cl\(^-\) and low SO\(_4^{2-}\) concentration in the fluid may indicate a low \(f_{Te}\); a high concentration in the fluid would favor gold mobility. There is one factor connecting all Au-Te types of deposit in China, including the Dongping: they are located in high-Te geochemical fields. The S- and Pb-isotope compositions of Dongping indicate that S and Pb components originated from a mixed source of magma and metamorphic rocks. Therefore, the high Te content of the Shuiquangou complex and Archean metamorphic rocks is a base for the formation of Te-rich hydrothermal solutions. The mobility of greater amounts of tellurium from the source became another factor leading to the increase of \(f_{Te}\).

Thermodynamic analyses indicate that a low \(pH\) value favors Te mobility from the source; physico-chemical conditions favoring gold mobility include a low \(pH\), high temperature, high \(f_{S_2}\), and high Cl\(^-\) concentration. The early hydrothermal fluids evolved from magmas of the Shuiquangou complex and Archean metamorphic rocks is a base for the formation of Te-rich hydrothermal solutions. The mobility of greater amounts of tellurium from the source became another factor leading to the increase of \(f_{Te}\). Thermodynamic analyses also indicate the existence of beneficial physico-chemical conditions for the precipitation of tellurides: relatively low temperature, high \(f_{Te}\), and a high \(f_{Te}/f_{S_2}\) ratio. The tellurides of the Dongping deposit were formed at the medium-late stage of hydrothermal activity, when the temperature decreased to about 250°C. The high percentage of gold in native gold indicates a high \(f_{Te}\), whereas a low \(f_{S_2}\) leads to a high \(f_{Te}/f_{S_2}\) ratio. Therefore, a
favorable environment exists for the precipitation of tellurides in the Dongping deposit. For other gold telluride deposits not hosted in an alkaline complex, such as the Xiaoyinpan gold telluride deposit, common features indicate that they also have favorable conditions for Te mobility and for the precipitation of tellurides. However, many gold deposits that formed in high-Te geochemical fields, but in which tellurides are absent, may have had unfavorable conditions for Te mobility or for the precipitation of tellurides: these include the Zhangguanzhuang and the Jingjiazhuang gold deposits, which are in the same geochemical field as the Dongping deposit, but have high Te values that are too high. However, the presence of some gold deposits hosted in the alkaline complex, but lacking tellurides, indicates that the existence of an alkaline complex is not a unique condition for the formation of tellurides. The discussion above indicates that the three necessary conditions for the formation of tellurides are: (1) a high-Te geochemical field; (2) favorable physico-chemical conditions for Te mobility (low temperature); and (3) favorable physico-chemical conditions for the precipitation of telluride (low temperature, high Te/Ag ratio, and a high f_{Te}/f_{S}_2 ratio).

The alkaline magmas of the Shuiquangou complex contributed to the formation of the Dongping gold telluride deposit in several ways. They: (1) provided the heat source; (2) supplied a portion of the ore-forming fluids (water, volatile components, and Cl⁻); (3) served as the source of Au and Te (remobilization from the consolidated Shuiquangou complex leached by the convecting fluid); and (4) provided favorable conditions for Au and Te mobility.

The study of many Au-Te deposits implies that, for the most part, alkaline magmas play similar roles in Au-Te mineralization (Motschler et al., 1985; Porter and Ripley, 1983; Saunders and May, 1986; Anderson and Eaton, 1990; Moyle et al., 1990; Richards et al., 1991; Russell, 1991; Richards and Kerrich, 1993; Richards and Ledlie, 1993). These similarities can be useful in explaining the affinity of many gold telluride deposits with alkaline rocks.

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