

Characteristics and origin of sediment-hosted disseminated gold deposits: a review

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Abstract

Sediment-hosted disseminated gold (SHDG) deposits comprise a major portion of the gold production and reserves in the US. Although presently known to be common only in western North America, SHDG deposits are a significant source of world gold production. These deposits are characterized by extremely fine-grained disseminated gold, hosted primarily by arsenian pyrite. Other metals show very little enrichment although in addition to As, anomalous concentrations of elements such as Sb, Hg, Tl and Ba are utilized as exploration tools. The host rocks are dominantly silty carbonates, but ore concentrations are also present in siliceous and silicified rocks as well as intrusive rocks. Alteration consists of decarbonatization, silicification (jasperoid formation) and argillization, which are arranged both spatially and temporally in that order. Argillic alteration is zoned from kaolinite-dominated cores to sericite-dominated margins. The deposits commonly exhibit significant structural (faults) and stratigraphic (composition/permeability) controls. Until the last few years, SHDG deposits were considered as near-surface, epithermal type deposits in origin. Because of their fine-grained nature and the lack of macroscopic features such as veins, it has proven quite difficult to extract geochemical data that are clearly related to their genesis. However, fluid inclusion data indicate pressures corresponding to depths of 2–4 km under lithostatic conditions. Temperatures are constrained by fluid inclusions and phase equilibria to near 225°C. Stable isotope data from alteration minerals and fluid inclusions indicate that the ore fluids were dominated by meteoric waters, some of which had clearly exchanged oxygen with wallrocks during their passage through the crust. Although the data vary, most ore fluids probably had δD values near -150‰ and $\delta^{18}O$ values ranging from -10 to $+5\text{‰}$. Sulfur isotope values reported from SHDG deposits span a wide range, from -30 to $+20\text{‰}$ (sulfides) and 0 to $>45\text{‰}$ (sulfates). Ore-related sulfides (pyrite, realgar) fall at the upper end of the range reported for sulfides. The alteration and mineral assemblage indicate the ore fluids were probably near neutral and gold was likely carried as a bisulfide complex. The depositional mechanism(s) probably included mixing, cooling and oxidation. These mechanisms are consistent with the observed alteration features, i.e. quartz precipitation, calcite dissolution and sericite–kaolinite coexistence. It also explains the presence of both siliceous ores containing native Au and sulfide ores containing Au in pyrite. The extreme variations in sulfur isotopes as seen at Post and fluid inclusion data from Carlin may be indicative of some phase separation ('boiling'), but such relations have not been documented in other deposits and the importance of phase separation to gold deposition appears minimal.

Keywords: carlin-type; gold ore; fluid inclusions; isotopes; ore-forming fluids; Nevada

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1. Introduction

Sediment-hosted disseminated gold (SHDG) deposits, also known as Carlin-type gold deposits, have been recognized as a distinct type of ore deposit only relatively recently. Although high-grade portions of these deposits, such as at Mercur, have been mined since the last century (Jewell and Parry, 1987), they were not recognized as a class until the discovery and development of Carlin as a bulk-tonnage mine between 1962 and 1965 (Hausen and Kerr, 1968). SHDG deposits did not become economically important until the mid-1970's after the price of gold was freed from US government price restrictions and allowed to rise to free market values. At present, SHDG deposits are among the most important fossil hydrothermal systems in the USA.

Until the past few years, the prevailing hypothesis regarding the formation of these deposits was one of near-surface hydrothermal systems (e.g., epithermal hot springs model; Rye, 1985; Radtke, 1985), related to Tertiary or recent extensional tectonics of the Great Basin. Many data have been gathered recently to contradict this model and suggest a deeper and older environment of formation. This paper synthesizes the available data on SHDG deposits and presents an integrated model for the formation of these deposits.

2. General description

Fig. 1 (updated from Bonham, 1986) shows the location of many of the major SHDG deposits in western USA. The characteristics of many of these deposits are described in a series of papers included in USGS Bulletin 1646 (Tooker, 1985). Although additional deposits which may be SHDG or similar deposits are known from locations as diverse as southern China (Ashley et al., 1991; Mao, 1991); southeast Asia (Sillitoe and Bonham, 1990; Garwin et al., 1995); and Peru (Alvarez-A. and Noble, 1988), the majority of deposits for which there are more than descriptive data available are located in the Great Basin of western North America. The discussion presented herein draws most heavily on these more detailed studies.

2.1. Host rock lithology

SHDG deposits consist of extremely fine-grained gold that is generally submicron to possibly atomic in size (Wells and Mullens, 1973; Bakken and Einaudi, 1986; Arehart et al., 1993a). Most of the gold is hosted by hydrothermal arsenian pyrite and quartz, with considerably lesser amounts present in carbonaceous matter and fine-grained phyllosilicates. In the Great Basin, the host rocks consist primarily of Paleozoic sedimentary rocks, ranging in age from Cambrian to Mississippian. A few deposits are located primarily in younger sedimentary rocks (e.g., standard deposit, Triassic) and minor mineralization also occurs in igneous rocks in some deposits (e.g., Post, Getchell). The most common host rock types are silty carbonates to calcareous siltstones and calcareous shales. Many of the large ore deposits (e.g., Carlin, Post/Betze) are hosted by the Silurian–Devonian Roberts Mountains and Popovich Formations (Evans, 1980; Fig. 2). Rocks of similar lithology are

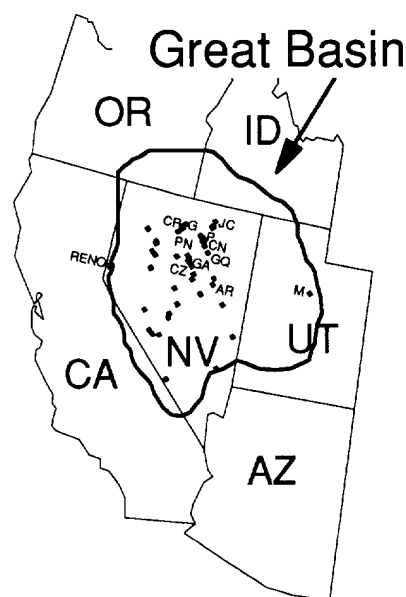


Fig. 1. Location map of the Great Basin showing major SHDG deposits. Modified and updated after Bonham, 1986). Deposits discussed in the text include: A = Alligator Ridge; CN = Carlin; P = Post/Betze; GQ = Gold Quarry, Rain; CZ = Cortez; GA = Gold Acres; G = Getchell; CR = Chimney Creek/Rabbit Creek and Pinson; JC = Jerritt Canyon; M = Mercur.

reported for other deposits (Rota, 1987; Ilchik, 1990; Osterberg, 1990; Ashley et al., 1991; Garwin et al., 1995; Wilson and Parry, 1995). In many orebodies, there is a very strong lithologic control on ore. For example, at Post/Betze, individual ore horizons representing original units of only a couple of meters in thickness can be traced over hundreds of meters laterally in drill core. In contrast to the favorability of rocks described above, igneous (dike and sill rocks) and clean carbonates or shales are generally less mineralized, and can form barriers to fluid flow.

2.2. Structure

There is a significant structural control on the localization of ore zones in SHDG deposits, consisting of both folds and faults. The general geometry of the ore zones comprises a feeder zone, commonly controlled by high-angle normal (or possibly reverse) fault zones and an upper stratiform zone. In competent brittle rocks having less favorable lithologies, the bulk of the ore may be in such structures or their branches, such as at Pinson (Madrid et al., 1988). Where gold-bearing solutions encountered reactive lithologies (which also are generally less competent structurally), larger and more tabular deposits were formed such as at Carlin (Bakken and Einaudi, 1986). The largest orebodies, such as Post/Betze, contain

both strongly fault-controlled and stratiform ores (Arehart et al., 1993a; Leonardson and Rahn, 1995).

In addition to fault control, there appears to be some primary fold control on the ores at some localities. Within the Carlin trend, many of the orebodies lie along the culminations of NNW-trending fold axes (Madrid and Bagby, 1986; Leonardson and Rahn, 1995). The Mercur deposit is on the margin of the Ophir anticline where fracturing associated with parasitic(?) flexures in the mine area may have played an important role in localizing mineralization (Kroko, 1992). Fold crests may have trapped or inhibited movement of ore fluids, or acted as a release point for overpressured hydrothermal fluids, all of which could have resulted in ore deposition. Although less structural information is available for other areas, it is likely that folding also influenced ore fluid migration and gold deposition in other deposits as well.

2.3. Hydrocarbons

These same fold crests (and gold ores) typically contain a significant amount of 'mature' hydrocarbon material and, in a regional sense, SHDG deposits are often correlative with organic-rich rocks. Maturation of organic material is evidenced by a decrease in atomic H/C ratio of kerogen (Ilchik et al., 1986), decreases in the abundance of short-chain

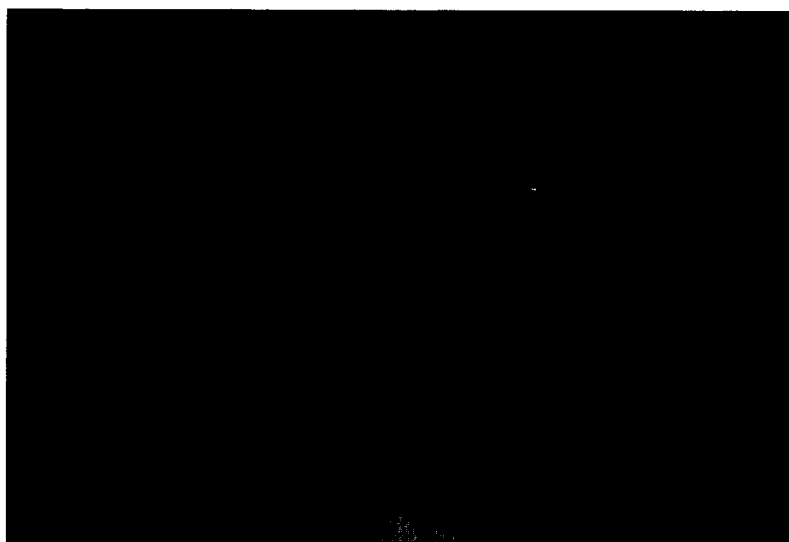


Fig. 2. Photograph of silty carbonate of the Roberts Mountains Formation in Maggie Creek Canyon near Carlin.

hydrocarbons (Kettler, 1990), and changes in reflectance values of pyrobitumen (Kuehn, 1989). There has been significant debate over the origin and role of these hydrocarbon materials. Hausen and Kerr (1968) reported gold “in mutual contact with...carbonaceous materials” (p. 931). Radtke and Scheiner (1970) proposed that the carbonaceous matter at Carlin effected gold deposition. In a later paper, however, Radtke (1985) noted that there is a poor correlation between gold and organic matter on a small scale (also at Carlin) and proposed that most or all of the organic material was introduced during the late stages of hydrothermal activity associated with the gold mineralization. In contrast, Ilchik et al. (1986); Ilchik (1990) suggested that the organic matter at Alligator Ridge is significantly older than the SHDG mineralization and was matured by elevated temperatures associated with hydrothermal activity. The speciation of *n*-alkanes is consistent with, but does not necessarily require, oxidative alteration of that organic matter at Alligator Ridge (Kettler, 1990). At Carlin, Kuehn (1989) and Kuehn and Rose (1995) concluded that hydrocarbon maturation took place

significantly prior to SHDG mineralization, perhaps as the result of a pre-SHDG hydrothermal or metamorphic event.

2.4. Hydrothermal alteration

Three major types of hydrothermal alteration have been recognized in SHDG deposits: decarbonatization, silicification (jasperoid formation) and argillization. The relative paragenetic position of these events and their relation to gold ore may vary between deposits, or within a single deposit, probably reflecting fluctuations in each hydrothermal system through time. For example, penecontemporaneous decarbonatization and silicification have been documented at Carlin (Bakken and Einaudi, 1986) and Alligator Ridge (Ilchik, 1990); gold mineralization is strongest in moderately to completely decarbonatized and weakly silicified portions of these deposits. In contrast, significant portions of the ore at Jerritt Canyon (Hofstra et al., 1988) and Pinson (Madrid et al., 1988) are in intensely silicified (nearly complete replacement) zones. In addition, pre-mineralization

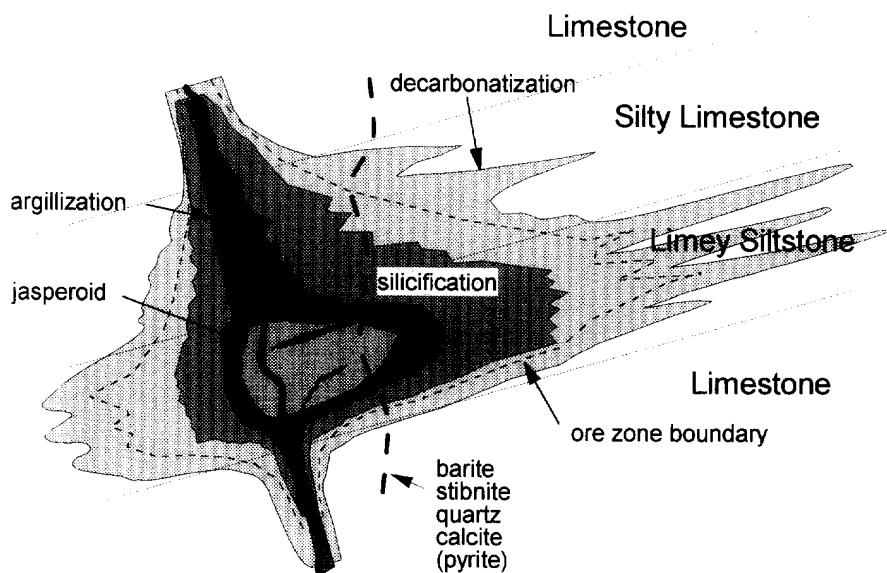


Fig. 3. Schematic cross-section through a SHDG deposit showing major alteration and mineralization features adjacent to a fluid feeder structure.

silicification has been described at Jerritt Canyon (Hofstra and Rowe, 1987; Northrop et al., 1987) and probably is present at other deposits as well. In the cores of these hydrothermal systems, intense argillization may be present (Arehart et al., 1992; Kuehn and Rose, 1995). The areal extent of phyllosilicates of hydrothermal origin is poorly documented, probably because of the sparse occurrence and fine-grained nature of these minerals, as well as the presence of preore (predominantly detrital) phyllosilicates in the host rocks.

In general, the sequence of alteration associated with gold mineralization proceeds from distal decarbonatization through silicification to argillization, both spatially and probably temporally (Fig. 3). Decarbonatization is in many cases the most areally

extensive ore-related alteration effect. At Alligator Ridge, however, Ilchik (1990) described calcification of the dolomitic host rocks as the most distal alteration effect. Decarbonatization comprises removal of the carbonate portion of the rock and is highly variable within a given deposit and between deposits. Strong stratigraphic control on decarbonatization is common, which appears to be related to original lithology. The most strongly decarbonatized rocks range from silty or sandy carbonates to calcareous siltstones or mudstones, probably because of original porosity/permeability characteristics. Non-permeable rocks, including mudstones and shales and purer carbonate beds are much less extensively decarbonatized. In areas of intense alteration, volume decrease, where present, may be indicated by thinned strati-

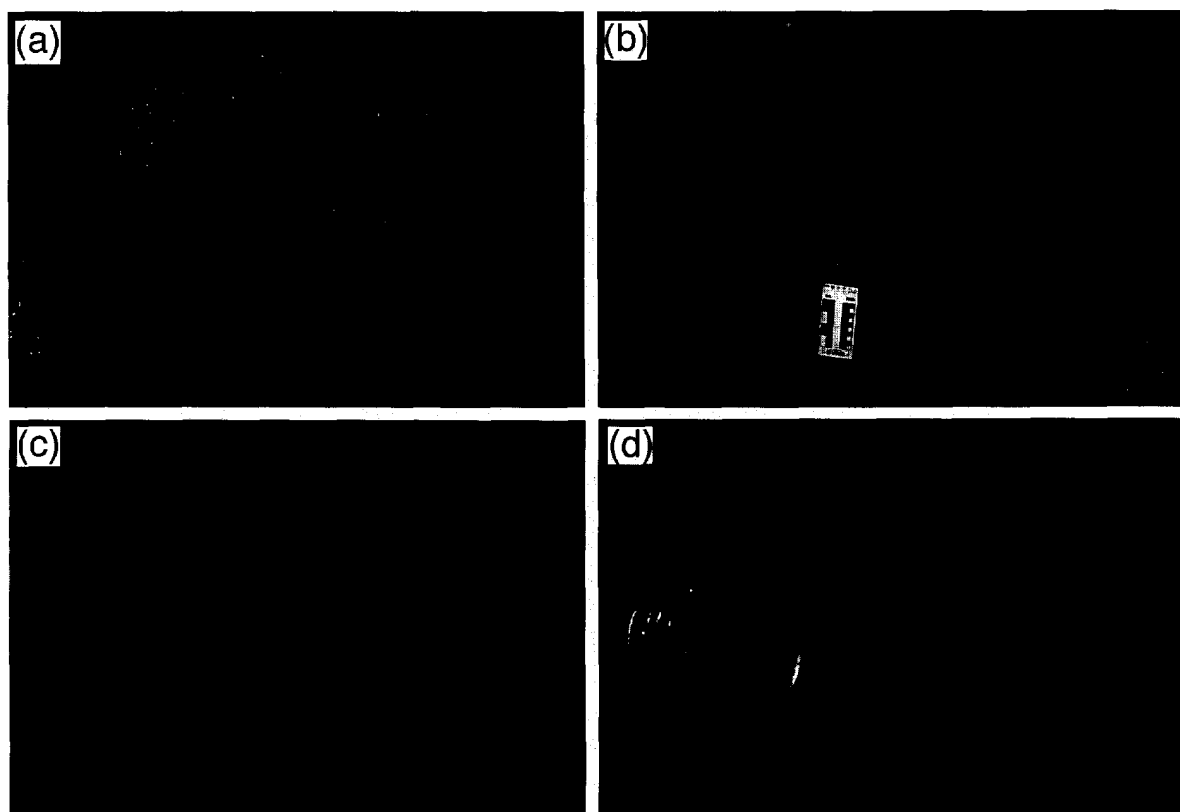


Fig. 4. (a) Photograph of weakly silicified (boxworks) carbonate rocks. Specimen is approximately 10 cm across. (b) Photograph of silicified breccia from Gold Quarry. Right bar scale is in cm. (c) Photomicrograph of jasperoid showing interlocking quartz crystals. Note lack of any structures indicating an original chalcedonic or amorphous silica structure. Field of view approximately 1 mm, crossed polars. (d) Photograph of intense argillic alteration (white pods) in jasperoid. Coin is 2 cm diameter.

graphic units, higher proportions of insoluble minerals, and in extreme cases, by disruption of bedding and brecciation (Bakken and Einaudi, 1986; Hofstra et al., 1991).

A second, extensive alteration type is silicification, producing a rock commonly referred to as jasperoid. Although initially the term jasperoid was used to describe intensely silicified carbonate rocks (Lovering, 1972), the usage has been expanded in the SHDG literature to include any extensive silicification associated with (or potentially associated with) gold mineralization. Silicification ranges from weak cementation-type effects and boxworks (Fig. 4) to more pervasive quartz replacements that grade into complete replacement (> 95% quartz; Fig. 4B and C) of the preexisting rock in many cases. As a result, these rocks form resistant outcrops and have been used extensively as prospecting tools. The two most common occurrences of intense silicification are as replacement of beds or groups of beds and along faults and along fractures. Generally, hydrothermal-related silicification is less extensive areally than decarbonatization. However, the genetic link between jasperoids and gold is obscure in many cases; therefore, some gold-related jasperoids may exist that are distal to ore. In addition, there are preore jasperoids that are genetically unrelated to ore in some areas (Hofstra and Rowe, 1987; Northrop et al., 1987).

The third major type of alteration is argillization. Argillization related to gold mineralization is difficult to discern macroscopically except in areas of very intense alteration (usually the core of the system, or in areas of igneous rocks). Peripherally, argillization takes the form of sericitization of feldspars (both detrital and primary igneous minerals), although this sericitization can be difficult or impossible to distinguish from sericitization related to non-gold related events. Some hydrothermal clay minerals (montmorillonite and kaolinite) are undoubtedly developed away from the center of the hydrothermal system, but it is generally quite difficult to establish a gold-related origin for these minerals where they occur distal to ore. Some attempts have been made at using crystallinity indices to outline the areal extent of hydrothermal phyllosilicates (Hauff et al., 1989), but in many cases it is unclear whether the more crystalline phyllosilicates

are related to hydrothermal events or pre-hydrothermal metamorphic events, or are simply mature detrital materials. Near the center of the hydrothermal system(s), the intensity of sericitization increases and the hydrothermal origin of these alteration minerals becomes somewhat more evident. In the core of the systems, sericite is replaced by a kaolinite-dominated phyllosilicate assemblage. In some cases (Fig. 4D), kaolinite is found as pods in jasperoid (Bakken and Einaudi, 1986; Arehart et al., 1993c), and may replace quartz. X-ray diffraction and differential thermal analysis studies at Carlin (Kuehn, 1989) and Alligator Ridge (Ilchik, 1990) have established that some of this kaolinite in the centers of the deposits is the variety dickite.

2.5. Mineral paragenesis and zoning

Fig. 5 is a generalized paragenetic diagram for SHDG deposits. Preore events (regional metamorphism and possibly hydrothermal) resulted primarily in recrystallization and remobilization of preexisting minerals, including quartz (jasperoid), and calcite, pyrite and barite (from bedded barite deposits common in the Paleozoic rocks). Gold is closely associated with hydrothermally-generated arsenian pyrite,

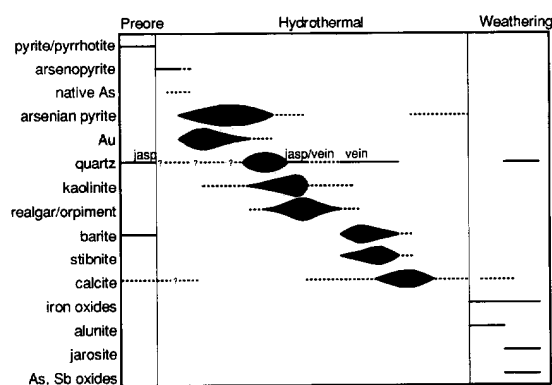


Fig. 5. Generalized paragenetic sequence for SHDG deposits from descriptions of Getchell (Joralemon, 1951), Carlin (Hausen and Kerr, 1968; Radtke, 1985), Cortez (Wells and Mullens, 1973), Jerritt Canyon (Birak and Hawkins, 1985), Chimney Creek (Osterberg, 1990), Rabbit Creek (Bloomstein et al., 1991) and Post/Be-tze (Arehart et al., 1993a). Minor and trace minerals are not shown because of their uncertain paragenetic position in most deposits.



Fig. 6. Composite back-scattered electron image (A) and secondary ion images (As (B) and Au (C)) of pyrite grains from gold quarry.

pyrite and locally arsenopyrite (Fig. 6) (Wells and Mullens, 1973; Mao, 1991; Arehart et al., 1993a). Temporally late in the gold stage are arsenic sulfides

(realgar and orpiment), that slightly postdate alteration minerals such as kaolinite and quartz. Significantly later in the paragenetic sequence are barite, stibnite and late calcite, commonly as open fracture fillings. A wide variety of trace minerals (not shown in Fig. 5), including thallium-bearing minerals (e.g., Radtke, 1985) probably are temporally associated with the main gold-depositing event. Alunite, jarosite and a variety of other oxide minerals commonly are present in the oxide zone of the orebodies and locally extend to depths of 700 m along fractures. These minerals are an oxidation product of the ores due to weathering (Arehart et al., 1992).

One of the most striking characteristics of many of these ore deposits is the lack of accessory minerals of hydrothermal origin, particularly base metal sulfides. Another distinctive characteristic is the relative lack of zonation of mineral occurrences. In many of the smaller and near-surface deposits, primary zoning is obscured because of extensive weathering effects. In deposits that still retain primary zoning features, many of the same minerals found deep in the deposits (realgar, orpiment, arsenian pyrite) also are found much closer to the surface, or at least evidence of their former presence is found. In some larger deposits, such as Post/Betze or Getchell (Arehart et al., 1993b; Joralemon, 1951), there is zoning of arsenic sulfide phases from arsenopyrite to native As to realgar/orpiment outward from the center of the deposits, but this zoning is over a large vertical and/or horizontal extent. Common late-stage minerals, such as calcite, barite and stibnite, are

Table 1

Summary of fluid characteristics of several SHDG deposits as inferred from fluid inclusions. Data from Hofstra et al. (1988); Jewell and Parry (1988); Osterberg (1990); Cline et al. (1995); Kuehn and Rose (1995)

Deposit	Temperature ^a	Salinity ^b	Gas composition ^c
Carlin main gold	185–245	2–4%	major CO ₂ ; minor H ₂ S, CH ₄ , SO ₂
Carlin late gold	175–250	< 1.5%	minor CO ₂ , trace H ₂ S, SO ₂
Jerritt I	200–250	3–10%	4% CO ₂ ; 1% H ₂ S ± SO ₂ ± HC ± Ar
Jerritt II	< 200	< 0.1%	< 0.5% CO ₂ ± H ₂ S ± SO ₂ ± HC ± Ar
Chimney Creek I	250–300	6.4–9.2%	CO ₂ present
Chimney Creek II	250–300	~ 2.8%	n.d.
Mercur	220–270	4.4–7.4%	CO ₂ present
Getchell mainstage	n.d.	n.d.	major CO ₂ ; minor N ₂ ± H ₂ S ± HC
Getchell late	n.d.	n.d.	minor CO ₂ ± N ₂ ± H ₂ S ± HC

^a Homogenization temperature, uncorrected.

^b Salinity in weight percent NaCl equivalent.

^c HC = short chain hydrocarbons. n.d. = not determined.

found throughout the ore zones, primarily in late veins, but with no apparent zoning.

3. Fluid inclusion data

Fluid inclusions in SHDG deposits are small and many are difficult to relate directly to gold mineralization. Initial observations of inclusions at Carlin (Nash, 1972) included a few inclusions that contained three-phase fluids (liquid water, liquid CO₂ and vapor CO₂). Subsequent more detailed observation (J.F. Slack reported in Radtke, 1985) yielded three types of inclusions: a liquid–vapor (liquid dominated) with salinities from 0 to 17 wt% NaCl equivalent; a liquid–vapor (vapor dominated) with salinities < 1.0 wt% NaCl equivalent and highly variable homogenization temperatures; and the three-phase inclusions described by Nash. These data were interpreted by Radtke (1985) as evidence of boiling in the Carlin system. Additional data have recently become available from this and other deposits and a relatively consistent pattern is emerging (Table 1).

Further detailed work at Carlin (Pasteris et al., 1986; Kuehn and Rose, 1995) suggests two major fluid events. High-salinity fluids (possibly basinal brines) were associated with the generation of oil and gas during late Paleozoic or early Mesozoic time and are unrelated to gold-mineralizing events. These fluids have salinities of 10–21 wt% equivalent NaCl and homogenization temperatures of the inclusions are $155 \pm 20^\circ\text{C}$. The gold-stage fluid inclusion assemblage is quite complex, comprising an array of inclusions having variable salinities and gas/liquid ratios. This assemblage is explained by the presence of two endmember fluids by Kuehn and Rose (1995): a gas-rich, moderate salinity fluid and a gas-poor, low salinity fluid (Table 1). The inclusion assemblage requires the existence of two-phase conditions (either boiling, effervescence, or immiscibility), at least intermittently. Ore-related fluid inclusions have homogenization temperatures of $215 \pm 30^\circ\text{C}$ (Kuehn and Rose, 1995). Calculated minimum trapping pressures based on homogenization of CO₂-rich fluid inclusions are 800 ± 400 bar (Kuehn and Rose, 1995).

Two hydrothermal episodes also are proposed for

the deposit at Chimney Creek based on fluid inclusion data (Osterberg, 1990). Earliest fluids predated gold mineralization and comprise saline brines (14–19.6 wt% eq. NaCl), probably basinal brines expelled by regional metamorphism. Two types of fluid inclusions were observed in gold-stage quartz and calcite: H₂O–CO₂ fluids of low salinity (~ 2.8 wt% eq.) and H₂O-rich inclusions of moderate salinity (6.4–9.2 wt% eq.). Homogenization temperatures are between 250 and 300°C for both types and calculated pressures are between 800 and 1000 bar.

In addition to an early regional metamorphic fluid, two fluids also are reported from the ore stage at the Jerritt Canyon deposit (Hofstra et al., 1988): (1) saline fluids with ~ 5 mol% gas and (2) very low salinity (< 0.1 wt% eq.) and low gas (< 0.5 mol%) fluids. Minimum trapping conditions reported by Hofstra et al. (1988) are 200–250°C and 500–1000 bar for saline inclusions. There was no evidence of phase separation reported in fluid inclusions at Jerritt Canyon.

In contrast to the complexity documented at Carlin, Chimney Creek and Jerritt Canyon, only one type of inclusion is reported from the Mercur deposits (Jewell and Parry, 1988). These two-phase, liquid-dominant inclusions have a wide range of homogenization temperatures but fairly restricted salinities. Homogenization temperatures for fluid inclusions hosted by jasperoid range between 220 and 270°C and salinities range between 4.4 and 7.4 wt% NaCl equivalent. Homogenization temperatures of hydrothermal calcite (usually with realgar) are 150–190°C and salinities between 3.3–8.9 wt% NaCl equivalent. Jewell and Parry (1988) reported a positive correlation between salinity and homogenization temperature. In addition, expansion of inclusion contents on crushing in oil is indicative of internal pressures in excess of atmospheric, attributed to the presence of significant CO₂ in the inclusions.

Trace gas analyses of fluid inclusions have been done at Jerritt Canyon (Hofstra et al., 1988), Carlin (Pasteris et al., 1986; Graney et al., 1991; Kuehn and Rose, 1995), Standard (Graney et al., 1991) and Getchell (Cline et al., 1995). The results indicate that the non-aqueous portion of the ore-related fluids comprises dominantly CO₂, with minor to trace amounts of N₂ + CH₄ + H₂S \pm SO₂ \pm short-chain hydrocarbons \pm Ar (Table 1). Graney et al. (1991)

used quadrupole mass spectrometry to measure fluid gas compositions and ratios from fluid inclusions in jasperoid at Carlin and Standard. They noted that jasperoids from mineralized areas exhibit higher H_2S/CO_2 and higher O_2 than do unmineralized jasperoids. N_2/Ar ratios are near 38, approximating air-saturated water values.

In summary, fluid inclusion data support the presence of two hydrothermal fluids during gold mineralization in most cases, but there is limited evidence for immiscibility or boiling. Both fluids are relatively low salinity, although one component appears to be more saline than the other and also contains significant dissolved gas. The contribution of CO_2 to salinity estimates for these fluids has not been quantified; in light of the sometimes significant amounts of CO_2 in the inclusions, true salt contents are likely to be lower than reported NaCl equivalents. In addition to the gold-related fluid events, jasperoids in some areas record earlier fluid events, possibly related to basinal brine movements. Estimated pressures are of the order of 1 kb, corresponding to depths of at least 2–3 km (at lithostatic) and 6–7 km (at hydrostatic).

4. Stable isotope data

4.1. Sulfur

Sulfur isotope values reported from SHDG deposits span a wide range, from -30 to $+20\text{‰}$ in sulfides and 0 to $>35\text{‰}$ in sulfates (Fig. 7). Ore-related sulfides (pyrite, realgar) span nearly the entire range reported for sulfides. Because of the fine-grained nature of the gold, and the difficulty of correlating gold with any of these sulfides, it has been difficult to interpret the wide spread in the sulfur isotope data.

It has long been observed that chemical and optical zoning may exist in many sulfides, and it has been shown recently that many sulfides are isotopically zoned as well (Eldridge et al., 1994). Recent work on the sulfides of the Post/Betze SHDG deposit by Arehart et al. (1993b) using the SHRIMP technique for analysis of ore sulfides has revealed extreme sulfur isotope zoning in some pyrite. This isotopic zoning correlates with chemical zoning of As and Au in the pyrite. Pyrite having the highest

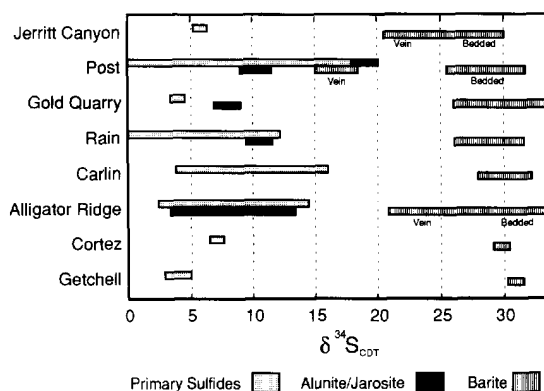


Fig. 7. Sulfur isotope data from SHDG deposits. Modified from Arehart et al., 1992. Doubly cross-hatched area at Post/Betze represents the calculated and measured $\delta^{34}S$ values of gold-bearing pyrite (see Arehart et al., 1993b). $\delta^{34}S$ values of vein and bedded barite, where distinguished, are indicated.

$\delta^{34}S$ values also contains the highest concentrations of gold. These zones also contain high concentrations of As (but not all high-As zones contain Au). Calculated and measured $\delta^{34}S$ values of gold-bearing pyrite are near $+20\text{‰}$. In contrast to the gold-rich zones, preore pyrite has $\delta^{34}S$ values ranging from -5 to $+10\text{‰}$ (-5 to $+7\text{‰}$ in sedimentary rock, $+6$ to $+10\text{‰}$ in igneous rock). Postore arsenian pyrite and marcasite (non-gold-bearing) has $\delta^{34}S$ values between -12 and -30‰ . Such small-scale isotopic measurements have not been made for other deposits.

Sulfur and oxygen isotope values of sulfates that are spatially associated with the ore also are highly variable (Figs. 7 and 8). Some of these values are clearly related to primary sedimentary sulfate, which is abundant in the Paleozoic section in the Great Basin. Other sulfates are clearly related to postore processes (e.g., alunite, Arehart et al., 1992) and have inherited their sulfur (and consequently their $\delta^{34}S$ values) through the process of near-quantitative oxidation of sulfides (Field, 1966). When these non-hydrothermal sulfates are excluded from the data, the remaining sulfates (barite) have $\delta^{34}S$ values ranging between $+15$ and $+25\text{‰}$ and $\delta^{18}O$ values near zero. Although these sulfates are clearly hydrothermal, in most cases barite veins are later than the gold mineralization. Commonly, barite forms late vein-filling crystals along with quartz, stibnite and calcite,

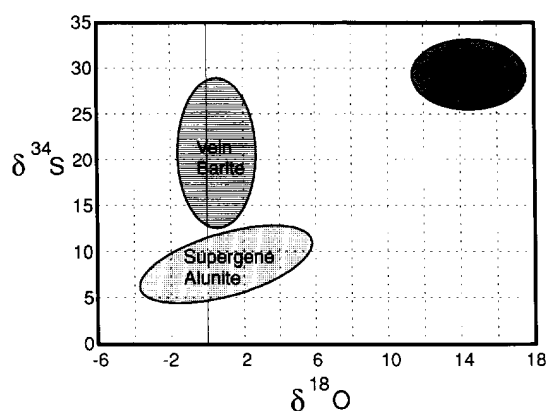


Fig. 8. Sulfur and oxygen isotope data from sulfates from SHDG deposits. Data from Arehart et al., 1992, Howe et al. (1995) and G.B. Arehart, unpublished data.

typically growing in open fractures (Fig. 3). In a few places at Post/Betze, similar late veins contain arsenian marcasite that hosts no gold and has very negative $\delta^{34}\text{S}$ values (-30 to -12‰) (Arehart et al., 1993b). Although no published isotopic data are available for other deposits, veins of a similar character are present in many orebodies.

4.2. Oxygen/hydrogen

Hydrogen and oxygen isotopes have been measured on a variety of minerals and some inclusion fluids from SHDG deposits. Because the amount of hydrogen in the fluid far exceeds that in the rock through which it passes, δD values from fluid inclusions and minerals have been little changed from their source fluid(s) values. δD measurements made on fluid inclusions comprise direct measurement of the fluids of interest, whereas the isotopic composition of waters must be calculated from measurements made on kaolinite. Data presented in Fig. 9 are calculated based on the kaolinite–water fractionation factors of Gilg and Sheppard (1996). For most SHDG deposits, the inferred hydrogen isotope values range between -140 and -170‰ . In contrast, δD values from Jerritt Canyon cluster near -110‰ .

Oxygen isotope values for the hydrothermal fluid(s) must be inferred (calculated) from measurements made on ore-related alteration phases. The one exception to this is the datum measured for fluid

inclusion waters hosted in realgar and orpiment (G.B. Arehart, unpublished data), where there is no possibility of retrograde oxygen exchange between the host mineral and inclusions. The most common alteration phase on which oxygen measurements have been made is quartz (jasperoid). Jasperoid $\delta^{18}\text{O}$ values range widely between deposits and even within deposits (O'Neil and Bailey, 1979; Holland et al., 1988; Hofstra et al., 1988). At Jerritt Canyon, there is a strong correlation between $\delta^{18}\text{O}$ of jasperoid and gold grade (Hofstra et al., 1988); Holland et al. (1988) documented higher $\delta^{18}\text{O}$ values of jasperoids from ore deposits versus non-economic prospects. Most investigators have interpreted the most positive oxygen values as being representative of the ore-bearing fluid (or at least the most exchanged fluid), although caution must be taken to ensure that the mineral measured is truly ore-related (see Hofstra et al., 1988). Similar wide ranges of $\delta^{18}\text{O}$ values exist at other deposits (Fig. 9), including Alligator Ridge (Ilchik, 1990), Post/Betze (G.B. Arehart, unpublished data) and Carlin (Radtke, 1985; Kuehn, 1989),

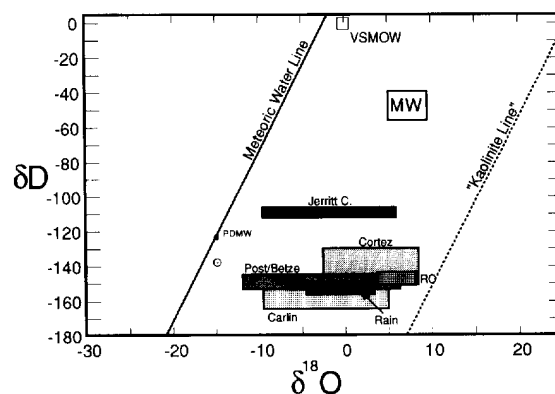


Fig. 9. Calculated isotopic composition of fluids responsible for ore deposition, based on fluid inclusion measurements and hypogene kaolinite (δD values) and equilibrium fractionation between quartz (jasperoid) or kaolinite and water at temperatures between 200 – 250°C ($\delta^{18}\text{O}$ values). Fractionation factors are from Clayton et al., 1972 (quartz–water) and Gilg and Sheppard, 1996 (kaolinite–water). RC = Rabbit Creek/Chimney Creek. MW = 'magmatic waters.' ■ = Approximate composition of present-day meteoric water (PDMW) from the northern Great Basin, ○ = direct measurement of fluid inclusion fluids from late-stage realgar/orpiment at Post/Betze. Data summarized from Rye, 1985; Radtke, 1985; Hofstra and Rowe, 1987; Hofstra et al., 1988 and G.B. Arehart, unpublished data.

although no detailed correlations between $\delta^{18}\text{O}$ and gold grade have been published. Based on measured δD values, temperatures between 200 and 250°C, and equilibrium fractionation between quartz and water, the calculated isotopic composition of the fluids responsible for several SHDG deposits are shown in Fig. 9.

The most obvious conclusion that can be drawn from these data is that SHDG deposits appear to have formed from evolved meteoric waters. Many of the deposits have similar isotopic signatures for their waters, although Jerritt Canyon stands out as significantly different. The majority of deposits for which there are data appear to have formed from meteoric waters that were significantly lower in δD values than present-day meteoric waters. Jerritt Canyon data suggest formation from waters having slightly higher δD values than those of the present-day northern Great Basin (Fig. 9). It is unclear, given the available age constraints (see below) whether this difference represents temporal variations in the isotopic composition of source waters, geographical differences such

as elevation of the source region(s), or some combination of factors.

4.3. Carbon/oxygen

Isotopic data from carbonate minerals (dominantly calcite and dolomite) may provide some insights into sources of fluid components and water-rock interaction. Fig. 10 summarizes the published data from SHDG deposits. Unaltered carbonate rocks which host SHDG deposits generally have $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of +18 to +24‰ and –2 to +1‰, respectively. There is a broad decrease in $\delta^{18}\text{O}$ in the host rock with increasing hydrothermal alteration. $\delta^{13}\text{C}$ values in rock may decrease or increase slightly in a general sense; however, locally (adjacent to fractures and veins) there is a clear increase in $\delta^{13}\text{C}$ values and decrease in $\delta^{18}\text{O}$ values as veins are approached (Ilchik, 1990; G.B. Arehart, unpublished data). Whether these changes are the result of reequilibration of the rock with the fluid at higher temperatures or input of hydrothermally-sourced CO_2 is as yet undetermined. With some exceptions, there are similar changes in isotopic values of ore-related veins on a broad temporal scale within individual ore systems (Fig. 10) to lower $\delta^{18}\text{O}$ values and higher $\delta^{13}\text{C}$ values. In addition to the hypogene assemblage, there is a distinct isotopic signature to supergene (oxidation-related) veins, which have lower $\delta^{13}\text{C}$ values, consistent with derivation from atmospheric CO_2 and meteoric water.

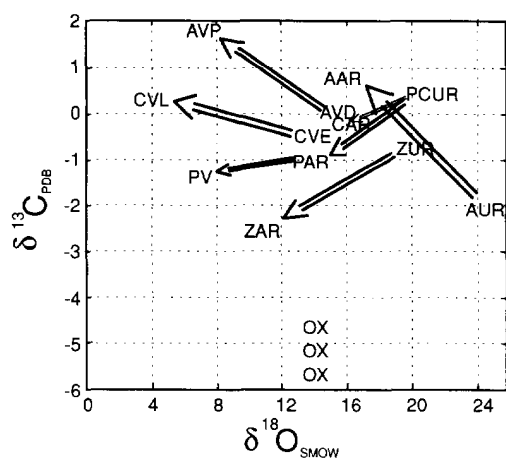


Fig. 10. Carbon/oxygen isotope data from rock and vein carbonates in SHDG deposits. AAR = Alligator Ridge altered rock; AUR = Alligator Ridge unaltered rock; AVP = Alligator Ridge proximal gold-related veins; AVD = Alligator Ridge distal gold-related veins; PCUR = Post and Carlin unaltered rock; CAR = Carlin altered rock; CVE = Carlin early gold-related veins; CVL = Carlin late gold-related veins; PAR = Post/Betze altered rock; PV = Post/Betze gold-related veins; ZUR = Cortez unaltered rock; ZAR = Cortez altered rock; OX = oxide stage calcite veins, all deposits. Data summarized from Rye et al., 1974; Radtke, 1985; Ilchik, 1990; Kuehn, 1989; Osterberg, 1990; and G.B. Arehart, unpublished data.

5. Gold transport and deposition

5.1. Gold transport

Gold is soluble under acid, oxidizing conditions as a chloride complex (Cloke and Kelly, 1964; Henley, 1973) or under near-neutral, reducing conditions as a bisulfide complex (Seward, 1973; Shenberger and Barnes, 1989). Fig. 11 is a plot of gold solubility in $f\text{O}_2$ -pH space; the stability fields of a variety of other common hydrothermal minerals are also plotted. From these data it is clear that the only reasonable mechanism for transport of gold in fluids ultimately responsible for SHDG deposits is as a bisulfide complex $[\text{Au}(\text{HS})_2^-]$ in near-neutral, reducing

environments. This is supported by the phase assemblage present in the deposits (kaolinite–sericite–pyrite–organic matter \pm arsenopyrite \pm arsenic). In addition, the lack of significant base metals and relatively low salinity of fluid inclusions in SHDG deposits suggests that chloride transport was not operative.

5.2. Source of fluid components

Sulfur is the component for which there is the best indication of likely source rocks. Sulfur isotope signatures for ore-stage sulfides (pyrite and As sulfides) are generally in the range of $+10$ to $+20\text{‰}$ (Rye et al., 1974; Radtke, 1985; Ilchik, 1990; Osterberg, 1990; Arehart et al., 1993b, see also Fig. 7). In contrast, sulfides from igneous rocks have compositions between 0 and $+8\text{‰}$. Sulfides in Paleozoic sedimentary rocks in the Great Basin have $\delta^{34}\text{S}$ values ranging from -15 to $+5\text{‰}$. Paleozoic sulfates have $\delta^{34}\text{S}$ values in excess of 25‰ . Because it

is difficult to fractionate the preore sulfides to yield the values observed in the ore-stage sulfides, the most likely source of sulfur is sulfates that have been reduced. Rayleigh-type fractionation of sulfate sulfur during reduction processes could easily yield the observed $\delta^{34}\text{S}$ values of ore sulfides.

Gold is present in anomalous abundances (above global averages for similar rock types) in several of the Paleozoic rock units in the Great Basin (C.J. Zimmerman, personal communication, 1989). It is unclear if this represents a real regional enrichment that took place prior to the time of SHDG mineralization. If it represents elevated pre-SHDG gold contents, it is tempting to hypothesize that much or all of the gold could have been relatively locally derived. A second possibility is derivation of gold from deeper in the circulation path. Neither of these hypotheses have any supporting data. Model calculations by Ilchik and Barton (1995) using a variety of source rocks generated sufficient gold in solution to account for the gold found in the ore deposits,

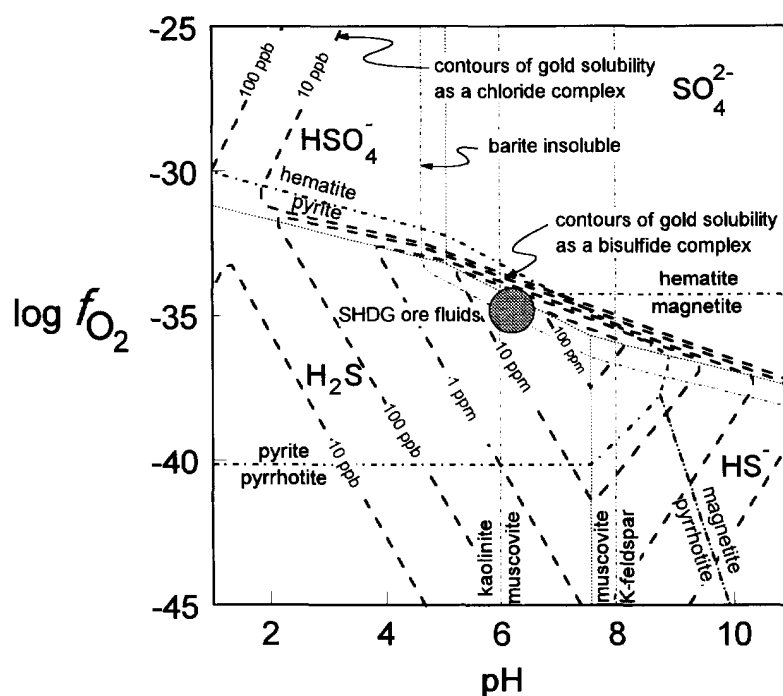


Fig. 11. f_{O_2} versus pH diagram showing stability fields of various minerals, aqueous sulfur species and gold solubility contours. $T = 250^\circ\text{C}$, $\Sigma\text{S} = 0.05 \text{ m/kg H}_2\text{O}$, $\text{K}^+ = 10^{-4} \text{ m/kg H}_2\text{O}$, $\text{Ba}^{+2} = 0.02 \text{ m/kg H}_2\text{O}$; barite soluble/insoluble assumed at $m_{\text{Ba}} \cdot m_{\text{SO}_4} = 10^{-3}$. Data from Ohmoto et al., 1983; Shenberger and Barnes, 1989; Helgeson et al., 1978.

yielding no constraints on source rock type. The general lack of base and other chloride-complexed metals, as well as the low salinities of fluid inclusions would suggest that any magmatic connection is probably distal (Sillitoe and Bonham, 1990). However, based on the available data, such magmatic contributions to the gold budget cannot be ruled out.

5.3. Gold deposition

Deposition of the gold could be effected by the destabilization of the bisulfide complex in a number of different ways, including phase separation (boiling), mixing (dilution), oxidation, reduction, changes in pH or temperature and sulfidation.

Temperature decrease is an excellent method of depositing jasperoid while dissolving calcite (Fournier, 1985; Fournier, 1986). However, the stability constants for gold bisulfide complexes change very little with temperature over the range 150–300°C and in fact may be retrograde under some conditions (Seward, 1973; Shenberger and Barnes, 1989; Hayashi and Ohmoto, 1991). Therefore, for gold complexed as a bisulfide, temperature decreases alone are not very effective in depositing that gold.

Although *reduction* is a possibility for gold deposition, there is currently no evidence supporting this as an active mechanism. Early studies (Hausen and Kerr, 1968; Radtke and Scheiner, 1970) suggested that organic matter played an active role in gold deposition; however, these studies assumed that chloride complexes were the major contributors to gold solubility. More recent work suggests that organic matter played no active role in gold deposition (see previous discussion of organic matter). Redox reactions have negligible effects on reactions involving carbonate dissolution or jasperoid formation, and would not affect (directly) conversion of feldspars to illite and clay minerals (Fig. 11). In addition, mineral zoning in the largest of these deposits is from arsenopyrite (more reduced) in the very center of the deposits to pyrite-realgar away from the centers (more oxidized), and late stages of mineralization are characterized by the presence of barite (an oxidized phase) suggesting that the fluids increased in oxidation state through time and space. Therefore it is unlikely that reduction was responsible for gold deposition.

Oxidation across the sulfide–sulfate boundary (Fig. 11) results in a sharp decrease in bisulfide concentrations and can be a particularly effective precipitation mechanism (Romberger, 1986). However, the paucity of gold in or associated with barite, and abundant gold in arsenian pyrite suggests that gold deposition took place before the fluids were out of the pyrite stability field. Because the solubility maximum for gold bisulfide is below the sulfide–sulfate boundary, oxidation could cause gold deposition while the system is still in the pyrite stability field. Because of the close spacing of the gold solubility contours, only slight changes in the oxygen fugacity can cause nearly complete precipitation of dissolved gold. Oxidation as a mechanism is consistent with observations of the gold distribution in the pyrite rims on a microscopic scale (Arehart et al., 1993a). The highest gold concentrations in pyrite rims are in the earliest-deposited portions of the rims, suggesting rapid and almost quantitative deposition of gold. Phase relations among As-bearing minerals (arsenopyrite, native As, realgar, orpiment) also are consistent with oxidation in time and/or space as is the presence of paragenetically late barite. However, as in the case of reduction, oxidation has no effect on silicate and carbonate phase relations; therefore, oxidation by itself cannot account for all of the observed features in SHDG deposits.

Any *pH change* away from near the H_2S – HS^- boundary should result in deposition of gold from bisulfide complexes because of the dependence of solubility on the activities of both H_2S and HS^- . However, gold solubilities below $\text{pH} \sim 5$ may be essentially constant due to the presence of a neutral gold bisulfide complex (Hayashi and Ohmoto, 1991). If this is the case, little gold deposition will accompany decreasing pH. Increasing pH, while effective in depositing gold, is incompatible with the observed alteration effects (quartz deposition, calcite dissolution). In addition to the above considerations, pH changes are likely to have been buffered at near-neutral values by reactions such as sericite–K feldspar, magnetite–pyrite (at fixed oxygen fugacity) and carbonate equilibria. Therefore, changes in pH are unlikely to have been a major cause of ore deposition.

Boiling has been shown to be an effective mechanism for gold deposition in a variety of epithermal systems (e.g., Drummond and Ohmoto, 1985). Boil-

ing effects gold deposition as H_2S is rapidly removed from the liquid portion of the hydrothermal system and the concentration of the bisulfide ligand decreases. Fluid inclusion data presented by Radtke (1985); Kuehn and Rose (1995) were interpreted to suggest that phase separation may have occurred sporadically and locally in the Carlin deposit; however fluid inclusion data from Chimney Creek (Osterberg, 1990) and Jerritt Canyon (Hofstra et al., 1988) do not require phase separation. Boiling is a particularly ineffective mechanism for generation of jasperoid, as it generally decreases calcite solubility (Fournier, 1986; Simmons and Christenson, 1994), making limestone replacement unlikely. In addition, the most likely silica phase to form associated with a boiling system is amorphous silica or chalcedony (Fournier, 1985). No evidence for either of these polymorphs has been described in SHDG deposits. However, late-stage (post-gold) phase separation may be indicated by extreme fluctuations in sulfur isotope values in late-stage arsenian pyrite at Post/Betze (Eldridge et al., 1994).

Although *sulfidation* (generally of iron) can be an effective mechanism for gold deposition by decreasing the concentration of H_2S in solution, little has been done to document this mechanism in SHDG deposits. Hofstra et al. (1991) briefly described measurements and modeling calculations suggesting that sulfidation may have been operative at Jerritt Canyon, but gave no details, such as petrographic evidence. Sulfidation should have no effect on calcite or quartz solubilities, which suggests that if operative, it was not the only process occurring.

Another process that may account for gold deposition in these deposits is fluid *mixing*. Mixing of a relatively gold bisulfide-rich fluid with a second fluid (at similar P and T) having a relatively lower concentration of H_2S would dilute the ore fluid, decreasing the H_2S concentration, and therefore gold solubility. Although gold deposition can be accomplished by this mechanism, simple mixing of two fluids (at similar temperatures) are unlikely to cause either deposition of quartz (jasperoid) or dissolution of carbonate (e.g., Fournier, 1985). If the fluids are at different temperatures, however, then quartz deposition and calcite dissolution are more likely to occur.

Given the limitations of processes described

above, none of which adequately explains the observed features, the more reasonable model involves several of these processes operating simultaneously. Because of the very high grades found locally in these deposits, the model fluid must either rapidly deposit its dissolved gold load or have a relatively high residence time in the future ore site. A boiling epithermal model fails because of its lack of explanation of alteration features and mineral zonation. Boiling systems should deposit both silica (generally as amorphous silica) and carbonate. This is particularly true of carbonate, considering the very high concentrations of CO_2 reported from some fluid inclusions (e.g., Nash, 1972; Kuehn and Rose, 1995). Neither of these features is found in SHDG deposits. In addition, boiling systems commonly have distinctive metal zonations, a consequence of their proximity to the paleosurface and locally high geothermal gradients.

It is most likely that more than one mechanism is responsible for the formation of all of the observed features of SHDG deposits. A most reasonable model that fits all of the observations is that of mixing of two fluids: the reduced, gold bisulfide-bearing solution at relatively higher temperature with a cooler, oxidized, less saline groundwater at the site of a structural breach in an overpressured reservoir. This would result in oxidation, dilution and cooling (of the ore fluid). The major causes of gold deposition would likely be oxidation and dilution. Locally this breach of a pressure seal may have caused pressures to reduce enough to generate immiscibility (CO_2 effervescence?) evidenced at Carlin (Kuehn and Rose, 1995). Dissolution of calcite in the host rock would be effected by the CO_2 -charged fluid; significant undersaturation with respect to calcite would be likely in mixed fluids on temperature constraints alone (Ellis, 1959). Relatively slow cooling would promote the formation of jasperoid rather than other forms of silica, and if the fluid compositions were appropriate, mixing could add to jasperoid precipitation.

6. Age of mineralization

There are no data on the age of SHDG type deposits from locations outside of the Great Basin;

therefore, this discussion is restricted to the age of deposits in the Great Basin. Because of the paucity of veins or alteration minerals that are clearly related to gold mineralization and that contain good material suitable for radiometric dating, the ages of these deposits are poorly established at best. Until recently, most deposits were considered to be quite young (upper half of the Tertiary) and were thought to have formed in epithermal, hot springs-type environments because of the association with typical epithermal elements such as As, Sb and Hg (e.g., Radtke, 1985). There are a wide variety of other mineral deposit types (including epithermal veins) throughout the Great Basin that have been shown to be Tertiary; by inference SHDG deposits have often been grouped in the same age bracket. Recent information (see discussion below) indicates that these deposits probably formed deeper in the crust, and therefore the conclusion that “if it is hot-springs type, it must be young” has been called into question.

6.1. Radiometric age data

There are only scattered radiometric age data available for SHDG deposits, many of which are from rocks or minerals not demonstrably related to mineralization. Reported ages range from 152 Ma (Mercur: Wilson and Parry, 1995) to as young as 8 Ma (Alligator Ridge: Ilchik, 1990). These age data fall into three general groups: (a) alunite minerals, (b) biotite, feldspar and whole-rock and (c) sericite (Fig. 12). Of the data, the first two groups are unlikely to record events that are genetically linked to gold mineralization. Alunite in SHDG deposits has been demonstrated to be supergene and therefore unrelated to gold mineralization (Arehart et al., 1992). The biotite, feldspar and whole rocks that have been dated comprise minerals that are normally altered by hydrothermal activity associated with SHDG mineralization. Therefore, if these minerals (and rocks) are fresh enough to be dated, they are not reliable indicators of the timing of hydrothermal activity. However, they may place lower limits on the age if the minerals are demonstrably younger than gold mineralization. For example, Maher et al. (1993) described two SHDG occurrences in the Roberts Mountains on which they place minimum age constraints of 33 to

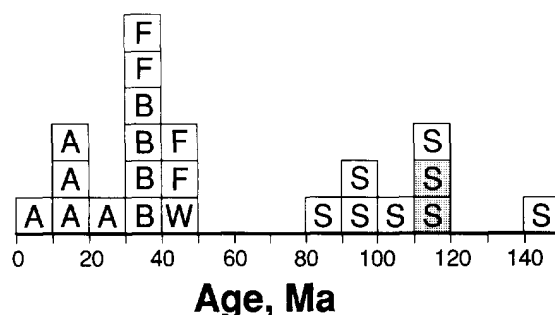


Fig. 12. Summary of published radiometric age data reported for SHDG deposits. Dates reported are from: A = alunite; F = feldspar; B = biotite; W = whole rock and S = sericite. Shaded blocks are data from $^{40}\text{Ar}/^{39}\text{Ar}$ stepheating experiments. Data from sources cited by Arehart et al. (1993c) and from Maher et al. (1993), Groff et al. (1995) and Phinisey et al. (1995).

34 Ma. These limits are based on K/Ar dates on biotite and plagioclase in unaltered igneous rocks that postdate mineralization. At one of these deposits (Tonkin Springs), altered rocks that overlie the gold deposit give K/Ar dates of 37 Ma (biotite from fresh rock away from hydrothermal alteration) and “if the hydrothermal alteration is related to gold mineralization” (Maher et al., 1993 p. 471), the age of the deposit is between 37 and 33 Ma.

The third group, comprising sericites, are more likely to be hydrothermal alteration products related to gold, although in most cases the connection between sericite and gold is not well documented. There has been only one detailed study of the age of a SHDG deposit published (Arehart et al., 1993c). At the Post/Betze deposit, age limits are defined by pre- and post-ore igneous events that are well dated by $^{40}\text{Ar}/^{39}\text{Ar}$ stepheating techniques at 157 Ma and 39 Ma respectively. Although the radiometric age data on hydrothermal minerals (sericite) are somewhat scattered, they cluster between 100 and 117 Ma. In addition, fission-track annealing data on zircons in mineralized rock show a peak of thermal activity near 110 Ma. From these data, it was concluded that the hydrothermal event took place at approximately 117 Ma (Arehart et al., 1993c). Two samples of hydrothermal sericite from altered dikes at Carlin yielded $^{40}\text{Ar}/^{39}\text{Ar}$ stepheating dates near 120 Ma, similar to the Post/Betze dates (Kuehn, 1989). Recent data from the Jerritt Canyon district (Phinisey et al., 1995) include several dates similar

to those of Arehart et al. Sericite from an altered andesite dike gave an age spectrum from 116 to 140 Ma. In contrast to Post/Betze, these dates are interpreted as a hydrothermal event not related to SHDG mineralization. A date from an unaltered basalt dike near the ore deposits yielded an age of 41 Ma; the same dike is reported to be altered and locally ore grade nearby. A hornblende separate from a quartz monzonite dike that is interpreted to be postore (but is from 3–5 km from the ore zone) yielded an age of 39 Ma. From these data, Phinisey et al. (1995) interpreted the age of SHDG mineralization at Jerritt Canyon to be between 41 and 39 Ma. Folger et al. (1995), in a companion study, attempted to evaluate the applicability of dating of sericites to SHDG deposits. At Jerritt Canyon, they reported an age of 402–435 Ma for the 20–40 micron fraction of sericite, which is similar to the age of diagenesis. Finer-grained fractions ($< 1 \mu\text{m}$) yielded younger ages (149 and 285 Ma). Based on geochemical data indicating K immobility, they concluded that because no new sericite was likely to form, any dates obtained must be the result of resetting of the isotopic clock of the preexisting micas. The two ages (149 and 285 Ma) were suggested to be partially reset detrital/diagenetic sericites. They further concluded that most or all of the sericite dates reported for other SHDG deposits are suspect unless a distinct gold-related sericite fraction can be analyzed separately. In contrast to K immobility or depletion at Jerritt Canyon, Radtke (1985) reported K enrichment in the ore zones at Carlin, indicating that hydrothermal K-bearing phases such as sericite are likely to have formed. Hydrothermal adularia also is reported to be present at the Twin Creeks deposit (Groff et al., 1995), also indicating K mobility.

Even given the caveats described above, it is interesting to note that most of the data from sericites fall in a reasonably limited age range, 80–130 Ma (Fig. 12). This age range includes the more detailed study at Post/Betze (Arehart et al., 1993c), as well as the two $^{40}\text{Ar}/^{39}\text{Ar}$ stepheating data from Carlin (Kuehn, 1989) and a 'hydrothermal event' age at Jerritt Canyon (Phinisey et al., 1995). This coincidence of Mesozoic ages was interpreted by Folger et al. (1995) as related to Mesozoic plutonism and not to SHDG mineralization. In contrast, the samples of Arehart et al. (1993c) were reasonably well-con-

strained as being related to gold mineralization rather than the pre-existing (and older) pluton.

6.2. Geological age constraints

Seedorff (1991) summarized geological information on a number of small deposits and prospects in Nevada that he classified as SHDG deposits. At several of these localities, alteration and/or mineralization are reported to be hosted in rocks of Tertiary age, thus establishing age maxima. Seedorff also noted the association of some of these mineralized zones with structures of mid-Tertiary age. He concluded that because the major structural control on alteration/mineralization in these occurrences is of Tertiary extensional origin, the deposits must be of a similar or younger age.

6.3. Summary

From the above discussion it should be clear that the ages of these deposits are still in question. Both radiometric and geologic constraints place the majority of deposits at older than ca. 35 Ma. The available radiometric age data suggest that most of these deposits in the Great Basin are Cretaceous in age, although data from Jerritt Canyon and the Roberts Mountains have been interpreted to indicate that some deposits may be Tertiary. Groff et al. (1995) suggested a multiple mineralization history for deposits in the Getchell area. A few small deposits and occurrences can be assigned a Tertiary age based on geologic constraints. It is possible that the deposits formed over a large time span between the oldest reported date (152 Ma) and ca. 35 Ma. More detailed work must be done on gold-related alteration phases at other deposits to establish the ages of the deposits more clearly and enable the deposits to be better related to tectonic events and environments.

7. Depth of formation

The best available estimates of the depth of formation of SHDG deposits come from fluid inclusion studies. Unfortunately, as noted above, most of the fluid inclusions at SHDG deposits are quite difficult to work with, both because of their size and uncer-

tain relationship (of any given inclusion type) to gold mineralization. In addition, there are significant uncertainties in correlating fluid inclusion phase relationships and homogenization temperatures to pressures at the time of trapping (see Roedder, 1984 for a detailed discussion). Given those uncertainties, minimum trapping conditions for ore-related fluid inclusions in jasperoids from Jerritt Canyon were estimated at 500–1000 bar by Hofstra et al. (1988). Pressure estimates for Carlin (800 ± 400 bar; Pasternis et al., 1986; Kuehn and Rose, 1995) and Chimney Creek (800–1000 bar; Osterberg, 1990) are very similar. Both Kuehn and Rose (1995); Osterberg (1990) found no evidence of spatial zoning in fluid inclusion compositions or homogenization temperatures across extensive orebodies. Although no specific estimates of total pressure were calculated, data from Mercur are consistent with pressures of near 800 bar (Jewell and Parry, 1988). These pressures translate into depths of 6–8 km (under hydrostatic conditions) or 2.5–3 km (under lithostatic conditions).

Seedorff (1991) made stratigraphic reconstructions of the cover over some SHDG occurrences that he considered Tertiary in age, from which he concluded that depths of formation ranged between 1.5 and 4.5 km. Because of the lack of age control, such reconstructions have not been attempted on most of the major deposits.

Additional qualitative information may be gained from studies of mineral zoning. Of the deposits described in the literature, none have any significant changes in primary ore mineralogy over the explored portions of the systems. The most vertically extensive orebody known is the Post/Betze system (Arehart et al., 1993a,b). Over a vertical interval in excess of 750 meters, there is very little change in mineral assemblage or textures. These data suggest that SHDG deposits form in areas with relatively low thermal gradients over the zone of mineralization, which necessitates formation in deeper portions of the earth's crust, in contrast to typical epithermal acid-sulfate or adularia-sericite systems.

8. Integrated geological and geochemical model

Any model which purports to explain the depositional process for SHDG deposits must account for

all of the observed features of these deposits. Primary among these are the paragenetic position of the gold and associated sulfide minerals, major alteration features, including decarbonatization and jasperoid formation, and fluid inclusion and stable isotope data. On a larger scale, the model must account for the age and tectonic environment of the deposits.

Based on the available data, the model which best fits is one of mixing of two fluids moderately deep (several kilometers) in the crust. Meteoric fluids (low δD and $\delta^{18}O$) circulated through the Paleozoic and Precambrian basement rocks where they probably picked up Au and S. Relatively abundant organic carbon in the Paleozoic section may have controlled the redox condition of the fluid, leading to dissolution of sulfates (barite). This abundant reduced sulfur made gold dissolution and transport by bisulfide complexes favorable. The ore fluids had high $\delta^{34}S$ values, inherited from the reduced sulfates. Metals that are commonly transported as chloride complexes were not mobilized. An alternative model for the source of some of the ore components (primarily Au) in magmatic fluids was proposed by Sillitoe and Bonham (1990), but there is little hard evidence for magmatic inputs to SHDG hydrothermal systems.

As this meteoric fluid travelled through the source rocks, it exchanged oxygen with the rock at high temperatures, with a consequent increase in the $\delta^{18}O$ value of the fluid. CO_2 was probably added from a different source, spatially closer to the deposition site, as any significant CO_2 in the fluid would have resulted in dissolution of the carbonate rocks through which it passed. Stable isotope data do not support an origin of this CO_2 from oxidized organic matter; the most likely sources are deep metamorphic fluids or from skarn formation accompanying igneous intrusion. Keeping this CO_2 dissolved in the fluid would require significant pressures, probably above normal hydrostatic, and may be maintained at relatively shallower depths by a pressure seal (Kuehn and Rose, 1995). Even so, depths of 2–4 km are required at lithostatic pressure.

Fluid upflow may have been focused along deep structures by either magmatic (thermal) events or tectonic events or some combination of the two. There is at least one major episode of igneous intrusion that corresponds well with many of the available dates for SHDG deposits (Arehart et al., 1993c).

Alternatively, it was postulated by Seedorff (1991) that the onset of extensional tectonics in the Great Basin may have allowed the release of some of these deep 'metamorphic' fluids to form SHDG deposits. Although Seedorff suggested that this was most likely a Tertiary event, there is some evidence that extensional tectonism took place in the Great Basin much earlier (Davis, 1980). The position of the orebodies in the Carlin trend suggest that these upflow zones were along the crests of folds, and similar structural features may have controlled fluid flow paths in other districts. Wherever these fluids breached the pressure seal and flowed into the overlying 'local' meteoric water contained in unaltered carbonate-bearing rocks (still at depths of several thousand meters), gold may have been deposited. The CO₂-charged deeper fluids were able to dissolve significant carbonate. Auriferous jasperoid was deposited because of the cooling effects of mixing of hot hydrothermal fluids with cooler 'local' meteoric water and consequent destabilization of the gold bisulfide complex. Mixing and sulfidation were the probable processes causing gold deposition, at least at the mixing front between the two fluids. Behind the active mixing front, diffusive oxidation due to mixing or gas exsolution may have resulted in the destabilization of the bisulfide complex and As species, and gold was deposited dissolved in metastable arsenian pyrite. Further evolution of the system resulted in more extensive oxidation in vein structures and precipitation of hydrothermal barite and quartz. The likely sources of the Ba and sulfate were the ore fluid and the 'local' meteoric fluid, respectively. Although there is evidence of boiling in the fluid inclusion data from Carlin (Kuehn, 1989) and sulfur isotope data (Arehart et al., 1993b; Eldridge et al., 1994), these effects were probably minimal in terms of causing gold deposition. In the case of the sulfur isotope data (Arehart et al., 1993b), the measurements are from very late-stage hydrothermal arsenian pyrites that do not contain gold. Some of the fluid inclusion measurements made by Kuehn (1989) are from similar material. It is possible, then, that phase separation (boiling) took place late in the history of these deposits, too late to have been an effective gold-depositing mechanism (that is, after the bulk of the gold had already been deposited).

Such a chemical model is consistent with a compressional tectonic environment, such as that which existed in the Great Basin in Cretaceous time. Overthrusting and stacking up of the sedimentary section as a result of compressional tectonics could lead to a regionally low geothermal gradient. This would allow maintenance of relatively high pressures with relatively low temperatures (as recorded in fluid inclusions) and is the opposite of conditions in the extensional environment of the late Tertiary. Hydrothermal circulation might then be initiated by either intrusive events, regional hydrologic pressure differentials, or some combination of the two. Deposition occurred where there were breaks in the overpressured regime allowing mixing of ore fluids with overlying fluids. Although this model emphasizes conditions extant during the Cretaceous in the Great Basin, it is possible (perhaps even likely?) that similar conditions existed at other times in the history of the Great Basin and other parts of the world. SHDG deposits from locations outside western North America are beginning to be discovered and should shed new light on the tectonic environment in which these deposits form.

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