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Geology and geochemistry of Carlin-type gold deposits in China

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Abstract The Carlin-type gold deposits in China lie mostly near the margins of the Proterozoic Yangtze and Aba cratons. Submicron-sized gold in micron-sized arsenian pyrite is disseminated in fractured Cambrian through Triassic carbonaceous shale and carbonate rocks, and is associated with anomalous Hg, Sb, As, U, and Tl. Alteration typically includes silicification, argillization, and sulfidation. Aqueous fluid inclusions contain CO₂, have relatively low temperatures of homogenization (250–150 °C), and salinities (8–2 wt% equiv. NaCl) that typically decrease from early to later stages. The indicated pressures of 105–330×10⁵ Pa correspond to depths in excess of 1.0 to 3.0 km, assuming hydrostatic conditions. The δD values of fluid inclusions (–87 to –47‰) and the calculated $\delta^{18}O$ values for water in ore fluids (–2.1 to 16.2‰) reflect interactions between meteoric water and sedimentary rocks. The $\delta^{13}C$ values of calcite (–9 to 2‰) and $\delta^{34}S$ values of sulfides (–24 to 17‰) suggest that C and S were derived from marine carbonate (and organic carbon) and diagenetic sulfides (and organic sulfur) in the vicinity of the deposits. Geologic relationships and geochronologic evidence indicate the deposits formed during a late phase of the Yanshanian orogeny (140–75 Ma).

These gold deposits share much in common with the Carlin-type gold deposits in Nevada, USA. Both occur in carbonaceous, pyritic, sedimentary rocks deposited on extended margins of Precambrian cratons. The smaller Chinese deposits are generally in more siliceous rocks and the larger Nevada deposits in more calcareous rocks.

In both countries, the host rocks prior to mineralization were affected by contractional deformation that produced many of the ore-controlling structures and the deposits do not show consistent spatial or genetic relationships with epizonal plutons. However, the Nevada deposits show broad spatial and temporal relationships with shifting patterns of calc-alkaline magmatism. The ore and alteration mineral assemblages, residence of gold, geochemical signatures, paragenetic sequence, and fluid inclusions are remarkably similar, which indicates that the deposits formed from low salinity, moderately acidic, CO₂ and H₂S-rich fluids by similar processes at relatively shallow levels in the crust (< 5 km). However, the deposits in China are associated with large Hg, Sb, As, U, and Tl deposits, which may reflect higher background abundances of these elements. Stable isotopic data suggest meteoric water evolved to become ore fluids through interactions with sedimentary rocks, although contributions of volatiles or metals from deeper levels are present in some deposits in Nevada. In both countries, the deposits appear to have formed in an area of high paleothermal gradients after a change in stress regime during the later phases of orogenic activity.

Keywords Carlin · China · Gold · Yangtze craton

Introduction

Carlin-type gold deposits are mainly known in Nevada, USA (e.g., Arehart 1996; Dobra 1997; Hofstra and Cline 2000; Hofstra et al. 1999) and southwest China (e.g., Cunningham et al. 1988; Tu 1990, 1992a, 1992b, 1994; Ashley et al. 1991; Liu et al. 1991, 1996; Mao 1991; Jia et al. 1992; Zhang and Yang 1993). They were not recognized as a distinct type of gold deposit until the discovery and development of the large-tonnage low-grade Carlin deposit in Nevada between 1962 and 1965 (Hausen and Kerr 1968). Since the discovery of the Carlin deposit, a large number of gold deposits with similar characteristics have been found in America,

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China, and other parts of the world. At present, these deposits are the most important producers of gold in America (Dobra 1997) and account for a small proportion of gold production in China (Tu 1992a, 1992b). The Banqi gold deposit in southwest Guizhou province was discovered in 1978 and is the first Carlin-type gold deposit recognized in China (Tu 1992a). Since then, more than 20 gold deposits and numerous occurrences of this type have been discovered in China (Liu et al. 1991, 1992, 1996; Tu 1992a) and gold production from Carlin-type deposits is expected to increase. This paper summarizes the geologic and geochemical characteristics of Carlin-type gold deposits in China to gain insights into their origin and for comparison with the Carlin-type gold deposits in the USA.

Geologic setting

The Carlin-type gold deposits in southwest China occur in two groups; one in Yunnan, Guizhou, and Guangxi provinces near the southwest margin of the Yangtze craton and another in northwest Sichuan province near the east and south margin of the Aba craton (Fig. 1). The Aba craton is presumed to be a northwestern prong of the Yangtze craton. In both areas, the Carlin-type gold deposits coexist with epithermal ore deposits of Hg, Sb, As, U, and Tl that together constitute two much larger metallogenic provinces (Fig. 1). Both areas also contain Mississippi Valley-type Pb–Zn deposits and bedded barite deposits (Tu 1987) that are older than the Au, Hg, Sb, As, U, and Tl deposits.

The Yangtze and Aba cratons are bounded by the South China Caledonian fold belt, the Sanjian Caledonian–Hercynian fold belt, and the Qinling Hercynian fold belt (Fig. 1). The Yangtze and Aba cratons consist of Proterozoic metamorphic rocks overlain by a thick carbonate and shale sequence of Cambrian through Triassic age. The Cambrian to Triassic rocks in the Qinling fold belt record two periods of rifting and two periods of subduction (Jian 1996). The final collision of North China with the Yangtze and Aba cratons (Fig. 1) in Late Triassic time resulted in the formation of the much larger Cathaysia platform (Jian 1996).

Prior to this, in the Devonian, the SW and NW margins of the Yangtze craton underwent extensional faulting, which resulted in formation of the Youjian rift basin and NW Sichuan depression (Fig. 1; Gao 1992; Mao 1992; Xia et al. 1992; Zheng et al. 1994; Jian 1996). In both the Youjiang rift basin and NW Sichuan depression, a thick sequence of Devonian to Triassic marine sedimentary rocks was deposited on a thin sequence of Cambrian to Silurian sedimentary rocks (Gao 1992; Mao 1992; Xia et al. 1992; Zheng et al. 1994). Basaltic volcanic rocks are present in a few places within the Devonian to Early Triassic sections in both basins. These rocks were uplifted and folded during the Early Jurassic to Latest Cretaceous (190–65 Ma) Yanshanian

Orogeny (Cunningham et al. 1988; Ashley et al. 1991; Gao 1992; Mao 1992; Yang 1992; Jian 1996), which formed in response to subduction of the Pacific Plate under the Eurasian plate.

Igneous rocks associated with the Yanshanian Orogeny are mainly found in eastern China, distant to the Carlin-type gold deposit area, and consist largely of granites and felsic volcanic rocks. Although intrusive rocks are absent at most of the gold deposits, a few contain small mafic dikes, quartz porphyries, granite porphyries, or dacite porphyries that were emplaced during the Yanshanian Orogeny (Gao 1992; Yang 1992).

Host rocks

The sedimentary host rocks of the Carlin-type gold deposits in China are highly variable in age (Table 1), ranging from Cambrian to Triassic. Most of the deposits ($n=13$), however, are hosted in Triassic sedimentary rocks, which is consistent with the fact that formations of this age are most extensive in the Youjian rift basin and NW Sichuan depression where most of the deposits are located (Fig. 2a, b). The remaining gold deposits occur in Cambrian (e.g., Laerma and Danzhai), Devonian (e.g., Gedang and Pingding, etc.), Carboniferous (e.g., Longhuo and Shijia), and Permian (e.g., Getang) sedimentary rocks. The host rocks are generally unmetamorphosed marine sedimentary rocks that consist primarily of impure carbonates and shale. In a few deposits (e.g., Shijia and Laerma) a small portion of the ore is hosted in Yanshanian igneous rocks including diabase and dacite-porphyry. The most common host rock types are pelitic siltstones, fine-grained sandstones, and silty carbonates (Table 1), which usually contain large amounts of organic matter (Liu et al. 1992; Yin 1992; Lin 1993; Tu 1994).

Structures

The Carlin-type gold deposits in China are structurally controlled by both faults and folds formed in the Yanshanian orogeny (Cunningham et al. 1988; Ashley et al. 1991; Gao 1992; Mao 1992; Yang 1992; Jian 1996). The deposits are usually located along second-order faults (e.g., Lannigou) near major faults, but a few of them are hosted in large regional fault zones (e.g., Dongbeizhai). Over large areas in the Youjian rift basin and NW Sichuan depression (Fig. 1), folds are gentle and strata are nearly horizontal, but, locally, folds are moderately tight and are cut by both normal and reverse faults that commonly parallel fold axes. The deposits are typically located along high-angle faults, intraformational faults, and faults developed along unconformable contacts (Fig. 3a–d). Generally, most of the movement on these faults pre-dates formation of the ore deposits (Gao 1992; Yang 1992; Ni et al. 1997).

Ore minerals occur in open faults, stockworks, and breccias and replace adjacent strata resulting in the formation of disseminated ore. In brittle, impermeable, and unreactive rocks, the bulk of the ore is along faults and fractures. In less competent, permeable, and reactive lithologies, ore extends outward from faults to form large tabular bodies.

Age of mineralization

Because of the lack of minerals that are suitable for isotopic dating, the age of Carlin-type gold deposits in China is still a controversial problem. In recent years, however, some progress has been made in this respect. According to the data available, a late Yanshanian age is most probable for the deposits. Diabase dikes of two ages are present in the Shijia gold deposit in northwest Guangxi province (Fig. 1). Fresh whole-rock samples of the diabase were chosen for K–Ar dating and yielded dates of 175.0 ± 0.6 and 140.0 ± 3.5 Ma, respectively (Hu et al. 1995). A total of nine gold orebodies in the deposit are known to occur in the contact zone between the younger diabase and adjacent Carboniferous strata or in fractures within the younger diabase. Therefore, the Shijia deposit must be younger than the 140 Ma age of the dike. Gold ore is present in a dacite-porphyry dike in the Laerma gold deposit, northwest Sichuan province. This porphyry yielded a K–Ar date of 102 Ma (Chen et al. 1994), which places an older limit on the age of the deposit. The authors measured the Rb and Sr isotopic composition of inclusion fluids in ore-stage quartz and calcite from the Lannigou gold deposit. The samples define a Rb–Sr isochron (Fig. 4) that gives a date of 105.6 ± 4.5 Ma, corresponding to the late Yanshanian. Zhang and Yang (1993) carried out fission-track dating on quartz from gold-bearing quartz–pyrite veins in the Baidi and Lannigou gold deposits in southwest Guizhou province (Fig. 1). The fission-track dates obtained are consistent with each other, within errors, 82.3 ± 7.5 to 90.8 ± 6.4 Ma, and correspond to the late Yanshanian. The Pb isotopic compositions of pyrite and realgar coexisting with gold from the Jinya gold deposit in northwest Guangxi province (Fig. 1) were measured (Hu et al. 1995). The Pb isotopic compositions of these minerals are characteristic of single stage leads. The minerals were therefore dated using the RFC single-stage dating method (Shen 1987). Although Pb isotopic ages are of diverse interpretation, this method gave dates of 130 and 82 Ma that correspond to the late Yanshanian.

From the above discussion, it can be seen that the dates obtained from five gold deposits, one from the NW Sichuan area and four from the Yunnan–Guizhou–Guangxi area, determined by four different methods, are generally consistent with each other despite the variable and appreciably older ages of their host rocks. Considering that the Carlin-type gold deposits in both areas

have similar characteristics, are in similar tectonic settings, and are commonly localized by faults inferred to have formed during the Yanshanian Orogeny (Cunningham et al. 1988; Ashley et al. 1991; Gao 1992; Mao 1992; Yang 1992), it is reasonable to conclude that the gold deposits formed between 140 and 75 Ma, corresponding to the late Yanshanian.

Hydrothermal alteration and mineralogy

The host rocks are silicified and argillized along faults and fractures, although the intensity is highly variable from deposit to deposit. Generally, quartz veins are present in faults and fractures, carbonate host rocks are preferentially silicified, and alteration extends away from faults and fractures along more reactive or permeable clastic or carbonate strata. Paragenetic relationships indicate that silicification took place throughout the hydrothermal event. In contrast, argillization, characterized by minerals such as sericite, illite, montmorillonite, and kaolinite, is mainly present in the later stages (Fig. 5). In addition, the host rocks were also sulfidized, as is evident from the obvious increase of sulfides in altered rocks relative to their fresh equivalents. Ore stage pyrite and arsenopyrite are disseminated in altered rocks and occur in fractures. Gold-rich arsenian pyrite rims on diagenetic pyrite are common.

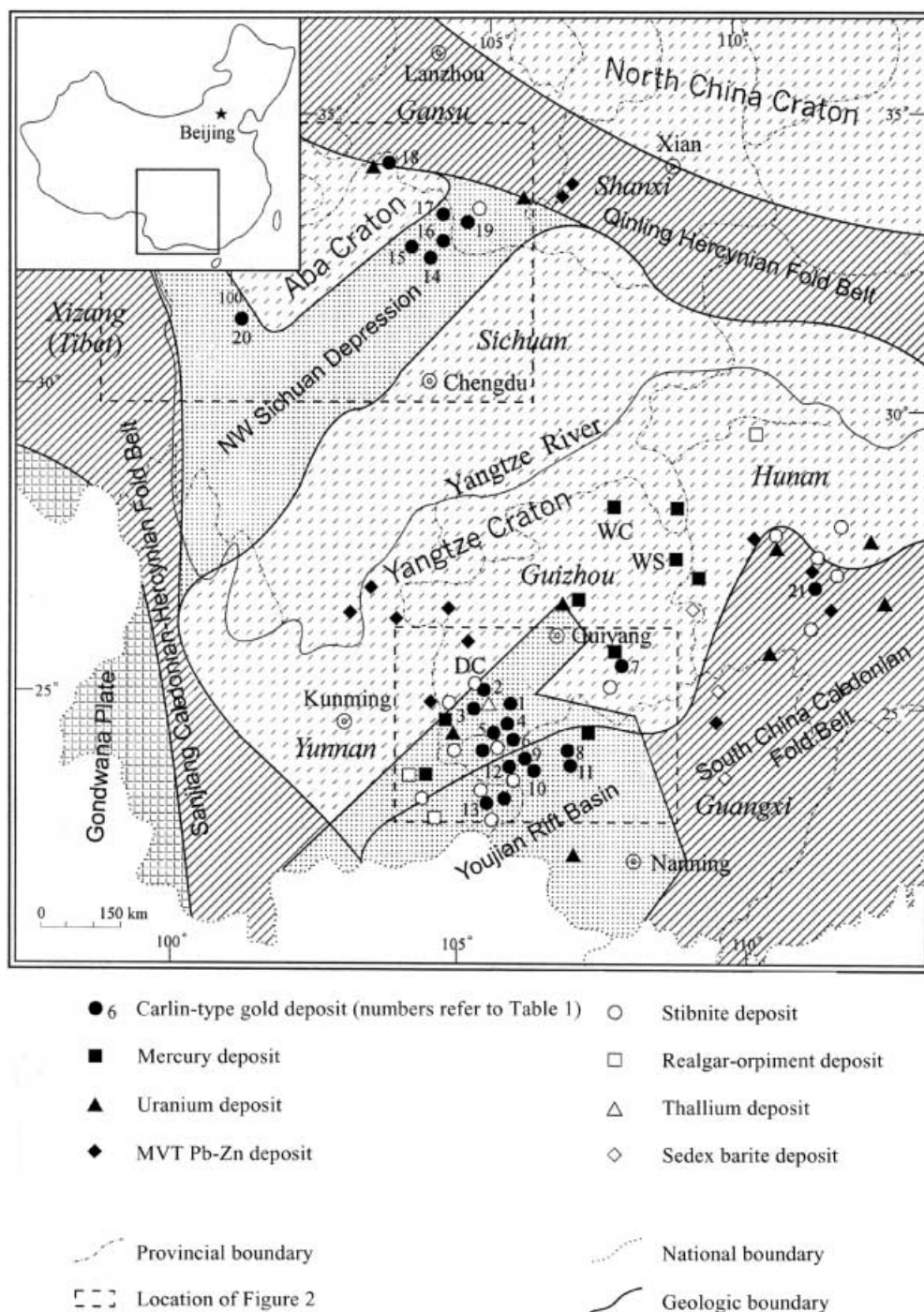
The main ore and gangue minerals in primary ores of 21 Carlin-type gold deposits in China are listed in Table 1. The textures, cross cutting relationships, and mineral assemblages in the ores indicate that mineralization took place in three stages: an early quartz–pyrite stage, a main auriferous quartz–arsenian pyrite–arsenopyrite stage, and a late quartz–calcite–realgar(orpiment)–cinnabar–stibnite stage (Fig. 5). However, each deposit exhibits minor differences. Although minor base-metal sulfides and barite are present in a few deposits (Ashley et al. 1991; Gao 1992; Liu et al. 1992; Mao 1992; Yang 1992), these minerals are paragenetically older than the main ore-stage minerals.

Liu and Wang (1994) and Mao (1991) studied the occurrence and distribution of gold at Yata, Getang, Banqi, Jinya, Laerma, and Dongbeizhai gold deposits. The results show that gold is mainly in the form of native gold, less than 1 μm in size, and that it resides in arsenian pyrite and arsenopyrite that are disseminated in quartz veins and throughout altered rocks. Stibnite, realgar, orpiment, and cinnabar are usually deposited later in the paragenesis in open cavities and fractures with only minor amounts of gold (Fig. 5).

Element associations and zoning

The mineral paragenesis is reflected by the geochemistry of the deposits. The occurrence of native gold in arsenian pyrite and arsenopyrite suggests that

Fig. 1. Location map showing major cities and province boundaries, major geotectonic units, locations of Chinese Carlin-type gold deposits, and other epithermal ore deposits in southwest China. *WC* Wuchuan Hg deposit; *WS* Wanshan Hg deposit; *DC* Dachang Sb deposit



geochemical data from the deposits will show positive correlations between gold and arsenic. This is well documented in the Danzhai deposit (Fig. 6), although in other deposits gold and arsenic do not always show such strong correlations (Fig. 6). In many deposits this is because of the presence of arsenic-rich minerals such as orpiment or realgar in the samples, which usually lack appreciable gold. In addition to As (10–10,000 ppm), the deposits are usually enriched in Sb (10–1,000 ppm), Hg (1–100 ppm), and Tl (1–100 ppm).

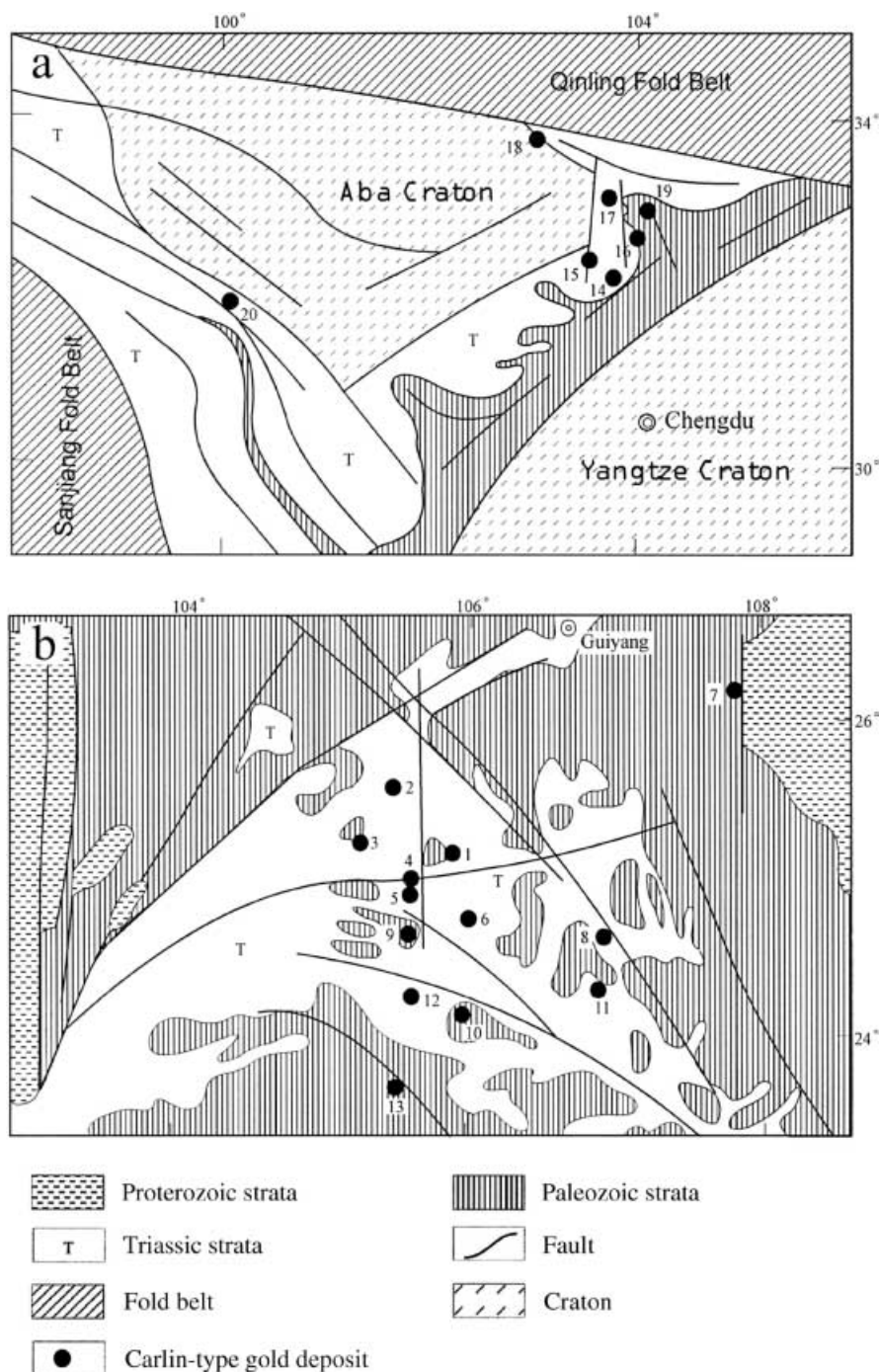
The Ag contents in the ores are usually low (generally <2 ppm) and gold–silver ratios, therefore, are relatively high ranging from 1–20. Concentrations of base metals and barium are typically low relative to average abundance values for sedimentary rocks, except in a few deposits where older base metal and barite mineralization is present.

The mineral deposits display evidence of elemental differentiation that likely reflects differences in the conditions of ore-formation. For example, most of the

Table 1. Geological and geochemical characteristics of major Carlin-type gold deposits in China. *Py* Pyrite; *Ar* arsenopyrite; *Stib* stibnite; *Re* realgar; *Orp* orpiment; *Cinn* cinnabar; *Chal* chalcocopyrite; *Sp* sphalerite; *Ga* galena; *Mc* marcasite; *Sch* scheelite; *Lo* lorandite; *Q* quartz; *Cal* calcite; *Dol* dolomite; *Ser* sericite; *Chl* chlorite; *Ill* illite; *Kao* kaolinite; *Mon* montmorillonite; *Ba* barite; *Gy* gypsum; *Fluor* fluorite; *n.a.* not available (summarized mainly after Hou 1991; Gao 1992; Zhang et al. 1995)

No.	Deposit	Age of host rock	Lithology of host rocks	Main metal minerals	Main gangue minerals	Element association	Mean grade (g/t)	Present reserve (t)
1	Lannigou	Middle Triassic	Carbonaceous siltstone, claystone	Py, Ar, Stib, Cinn, Re, Orp	Q, Cal, Dol, Ser, Ill, Kao	Au, As, Sb, Hg, Tl	7.0	80
2	Zimudang	Late Permian Early Triassic	Claystone, siltstone, silty limestone	Py, Ar, Re, Orp, Lo	Q, Cal, Ill, Kao, Ser	Au, As, Tl	6.0	60
3	Getang	Late Permian	Silicified limestone breccia, claystone	Py, Stib, Ar, Cinn, Re, Orp	Q, Cal, Fluor, Ba, Ill	Au, As, Sb, Hg, Tl	6.2	22
4	Yata	Middle Triassic	Carbonaceous siltstone, fine sandstone, claystone	Py, Stib, Ar, Mc, Re, Orp, Cinn	Q, Cal, Ser, Ba, Kao	Au, As, Sb, Hg, Tl	5.0	15
5	Banqi	Early Triassic	Claystone, pelitic siltstone	Py, Mc, Ar, Stib	Q, Cal, Ser, Kao, Ba	Au, As, Sb	5.0	10
6	Baidi	Middle Triassic	Claystone, siltstone, fine sandstone	Py, Ar, Stib	Q, Cal, Ser, Kao	Au, As, Sb	n.a.	n.a.
7	Danzhai	Cambrian	Calcareous argillite, silty limestone	Py, Ar, Cinn, Stib	Q, Cal, Ser, Kao	Au, Hg, As, Sb	5.3	7
8	Jinya	Middle Triassic	Dolomitic argillite, silty dolomite, siltstone	Py, Ar, Sp, Chal, Stib, Re, Orp	Q, Dol, Cal, Ill, Kao, Chl	Au, As, Sb, Cu, Tl	5.3	30
9	Longhe	Carboniferous	Siliceous argillite, silty limestone	Py, Ar, Cinn	Q, Cal, Ser, Kao	Au, As, Hg	5.0	10
10	Sijia	Carboniferous	Siliceous siltstone, chert, diabase	Py, Stib, Cinn	Q, Cal, Ser	Au, Hg, Sb	n.a.	n.a.
11	Mingshan	Middle Triassic	Argillite, siltstone, fine sandstone	Py, Ar, Sp, Ga, Chal	Q, Cal, Ser	Au, As, Tl	4.5	10
12	Gaolong	Middle Triassic	Calcareous Argillite, calcareous siltstone	Py, Stib, cinn	Q, Cal, Ser, Ill	Au, Sb, Hg	4.0	25
13	Gedang	Early Devonian	Carbonaceous siltstone, argillite	Py, Ar, Stib, Mc, Sp, Chal	Q, Cal, Ba, Ser, Ill	Au, As, Sb, Tl	5.3	7
14	Qiaojiao shang	Middle Triassic	Carbonaceous slate, pelitic siltstone, fine sandstone	Py, Ar, Stib, Re	Q, Cal, Ser	Au, As, Sb	4.0	15
15	Dongbeizhai	Early Triassic	Calcareous siltstone, carbonaceous slate	Py, Stib, Ar, Re	Q, Cal, Ser, Kao, Mon	Au, As, Sb, Tl	6.2	50
16	Lianhecut	Middle Devonian	Pelitic sandstone, granite porphyry	Py, Stib, Ar, Cinn, Re	Q, Cal, Ser, Kao, Ba, Gy	Au, As	4.0	10
17	Manaoke	Early Triassic Late Triassic	Siltstone, silty limestone	Py, Stib, Ar, Re, Sch	Q, Cal, Ser, Chl	Sb, Hg, Tl	5.0	20
18	Laerma	Cambrian Jurassic	Chert, carbonaceous slate, dacitic porphyry	Py, Stib, Cinn	Q, Cal, Ser, Ba	As, W	7.3	20
19	Pingding	Middle Devonian	Carbonaceous slate, chert, breccia	Py, Stib, Cinn, Re, Orp, Ar	Q, Cal, Ser	Hg, Se	11.5	5
20	Qiuluo	Late Triassic	Carbonaceous slate, pelitic siltstone, fine sandstone	Py, Stib, Ar, Chal	Q, Cal, Chl, Ser	Sb, Hg	16.0	26
21	Gaojiaao	Middle Devonian	Pelitic siltstone, silty argillite	Py, Ar, Stib, Sp, Chal, Sch	Q, Cal, Ba, Ser	Au, As Sb, W	3.5	10

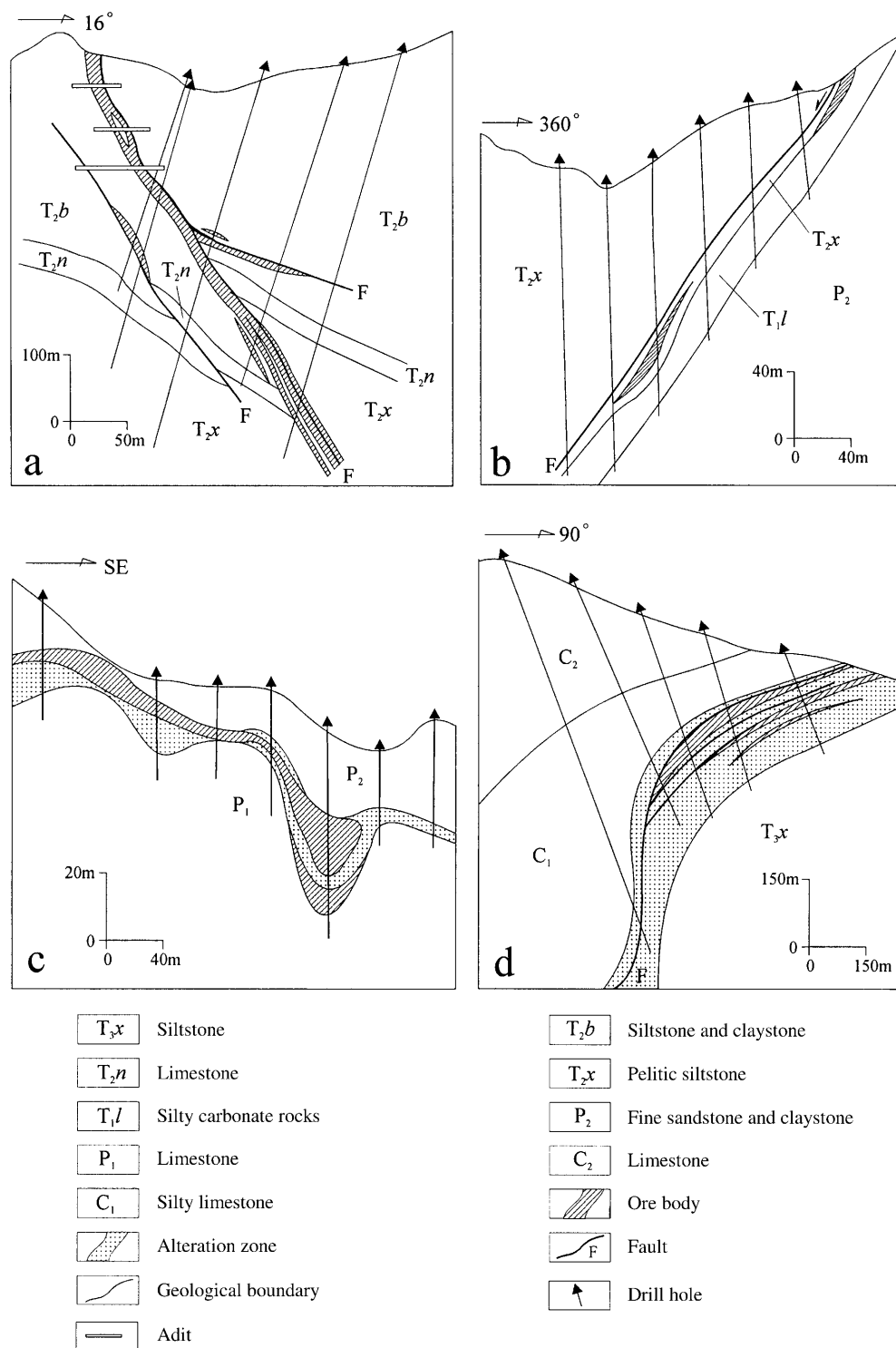
Fig. 2. Simplified geologic maps of **a** NW Sichuan area and **b** Yunnan–Guizhou–Guangxi area showing major sedimentary units, folds, faults, and Carlin-type gold deposits. 1 Lannigou; 2 Zimudang; 3 Getang; 4 Yata; 5 Banqi; 6 Baidi; 7 Danzhai; 8 Jinya; 9 Longhe; 10 Sijia; 11 Mingshan; 12 Gaolong; 13 Gedang; 14 Qiaoqiaoshang; 15 Dongbeizhai; 16 Lianhecun; 17 Manaoke; 18 Laerma; 19 Pingding; 20 Qiuluo; 21 Gaojiaao (after Yang and Hu 1994; Zheng et al. 1994)



Carlin-type gold deposits are spatially associated with ore deposits of Hg, Sb, As, or Tl that lack appreciable Au (Fig. 1). In fact, most of the gold deposits shown in Fig. 1 were found by exploration in, or near, known Hg, Sb, As, and Tl deposits and occurrences (Tu 1992a; Liu et al. 1992; Zhang et al. 1995). Furthermore, Au, As, Sb, Hg, and Tl rarely form composite ore deposits of commercial importance. Instead, one of the elements is usually dominant and the others subordinate. For example, the giant Wuchuan and Wanshan mercury de-

posits [$> 10,000$ tonnes (t) of Hg] in Guizhou province (Fig. 1) have low gold concentrations and moderate Sb and As concentrations. Similarly, the giant Dachang antimony deposit ($> 500,000$ t Sb) in Guizhou province has low concentrations of Au, As, Sb, and Hg. In the well-documented gold deposits such as Banqi, Getang, Zimudang, and Lannigou in southwest Guizhou province, the amounts of As, Sb, and Hg in the ores are of no commercial importance and could never be mined independently.

Fig. 3a–d. Cross sections illustrating the variety of stratigraphic and structural ore controls observed in the deposits. **a** Lannigou deposit; **b** Banqi deposit; **c** Getang deposit; **d** Dongbeizhai deposit. *T* Triassic; *P* Permian; *C* Carboniferous



Within individual deposits, Au, As, Sb, Tl, and Hg are typically zoned in the order of Au→Sb→Tl→As→Hg from the lower part of the orebodies upwards and from the center outwards (Zhu 1998). For example, in the Danzhai deposit, Guizhou province, the gold orebodies are located in the middle–lower parts of the mercury orebodies, showing separation of gold from mercury.

Different host rocks are also preferentially mineralized by different elements. The gold deposits are mainly hosted in siltstones and mudstones whereas the Hg, Sb, and As deposits (and occurrences) occur chiefly in carbonate rocks (Zhu 1998). For example, mercury mineralization associated with the Sanchahe gold deposit in SW Guizhou occurs mainly in a carbonate horizon of

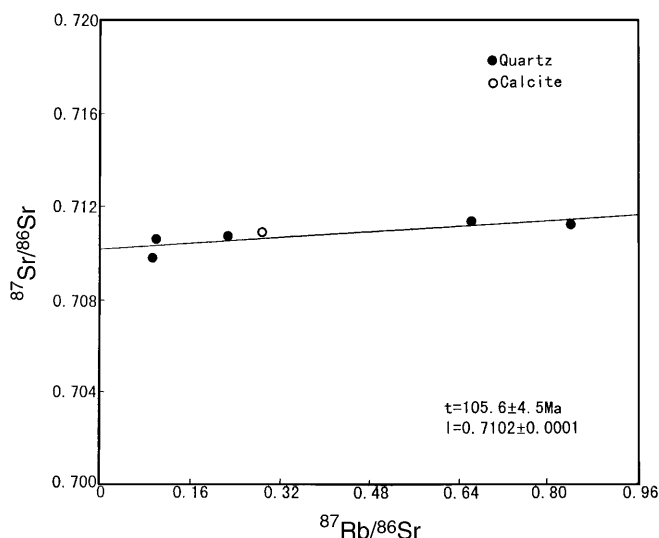


Fig. 4. Rb-Sr isochron based on analyses of inclusion fluid in quartz and calcite from the Lannigou deposit

Minerals	Hydrothermal stages			Weathering
	Early	Main	Late	
Quartz				
Pyrite				
Arsenian pyrite				
Arsenopyrite				
Gold				
Sericite				
Cinnabar				
Realgar				
Orpiment				
Stibnite				
Calcite				
Illite				
Kaolinite				
Iron oxides				
As,Sb oxides				

Fig. 5. Generalized paragenetic sequence for Carlin-type gold deposits in southwest China based on descriptions in Yang (1992), Gao (1992), and Hou (1991)

the Changxing Formation, but gold mineralization is restricted to the lower mudstone units of this formation.

Fluid inclusions

Several studies on the petrography, microthermometry, and chemical composition of fluid inclusions have recently been conducted on samples from the Lannigou, Getang, Baidi, Yata, Banqi, Gedang, Mingshan, Shijia, Jinya, and Dongbeizhai deposits (Hou 1991; Gao 1992; Yang and Hu 1994). The fluid inclusion data from these deposits have many features in common and no major differences, indicating that all of them formed from fluids of similar origins. Fluid inclusions in quartz deposited during the three hydrothermal stages typical of these deposits have the following characteristics.

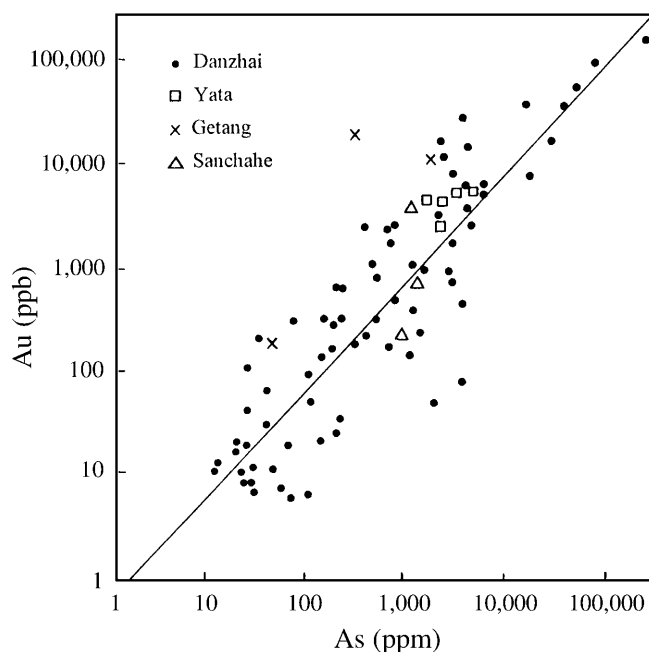


Fig. 6. Au and As composition of mineralized rocks from the Danzhai, Yata, Getang, and Sanchahe deposits based on data in Tu (1994) and Ashley et al. (1991)

1. The fluid inclusions are usually small in diameter, generally ranging from 3 to 10 μm . These fluid inclusions can be generally classified into three types: two-phase liquid-rich inclusions, two-phase vapor-rich inclusions, and three-phase (liquid H_2O , liquid CO_2 , and vapor CO_2) $\text{H}_2\text{O}-\text{CO}_2$ inclusions, although the third type is rare.
2. Ion chromatographic and gas chromatographic analyses of inclusion fluids show that they are rich in Na^+ and Cl^- relative to K^+ , Ca^{2+} , Mg^{2+} , SO_4^{2-} , and F^- , and that CO_2 is the dominant gas component (Fig. 7), suggesting the fluid could be described by the $\text{H}_2\text{O}-\text{NaCl}-\text{CO}_2$ system.
3. Homogenization temperatures of fluid inclusions are highly variable, but tend to decrease through the paragenetic sequence (Table 2) from about $250 \pm 50^\circ\text{C}$ in the early stage to $200 \pm 50^\circ\text{C}$ in the main stage, to $150 \pm 50^\circ\text{C}$ in the late stage. The data indicate that gold and ore-related minerals were deposited from relatively low temperature fluids as they cooled.
4. Salinities are moderate and tend to decrease from the early to the late stage (Table 2) from 3.0–8.5 wt% equivalent NaCl in the early stage to 3.1–6.7 wt% equiv. NaCl in the main stage to 1.9–5.0 wt% equiv. NaCl in the late stage. The decreases in salinity may indicate that the ore fluids were progressively diluted with less saline fluids. Together, the homogenization temperature and salinity data suggest that the ore fluids progressively mixed with cooler, more dilute fluids until mineralization ceased.

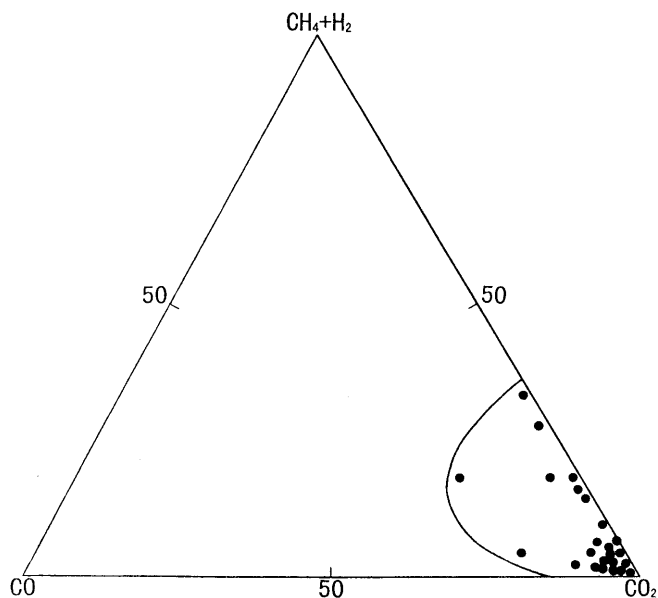


Fig. 7. Ternary diagram showing gas compositions of fluid inclusions from the Lannigou, Banqi, Yata, Getang, Jinya, Gaolong, Shijia, Mingshan, and Gedang deposits (after Yang and Hu 1994)

5. As three-phase $\text{CO}_2\text{--H}_2\text{O}$ fluid inclusions are rare, it is difficult to estimate the pressure of ore formation. Based on the few data available, minimum pressures of 105 to 330×10^5 Pa (105 to 330 bar) are indicated, which correspond to minimum depths of about $420\text{--}1,320$ m under lithostatic conditions (assuming 250×10^5 Pa/km), or $1,050\text{--}3,300$ m under hydrostatic conditions (assuming 100×10^5 Pa/km). Hydrostatic conditions, or conditions intermediate between hydrostatic and lithostatic, were likely as the deposits are mainly hosted in open faults and fractures.

Stable isotopes

Hydrogen and oxygen isotopes

The authors measured and collected hydrogen and oxygen isotope data on minerals separated from several tens of samples taken from the hydrothermal veins of seven Carlin-type gold deposits (Table 3). The $\delta^{18}\text{O}$ values of hydrothermal quartz ($9.6\text{--}27.5\text{‰}$) and calcite ($12\text{--}27\text{‰}$) are very similar to one another. The calculated $\delta^{18}\text{O}$ values for water in ore fluids that deposited these minerals (Table 3) range from -2.1 to 16.2‰ and the δD values of water in fluid inclusions range from -87 to -47‰ . The $\delta^{18}\text{O}$ and δD values of present day tunnel water, river water, and spring water from the mining districts range from -14 to -8‰ , and -95 to -56‰ , respectively.

Figure 8 shows the hydrogen and oxygen isotopic composition of water in ore fluids and the present day meteoric water relative to traditional references. The isotopic composition of present day meteoric water is probably similar to, or slightly lower than, late Yanshanian meteoric water because the ocean–continent framework has not significantly changed since then and the climate was warmer in the late Yanshanian (Frakes et al. 1992). The δD of water in ore fluids is approximately equivalent to that of meteoric water, but the $\delta^{18}\text{O}$ values are much higher. The positive “shift” of $\delta^{18}\text{O}$ values towards those of the sedimentary host rocks ($\delta^{18}\text{O}$ values of marine carbonate rocks range from 19 to 30‰ . Shen 1987) suggests the ore fluids consisted of meteoric water that evolved via isotopic exchange with shaley carbonate rocks at elevated temperatures (Shen 1987). Although the isotopic data from some deposits plot in the fields for magmatic and metamorphic water (Fig. 8), the lack of coeval igneous and metamorphic rocks in the vicinity of the deposits make it unlikely that magmatic or metamorphic fluids were important.

Carbon and oxygen isotopes

Figure 9 shows the carbon and oxygen isotopic composition of the limestone host rocks and hydrothermal calcite in the ores from several Carlin-type gold deposits relative to some traditional references (Liu et al. 1996). Unaltered limestones have isotopic compositions that are characteristic of typical marine limestones, although a few samples have lower $\delta^{18}\text{O}$ values that likely reflect some interaction with hydrothermal fluids. In comparison with unaltered limestone, the calcite in the ores has generally lower $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values that define two arrays: one with a nearly horizontal slope and one with a positive slope. The horizontal data array suggests that the CO_2 in ore fluids was derived from dissolution of marine limestone. The inclined array may be indicative of reactions between the host rocks and fluids containing CO_2 derived from oxidation of organic carbon, the atmosphere ($\delta^{13}\text{C}_{\text{CO}_2} = -7\text{‰}$), or both. Although the lower $\delta^{13}\text{C}$ values also permit a magmatic origin, the lack of coeval intrusions near most deposits discounts this possibility.

Sulfur isotopes

Figure 10 shows the range of $\delta^{34}\text{S}$ values obtained for diagenetic pyrite and hydrothermal sulfides from eight Carlin-type gold deposits. In each deposit, the calculated range of $\delta^{34}\text{S}_{\text{H}_2\text{S}}$ values is very similar to that of diagenetic pyrite in the host rocks suggesting that the H_2S in ore fluids was derived from dissolution of diagenetic pyrite. Organic sulfur may also have been a source of H_2S because it can have the same isotopic composition as diagenetic pyrite, but only in rocks that

Table 2. Homogenization temperatures and salinities of fluid inclusions from Carlin-type gold deposits in China. Based on data in Hou 1991; Gao 1992; Yang and Hu 1994, and this work

Deposit	Homogenization temperature			Salinity		
	(°C)			(equiv. wt% NaCl)		
	Early	Main	Late	Early	Main	Late
Lannigou	241–265	184–228	154–166	3.0–6.7	3.2–5.7	3.1–4.6
Getang	240–307	136–287				
Baidi	215–266	180–230	159–185	4.7–8.5	3.9–6.4	3.0–4.7
Yata		200–250	140–190			
Banqi	260–310	180–230	180–190		3.2	
Gedang	252–305	155–270	98–171	4.2–7.7	3.0–3.8	
Mingshan	250	166–213	124–147		4.5–5.4	3.2–4.8
Shijia	207–244	186–205	142–161	3.5–5.1	6.7	1.9
Jinya	280–305	220–270	143–220		4.6–5.3	2.2–3.5
Dongbeizhai		170–220	120–170			5.0
Gaolong	239–290	245–273	125–191	3.4–6.7	3.1	2.4

were deposited in euxinic environments. The $\delta^{34}\text{S}$ variations from deposit to deposit, therefore, are inherited from the host rocks and reflect variations in the $\delta^{34}\text{S}$ composition of marine sulfate when the rocks were deposited, the mechanisms of sulfate reduction, and whether the systems were open or closed (Ohmoto and Rye 1979).

For example, in the vicinity of the Lannigou deposit, only Permian and Triassic age marine sedimentary rocks are present and they are rich in organic matter. The $\delta^{34}\text{S}$ values of diagenetic pyrite (11.7 and 13.2‰) from unaltered host rocks are within the range of Permian and Triassic marine sulfate of about 10–15‰ (Shen 1987). Complete reduction of sulfate to sulfide in a closed system is required to produce pyrite with the same isotopic composition as marine sulfate. This process can occur in restricted basins or during the burial history of the sediments. The $\delta^{34}\text{S}$ values of hydrothermal sulfides in the deposit vary as follows: early hydrothermal pyrite (12.6–9.1‰), main-stage pyrite (13.6–13.0‰), late-stage realgar (11.9–11.4‰), late-stage cinnabar (11.9–10.2‰), and late-stage stibnite (9.8–8.4‰). The steady decrease in $\delta^{34}\text{S}$ values from main-stage pyrite through late-stage sulfides is consistent with the relative fractionation factors for these minerals (Ohmoto and Rye 1979). The calculated $\delta^{34}\text{S}$ values of H_2S in equilibrium with main- and late-stage sulfides at 200 °C ranges from 15.2 to 9.6‰, suggesting that H_2S was derived from dissolution of diagenetic pyrite (or possibly organic sulfur, as described above) in the host rocks.

In contrast, the Getang deposit is hosted in the Maokou and Longtan Formations of Permian age, which contain coal beds. The $\delta^{34}\text{S}$ values of diagenetic pyrite from unaltered host rocks (–12.4 to –21.5‰) are much lower than those of Permian marine sulfate (~11–14‰; Shen 1987). Biogenic reduction of marine sulfate in an open system is required to produce diagenetic pyrite with such low $\delta^{34}\text{S}$ values (Ohmoto and Rye 1979). However, as at Lannigou, the $\delta^{34}\text{S}$ values of hydrothermal pyrite (–10.0 to –24.2 ‰) and

the calculated $\delta^{34}\text{S}$ values of H_2S in ore fluids (–11.8 to –26‰) are similar to those of diagenetic pyrite, suggesting that H_2S was derived from dissolution of diagenetic pyrite (and possibly organic sulfur) in the host rocks.

Discussion

Processes of ore deposition

The absence of late Yanshanian igneous or metamorphic rocks in most mining districts make it unlikely that ore fluids were derived from magmas or metamorphic devolatilization of adjacent sedimentary rocks. The stable isotopic results suggest that the water in ore fluids consisted of variably exchanged meteoric water, that CO_2 was derived mainly from dissolution of carbonate rocks, and that H_2S was largely derived from dissolution of diagenetic pyrite in host strata. Given that gold was transported as a bisulfide complex, the source of H_2S is very important because it can place constraints on the source of gold. In this case, gold must have come from somewhere between the sites of H_2S generation and the sites of mineral precipitation. Given that H_2S appears to have been derived from the local host rocks, then gold too may have been leached from local host rocks. If so, the host rocks deserve further study to determine whether or not they were enriched in gold prior to ore formation such as in the Late Devonian Popovich Formation in the Carlin Trend (Emsbo et al. 1999). Alternatively, it is possible that the isotopic similarity between the diagenetic pyrite in the host rocks and H_2S in ore fluids is a coincidence and that H_2S was derived from isotopically similar sources at deeper levels.

Fluid inclusion studies show that during mineralization, temperatures decreased from 250 ± 50 to 150 ± 50 °C, salinities decreased from 6 ± 2 to 3 ± 2 wt% equivalent NaCl, and that CO_2 was the major gas present in ore fluids. The ore fluids were sufficiently acidic

Table 3. Hydrogen and oxygen isotope compositions of Carlin-type gold deposits in China. *N* Number of samples analyzed. $\delta^{18}\text{O}_{\text{H}_2\text{O}}$ is calculated from the $\delta^{18}\text{O}$ and homogenization temperature (average value of main stage) of quartz using the quartz–water fractionation factor; δD is measured from fluid inclusions in

quartz; H and O isotope composition of meteoric water is obtained by analyzing present tunnel water, river water, and spring water from mining districts. Data are mainly from Hou 1991; Gao 1992; Yang 1992; and this work

Deposit	N	Mineral	Temperature (°C)	$\delta^{18}\text{O}_{\text{mineral}}$ (SMOW)	$\delta^{18}\text{O}_{\text{H}_2\text{O}}$ (SMOW)	δD (SMOW)
Banqi	4	Quartz	200	9.6–16.1	–2.1–4.4	–59.1–78.8
	6	Calcite		15.6–23.1		
Jinya	7	Quartz	240	17.8–23.1	8.8–13.7	–56.2–70.0
	4	Calcite		12.0–24.8		
Gaolong	6	Quartz	250	11.3–18.6	2.3–9.6	–53.4–77.1
Gedang	5	Quartz	200	21.2–23.6	9.5–11.9	–62.1–83.4
Dongbeizhai	3	Quartz	200	23.9–27.5	12.5–16.2	–66.5–75.5
	26	Calcite		15.1–27.0		
Lannigou	8	Quartz	200	22.7–26.1	11.0–14.4	–74.0–79.0
	6	Calcite		15.5–22.6		
Maxiong	6	Quartz	250	11.5–17.0	2.5–8.0	–46.9–87.0
	12	Meteoric water			–8.0–14.0	–56.0–95.0

(pH~4–6) to dissolve calcite and alter detrital silicates in the host rocks to illite or kaolinite. At these conditions, the lack of appreciable Ag or base metals in the ores and the presence of gold, arsenian pyrite, arsenopyrite, realgar, stibnite, and cinnabar indicate that the ore fluids had relatively high H_2S concentrations ($\sim 10^{-2}$ to 10^{-3} m) and that Au was transported as a bisulfide complex (Hofstra et al. 1991).

The fluid inclusions provide clear evidence of cooling and dilution. The silicification and alteration zoning, from kaolinite and illite, without calcite outward into unaltered limestone, suggests that hot, moderately acidic ore fluids were progressively neutralized by reaction with cool carbonate host rocks and diluted with local ground water. The presence of disseminated pyrite, arsenopyrite, and gold in the host rocks suggests that these minerals precipitated by reaction of H_2S in ore fluids with iron in the host rocks (sulfidation). In this case, the iron sulfides and gold are disseminated because the sites of reactive iron in the host rocks are disseminated. Cooling and increases in pH because of reactions with the host rocks and dilution with local ground water may also have contributed to sulfide precipitation, especially where pyrite and arsenopyrite occur in veinlets. Quartz, stibnite, and realgar have prograde solubility and likely precipitated in open fractures as the fluids cooled during the waning stages of the hydrothermal systems. Cinnabar can also precipitate by cooling, but is especially sensitive to pH increases, thereby explaining its occurrence in limestones.

The apparent lack of magmatic activity in SW China is a problem for invoking magmatic heat as a main cause of the high geothermal gradient needed to drive ore fluids. Another possibility is that the high geothermal gradient could be the result of crustal thickening related to thick sediment deposition, or, more likely, contractional deformation associated with the Yanshanian orogeny. In this setting, the drive for fluid movement may have come from the high regional

heat flow and the increased permeability resulting from a relaxation of compressional stresses, which allowed influx and deep circulation of meteoric water. Meteoric waters apparently evolved to become ore fluids by circulation through carbonaceous, pyritic, calcareous sedimentary rocks. The fluids obtained their high CO_2 and H_2S contents by equilibration with sedimentary rocks at elevated temperatures and depths, scavenging Au and associated elements from rocks along fluid pathways. The resulting ore fluids were channeled upward through permeable faults, depositing Au and other elements in cool reactive permeable rocks at shallower levels. This model explains the spatial association of the deposits with high-angle faults, folds, and the lack of a specialized heat source. Further work is clearly needed to evaluate these possibilities and to determine how the deposits fit into the tectonic history of southwest China.

Comparison with Carlin-type gold deposits in Nevada

The Carlin-type gold deposits in China and those in Nevada have many features in common (Table 4). Both occur in carbonaceous, pyritic, sedimentary rocks deposited on extended continental margins of Precambrian cratons. In both countries, the host rocks were affected by contractional deformation prior to mineralization that produced many of the ore controlling structures. However, the rocks in Nevada appear to have been affected by several more episodes of deformation and magmatism prior to mineralization than those in southwest China. The gold districts in both countries contain a variety of ore deposits that are older than the gold deposits: sedex-Ba, Au, base metals, and pluton-related base metals and Au in Nevada; MVT base metals and sedex-Ba in southwest China. Deposits in both countries do not show consistent

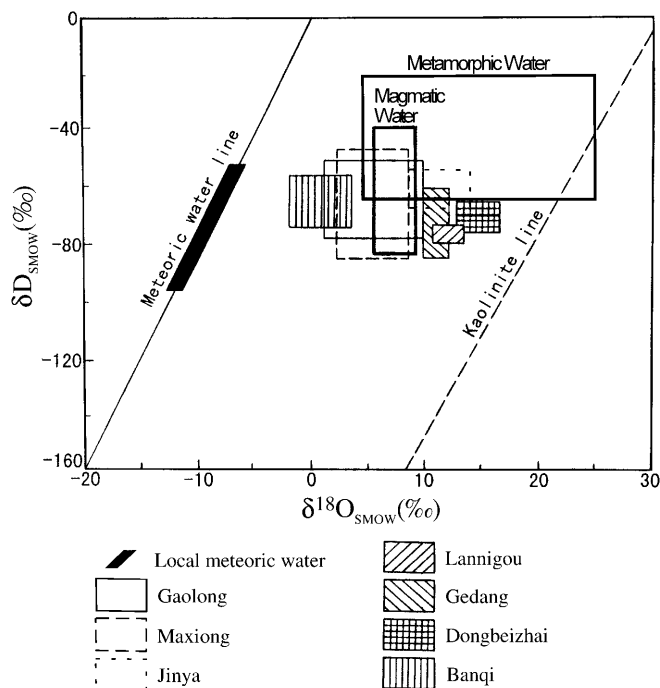


Fig. 8. Hydrogen and oxygen isotopic composition of local meteoric water and water in ore fluids based on analyses of fluid inclusions and quartz from seven gold deposits (see Table 3)

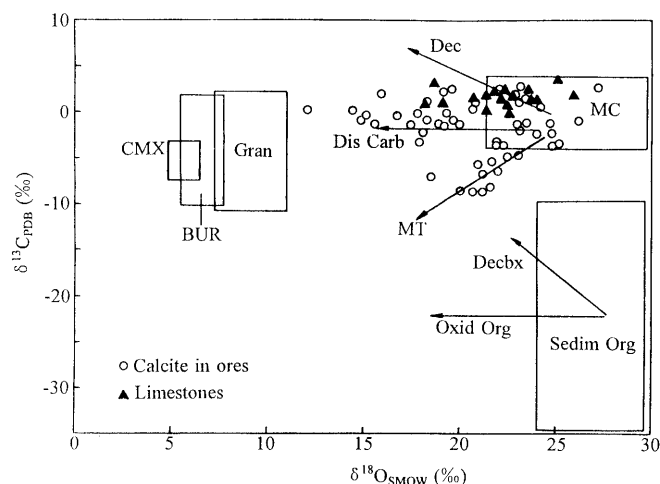


Fig. 9. Carbon and oxygen isotopic composition of calcite in ores and unaltered limestones from deposits in the Yunnan–Guizhou–Guangxi area. For reference, the fields for typical marine carbonates (MC), sedimentary organic matter (Sedim Org), and igneous carbonate (CMX) carbonatite and mantle xenoliths; BUR basic and ultrabasic rocks; Gran granite) are outlined. The arrows show typical isotopic trends resulting from carbonate dissolution (Dis Carb), decarbonation (Dec), decarboxylation of organic matter (Decbx), oxidation of organic matter (Oxid Org), and mixing trend (MT). See text for further description

spatial or genetic relationships with epizonal plutons. However, the deposits in Nevada are approximately contemporaneous with widespread calc-alkalic magmatism (which likely provided the drive for fluid

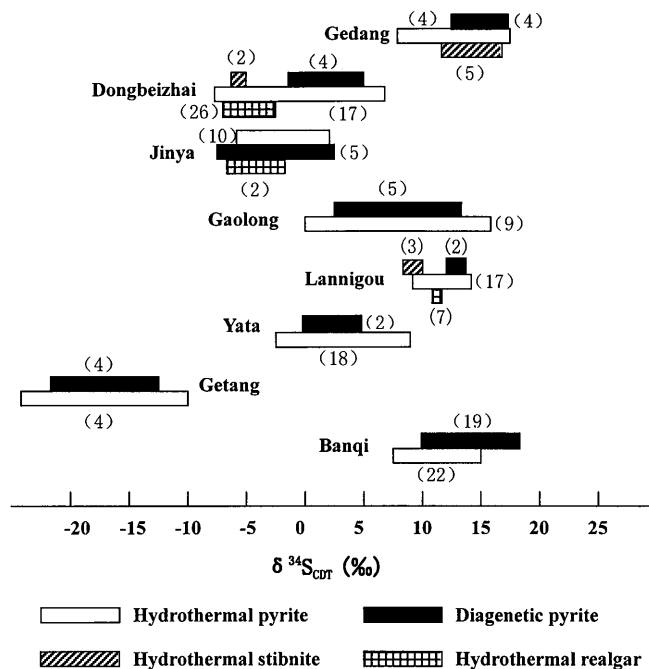


Fig. 10. Sulfur isotopic composition of diagenetic pyrite and hydrothermal sulfides from eight gold deposits. The samples with diagenetic pyrite were collected well away from known mineralization. The number in parentheses near range bars is the number of samples analyzed

movement), but in southwest China there is very little magmatism. In addition, subsequent extensional deformation and volcanism have obscured original geologic relationships in Nevada. Although deposits in both countries are commonly localized by fault intersections with favorable strata, especially in anticlines, the Nevada deposits more commonly extend well away from feeders in favorable strata. The Chinese deposits are smaller and hosted in more siliceous rocks, whereas the Nevada deposits are larger and are hosted in more calcareous rocks. The ore and alteration mineral assemblages, residence of gold, geochemical signatures, paragenetic sequence, and fluid inclusions are remarkably similar suggesting they formed from low salinity, moderately acidic, CO_2 - and H_2S -rich fluids by similar processes. However, the deposits in China are spatially associated with larger Hg, Sb, As, U, and Tl deposits, which may reflect higher background abundances of these elements in the sedimentary rocks. The mineral assemblages and fluid inclusion temperatures, salinities, and gas compositions suggest the deposits in both countries formed at relatively shallow levels in the crust. Stable isotopic data suggest meteoric water evolved to become ore fluids through interactions with sedimentary rocks, although contributions of volatiles or metals from deeper levels may be present. For example, the gold deposits in the Getchell district have characteristics very similar to other Carlin-type deposits in Nevada, but isotopic data indicate they formed from metamorphic or magmatic fluids (Cline and

Table 4. Comparison of Carlin-type gold deposits in southwest China and Nevada based on this paper and compilation by Hofstra (1999)

Characteristic	Chinese deposits	Nevada deposits
Deposit size	< 1–80 t Au	< 1–1,000 t Au
Mean deposit grades	4.0–7.0 g/t Au	0.6–22 g/t Au
Other deposit types present in some districts	Older MVT base metals; older sedex barite; coeval Sb, As, Tl, Hg, U vein and replacement deposits	Older sedex Au, barite, and base metals; older pluton-related base and precious metals; younger Hot spring Ag–Au and Hg
Age of deposits	Cretaceous	Mid-Tertiary
Setting during mineralization	Yanshanian orogeny	Cordilleran orogen soon after change from contraction w/o magmatism to extension w/magmatism
Relation to coeval plutons	No consistent spatial or genetic relationship	No consistent spatial or genetic relationship
Drive for fluid movement	Increased regional heat flow and increased permeability	Increased regional heat flow and increased permeability
District controls	Deposits are commonly in antiforms along near vertical faults; deposits are localized along structures that formed during Yanshanian orogeny	Deposits are exposed in windows (antiforms) through the Roberts Mountains thrust or shale–limestone contacts; deposits are localized along pre-existing structures that may have originated as normal faults during Late Proterozoic continental rifting
Deposit controls	Intersections of near vertical faults and fractures with permeable reactive strata	Intersections of near vertical faults and fractures with permeable reactive strata and locally low angle faults
Host rocks	Cambrian to Triassic calcareous marine sedimentary rocks deposited on a subsiding extended cratonic margin; locally igneous rocks as young as Cretaceous	Cambrian to Mississippian calcareous marine sedimentary rocks deposited on a subsiding extended cratonic margin; locally igneous rocks as young as Mid-Tertiary
Organic carbon		Locally, several percent carbon that is indigenous or introduced as petroleum and subsequently matured
Alteration	Silicification, argillization, sulfidation	Decarbonation, silicification, argillization, sulfidation
Ore and gangue minerals	Arsenian pyrite, arsenopyrite, orpiment, realgar, stibnite, cinnabar, thallium sulfides, quartz, calcite	Arsenian pyrite and marcasite, arsenopyrite, native arsenic, orpiment, realgar, stibnite, cinnabar, thallium-sulfides, quartz, barite, calcite
Residence of gold	Submicron inclusions in arsenian pyrite and arsenopyrite	Mainly in arsenian pyrite and marcasite as submicron inclusions and solid solution
Paragenetic sequence	Early silicification and acid leaching; later orpiment, realgar, quartz, stibnite in open pores and fractures	Early acid leaching and sulfidation, silicification; later orpiment, realgar, calcite, quartz, stibnite, or barite in open pores and fractures
Geochemical signatures	S, Au, As, Sb, Hg, Tl, Ag, U Au/Ag = 1–20	S, Au, As, Sb, Hg, Tl, Ag, Ba, $\pm$ W, $\pm$ Te, $\pm$ Se, Au/Ag > 5
Homogenization temp.	200 $\pm$ 50 °C	200 $\pm$ 50 °C
Salinities	5 $\pm$ 4 equiv. wt% NaCl	4 $\pm$ 4 equiv. wt% NaCl
Gases	CO ₂ $\gg$ CH ₄ + CO, H ₂ S	CO ₂ $\gg$ CH ₄ , H ₂ S
Pressure/depth	105 to 330 $\times$ 10 ⁵ P/1.0 to 3.5 km	100 to 1,000 $\times$ 10 ⁵ P/1 to 5 km
pH	Moderately acidic 4 to 6	Moderately acidic 4.0 to 5.5
δ D of clays and fluid inclusions	–87 to –47‰	Most districts –160 to –120‰; Getchell district –155 to –40‰
δ^{18} O of quartz	9.6–27.5‰	1 to 26‰
δ^{13} C and δ^{18} O of carbonates	–9–2‰ and 12–19‰	–4 to 3‰ and 7 to 13‰
δ^{34} S sulfides	–24–17‰	–7 to 17‰
Source of H ₂ O	Variably exchanged meteoric	Variably exchanged meteoric in most districts; metamorphic or magmatic and meteoric at Getchell
Source of CO ₂	Sedimentary rocks	Sedimentary rocks $\pm$ metamorphic or magmatic
Source of H ₂ S	Sedimentary rocks	Sedimentary rocks $\pm$ metamorphic or magmatic
Source of gold	Sedimentary rocks	Sedimentary rocks $\pm$ metamorphic or magmatic contribution; $\pm$ from older mineralization in districts
Gold depositional mechanisms	Sulfidation, cooling, dilution, pH increase	Sulfidation, cooling, dilution, boiling

Hofstra 2000). In both countries, a combination of high paleothermal gradients and tectonic disturbance seem to have played an important role in the formation of the deposits (Tu 1990; Hofstra et al. 1999).

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