THE MINERALOGY OF THE BELLEROPHON-NELSON TELLURIDE-BEARING GOLD DEPOSIT, ST. IVES CAMP, YILGARN CRATON, WESTERN AUSTRALIA

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ABSTRACT

The newly discovered Bellerophon-Nelson telluride-bearing gold deposit at the St. Ives camp, Western Australia, is hosted by meta-sedimentary rocks of the lower Black Flag Group and alkaline intrusions. Four stages of mineralization are recognized; from oldest to youngest these are: quartz-carbonate veins (Stage I), quartz-albite-carbonate-pyrite veins and sericite-pyrite seams (Stage II), quartz-pyrite veins (Stage III), and carbonate ± chlorite veins (Stage IV). Stages II and III contain economic gold mineralization, and the gold grains are strongly associated with pyrite. Intense albite and hematite alteration surround the mineralized veins, and trace amounts of gold precipitated in these altered rocks. The albite and hematite alterations are synchronous and derived from the same oxidized fluid as the auriferous veins. The occurrence and absence of hematite within the alteration zone reflects variable amounts of magnetite in the precursor rocks. Thirteen species of telluride and sulfosalt minerals have been identified in Stages II and III. The most common telluride minerals include calaverite, petzite, tellurobismuthite, and altaite, and these minerals have similar occurrences to native gold. In addition to native gold, telluride and sulfosalt minerals are also major Au carriers and account for at least 15% of the gold in this deposit. The mineral associations of PbCl(OH)–Pb2Cl3(OH)–Te–TeO2 and BiOCl–BiO(OH,Cl)–Te–TeO2 were formed as replacement of earlier telluride minerals as the result of reactions with Cl-bearing fluids. The intergrowth between native gold and Cl-bearing minerals + native Te/Te-oxide indicates that Au in telluride minerals was remobilized and re-deposited. The phase diagram for the telluride and sulfosalt mineral association suggests that during Stage II log ƒS2 decreased from −8 to −11, and that log ƒTe2 increased from −8 to the level required for the formation of Te-oxide. The values of log ƒTe2 and log ƒS2 in the Stage III veins were −8 to −11 and −9 to −11.5, respectively. The highly oxidized, tellurium-enriched hydrothermal fluid, which formed the Bellerophon gold telluride deposit, is consistent with the involvement of magmatic fluid, and sulfidation is the likely cause of gold precipitation.

Keywords: Bellerophon-Nelson gold deposit, St Ives, Yilgarn craton, telluride minerals, Archaean

INTRODUCTION

Archaean greenstone terranes host numerous gold deposits, e.g., the Yilgarn craton in Western Australia and the Superior craton in Canada, with >9200 tons and >8500 tons gold resources respectively (Robert et al. 2005). The most widely accepted hypothesis for the origin of the Archaean greenstone gold deposits in the Yilgarn craton is the orogenic gold model of Groves et al. (1998). However, this hypothesis cannot explain the well-known spatial associations between Archaean gold deposits and felsic alkaline intrusions (Witt 1992, Robert 2001). Two endmember hypotheses exist to explain the gold-intrusion link: (1) the auriferous fluid was derived from a magmatic intrusion (Robert 2001, Mueller et al. 2008, Stoltze 2006, Qui & McNaughton 1999, Walshe et al. 2008) and (2) the competency contrast between the intrusion and wall rocks produced a structural trap for younger mineralization (Duuring et al. 2000, Brown et al. 2002, Witt 1992). Because of the identical ages and Sr/Nd isotope compositions between the gold mineralization and the alkaline intrusions, the Granny Smith-Wallaby gold deposits have been interpreted as having a magmatic origin (Mueller et al. 2008, Stoltze 2006). In contrast, an age difference between the gold mineralization and adjacent intrusions has been reported for the Kanowna Belle gold deposit (Ross et al. 2004). The gold mineralization at the Golden Mile deposit in Western Australia has a different age from the mineralized dolerite, but identical dating results with the lamprophyre dikes (Heath 2003, Rasmussen et al. 2009, McNaughton et al. 2005). Various approaches have been applied to study this question, but no consensus has been reached, e.g., fluid inclusions (Ho et al. 1992), U-Pb dating (Duuring et al. 2004, Mueller et al.

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Gold has been mined from the St. Ives gold camp in the Eastern Goldfields Superterrane (EGST) of the Yilgarn Craton, Western Australia for over 100 years, and it has become the second largest gold producer in Australia (pers. commun. with Julian Woodcock, St. Ives Exploration Manager). Multiple deposits have been mined, the most important being Athena-Hamlet, Victory-Defiance, and Revenge. However, the Bellerophon-Nelson prospect (hereafter shortened to Bellerophon), about 60 km south-southeast of Kalgoorlie, is of particular interest. The host rocks are felsic meta-sedimentary rocks, in contrast to the other deposits of the St. Ives camp, which are hosted in basalt, komatiite, black shale, or dolerite. There is a close association with alkaline felsic intrusions, which are also mineralized (e.g., Granny Smith and Wallaby gold deposits, Mueller et al. 2008). This study describes the geology, alteration, and mineralogy of the Bellerophon prospect. Thirteen telluride and sulfosalt minerals have been found in this deposit, some of which, e.g., calaverite, petzite, buckhornite, are important Au carriers. These telluride and sulfosalt minerals will be described in detail, and the relationship between the gold mineralization and felsic intrusions will be discussed.

**GEOLOGICAL SETTING**

The Yilgarn craton, formed between >3.0 Ga and 2.6 Ga, has been divided into six terranes (Cassidy et al. 2006): Narryer, South West, central Youanmi, Kalgoorlie, Kurnalpi, and Burtville. The latter three constitute the Eastern Goldfields Superterrane (EGST). Among these terranes, the EGST is the most productive gold producer and hosts 15 of the 19 known world-class gold deposits in the Yilgarn (Cassidy & Hagemann 2001).

The greenstone rocks in the EGST are composed of the older Kambalda sequence (2720–2690 Ma, Swager 1997, Krapež et al. 2000), which is unconformably overlain by the Kalgoorlie sequence (2690–2660 Ma, Krapež et al. 2000, Squire et al. 2010). The Kambalda sequence consists of mafic-ultramafic volcanic rocks, locally over 5000 m thick, interbedded with minor black shale horizons. The Black Flag Group and the younger Merougil Group, which together form the Kalgoorlie sequence, consist of meta-sandstone and felsic volcaniclastic rocks, and in the case of the Merougil Group, thick layers of conglomerate. These sequences were intruded by numerous mafic and intermediate to felsic intrusions (Champion & Sheraton 1997, Nelson 1997). The mafic dolerite intrusions, e.g., Condenser, Junction, and Defiance dolerites, have been dated at 2680 ± 8 Ma (Bateman et al. 2001), whereas the intermediate to felsic intrusions have ages of 2680 to 2600 Ma (Nelson 1997, Krapež et al. 2000, Compton et al. 1986, Nguyen 1997). Lamprophyre intrusions in the Kalgoorlie and Kambalda areas have been dated to around 2.63 Ga, which is coeval with the gold mineralization (McNaughton et al. 2005, Perring & McNaughton 1992). The EGST is cut by crustal-scale NNW-trending faults. Blewett et al. (2010) recognized seven stages of deformation. Gold mineralization mainly occurred at the third extensional stage and the fourth and fifth stages, which involved shearing and thrusting (Czarnota et al. 2008, Blewett et al. 2010).

**METHODOLOGY**

Polished thin sections with a thickness of 30 μm were made for those samples of interest at the Australian National University. Detailed microscope observations elucidated the mineral relations and mineralization processes. Compositions of minerals of interest, dominantly sulfide telluride and sulfosalt minerals (samples were coated with ~20 nm of carbon), were determined using a JEOL 6400 SEM (Electron Microscopy Unit, now Center of Advanced Microscopy, ANU) with EDS operating at 15 kV and a focused beam (1 μm) of 1 nA, 120 s counting time, and a 17–20 mm working distance. Due to the high resolution achieved, mixed analyses can be avoided, which is important for analyzing fine-grained mineral phases. Back-scattered images were also used to investigate the phase relations.

**REGIONAL METAMORPHISM**

The metamorphic grade of host rocks to the Bellerophon gold prospect is upper greenschist. Common metamorphic minerals include quartz, carbonate, biotite, chlorite, sericite, and epidote. The metamorphic mineral assemblage reflects the mineralogy of the precursor rock; some samples are biotite-rich, whereas others are rich in quartz and plagioclase. Chlorite may replace epidote and biotite in response to retrograde metamorphic reactions. Magnetite and/or rutile can be found intergrown with chlorite following the breakdown of biotite, and plagioclase has been altered to sericite and calcite. Metamorphic rocks close to mineralization show foliation defined by chlorite, biotite, and sericite. Small veinlets occur parallel to the foliation. The mineral assemblage in the veinlets is similar to the host metamorphic rocks, suggesting that the veinlets formed during or after the regional metamorphism. Opaque minerals in these veins include pyrite + magnetite + chalcopyrite + pyrrhotite. Magnetite close to the mineralization can be oxidized to hematite, but this reaction is not seen elsewhere. As documented in the following text, the mineralization-related hydrothermal alteration overprinted the metamorphic minerals; hence, the mineralization should be later than the regional metamorphism. The dating
results of muscovite in the mineralized veins support this interpretation (Xue 2014). In regional geology, five phases of metamorphism have been recognized (Goscombe et al. 2007), with the dominant metamorphism in EGST being the combination of M2 and M3a. The M2 metamorphism is coeval with the emplacement of regional high-Ca granite and the East–West compression deformation. The M3a metamorphism is characterized by the extensional settings and regional rifting. It is worth noting that lamprophyre dikes younger than the regional metamorphism have been recognized in the Kalgoorlie and Kambalda areas (McNaughton et al. 1992, Rock et al. 1990).

**DEPOSIT GEOLOGY**

The host rocks of the Bellerophon gold prospect are mainly sandstone and siltstone, interlayered with chert and volcaniclastic rocks (Fig. 1). Monzonite and albite felsic bodies, which intrude the supracrustal rocks, are also mineralized (Fig. 1b) and can contain gold in concentrations >20 ppm. The mineralization at Bellerophon has been divided into four stages, based on a combination of field relationships, microscopic textures, and SEM studies (Fig. 2), which from oldest to youngest are: Stage I, quartz-carbonate veins; Stage II, quartz-albite-carbonate-pyrite veins and sericite-pyrite seams; Stage III, quartz-pyrite veins; and Stage IV, carbonate ± chloride veins. Stage II and Stage III account for most of the native gold and telluride mineralization.

Three alteration zones have been observed at Bellerophon (Fig. 1b). Most distal is the chlorite alteration zone, which also contains small amount of sericite. The mineral assemblage in this zone is attributed to the combined effects of alteration by auriferous fluid and regional metamorphism. Sericite abundances in the intermediate zone increase with proximity to mineralization. This zone also contains variable amounts of hematite and albite in association with hydrothermal carbonate and quartz. There are two types of alteration in the proximal zone, intense hematite alteration and albite alteration (Fig. 3). Both can envelop the mineralized veins. In some cases, the two types of alteration combine to form albite+hematite alteration, which has been found in both the sandstone and volcaniclastic wall rocks (Fig. 3). Monzonite intrusions show intense albite alteration, whereas the albite dikes are hematite altered (Fig. 3). Proximal albite and hematite altered rocks can carry native gold, and locally, the sericite alteration is also weakly mineralized (Fig. 1b).

**Stage I**

The principal minerals in the Stage I veins are quartz and carbonate (Fig. 4a), which are locally accompanied by rare apatite and pyrite. The veins are not common and mostly occur at the contact between dark green regional metamorphic rocks and pale to brick-red mineralized zones. The veins are generally only a few tens of centimeters in length, and trend sub-parallel to the regional foliation. Native gold has not been observed in Stage I veins. Thin, moderately weak alteration zones, consisting of small anhedral grains of carbonate, quartz, sericite, and rare pyrite, developed around the Stage I veins.

**Stage II**

Stage II veins (Figs. 5a–c) consist of quartz, albite, carbonate, and pyrite and are characterized by intense proximal alteration. Carbonate minerals are mainly ankerite, with traces of Fe-dolomite, and rare calcite. The Stage II veins have been found cutting the Stage I quartz-carbonate veins (Fig. 4a). Their width varies from several millimeters to centimeters, and they do not extend over a couple hundred meters. The carbonate minerals tend to crystallize at the edge of the veins (Figs. 5a). Common accessory minerals include rutile, muscovite, apatite, and anhydrite. Anhedral chloride was found in the interstitial spaces between quartz, albite, and pyrite grains, and probably crystallized from an evolved hydrothermal fluid. The ~8% sulfide minerals in the Stage II veins are mainly pyrite, with trace chalcopyrite, galena, rare sphalerite, and molybdenite. The pyrite is cubic in shape and can be up to 1 cm in size. Chalcopyrite is anhedral and fills the interstices between the other minerals. Galena is mostly found enclosed by pyrite, whereas sphalerite and molybdenite have crystallized along the edges of pyrite crystals. Sulfate minerals, such as baryte, celestite, and anhydrite, either co-existed with or were enclosed by pyrite. Brecciated wall rock fragments, up to ~3 centimeters in size, are found in larger veins and locally can constitute 15% of the vein. Irregular fine veins (~1.5 cm wide) cut through the wall rocks to form stockwork textures (Fig. 5d).

Both hematite and albite alteration are found adjacent to the Stage II veins (Figs. 5a–c), which changes the color of the wallrocks from dark green (the regional metamorphic rock color) to pale, brick red, or yellow. The alteration halo varies in width from ca. 2 cm in the case of small veins (Fig. 5a) to up to two meters for large veins. Tiny anhedral crystals of albite and quartz, found in the albite alteration zone, are interpreted to have crystallized in response to interaction between the auriferous fluid and the wall rocks. Hematite alteration takes two forms: dusty hematite grains on the surface of feldspar, and the microscopic hematite that formed by oxidation of magnetite. Hematite alteration can also be found in association with biotite+chlorite domains (Fig. 5b). However, the mineralized quartz-albite-carbonate-pyrite veins are
devoid of hematite. The hydrothermal mineral assemblage in the alteration halo is similar to that in the auriferous veins, and includes cubic pyrite, quartz, carbonate, albite, sericite, and apatite.

**Stage III**

Stage III veins are mainly composed of milky white quartz and pyrite (Figs. 5c–5h), but locally they...
can contain small amounts of carbonate and albite. They cut the Stage II veins (Figs. 4b–4c) and can be up to 2 m wide, especially those that are composed of almost pure quartz. Wall rock breccias, with fragments of monzonite and albite dikes, are found in association with Stage III veins (Figs. 5g–5h). Other common accessory minerals in these veins include apatite, rutile, scheelite, and tourmaline. The sulfide minerals (<∼5%) are cubic pyrite and anhedral chalcopyrite. The sulfate minerals, baryte and celestite, are also found in these veins. Carbonate and albite minerals are mostly found at vein edges as a consequence of a reaction between the vein fluid and wall rocks.

The color of the monzonite, where cut by Stage III veins, is changed from red to light gray or yellow in response to albite alteration (Fig. 5g). Albite dikes are strongly hematite altered and show both hematite dusting on albite surfaces and microscopically distinct hematite crystals (Fig. 5h). Both albite and hematite alteration are found in the meta-sedimentary and volcaniclastic rocks surrounding the quartz-pyrite veins. Locally, there are clear reaction zones between the quartz-pyrite-carbonate veins and wall rocks (Fig. 5f) that have resulted in the crystallization of carbonate, albite, quartz, scheelite, muscovite, pyrite, and native gold.

FIG. 2. Paragenesis of the Bellerophon mineral assemblage.

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<th>Mineral assemblage</th>
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Stage IV

There are two types of Stage IV veins: carbonate-chlorite veins and carbonate veins. The former can cut the latter and both are younger than Stage II and Stage III veins (Figs. 4b–c). The Stage IV veins are thin, with widths of no more than 5 mm. Minor minerals, which make up less than 15% of the veins, include quartz, pyrite, muscovite, biotite, magnetite, and tourmaline. The proportion of these minerals varies: some veins have tourmaline and pyrite, whereas others contain muscovite or magnetite (the magnetite locally altered to hematite). Wall rocks for the Stage IV veins are sedimentary and volcaniclastic rocks. There is no alteration halo surrounding the Stage IV veins, other than disseminated chlorite and carbonate minerals in wall rocks close to the veins.

Mineralized sericite-pyrite seams

Another type of mineralization that has been observed at Bellerophon takes the form of native gold and telluride minerals, which occur in sheared sericite-pyrite seams. These seams have a width of several centimeters, are tens of centimeters in length, and are oriented sub-parallel to the foliation in the wall rocks. Mineralized sericite-pyrite seams can be found in the quartz-pyrite veins as stringers (Fig. 5i). Muscovite grains are strongly aligned within the seams. The smaller grains of sericite, with sizes of up to tens of μm, are interpreted to be inherited from the primary fabric of the rocks, which indicate their metamorphic origin. In contrast, larger examples, with grain sizes up to 0.5 mm, are thought to have re-crystallized during mineralization. Pyrite is subhedral to anhedral and can be elongated with curvilinear boundaries (Fig. 5j). These features, combined with the pressure shadow textures between pyrite and fringed biotite + sericite + quartz, were produced by ductile deformation of the shear zones. Besides sericite and pyrite, other trace minerals in the mineralized sericite-pyrite seams include biotite, carbonate, rutile, locally chlorite, and small grains of albite. Elongated pyrite crystals enclose various minerals, including carbonate, sericite, native gold, tellurides, chalcopyrite, and pyrrhotite. The carbonate and sericite grains are usually found in the rim zone of the pyrite, and have the same microscope features as these minerals in the re-crystallized carbonate-sericite seams. These textures indicate that the pyrite was dissolved and re-crystallized during mineralization. Grains of pyrrhotite are not found in the auriferous veins and are therefore interpreted to be inherited from the precursor rocks.
The broken Stage I pyrite crystal seen in Figure 4d shows that the pyrite-sericite seams that fill the pyrite fractures are later than the Stage I veins. The pyrite-sericite seams and quartz-albite-carbonate-pyrite veins have similar telluride mineralogy, which suggests that they crystallized under conditions of similar $fS_2$/$Te_2$ as discussed below, and indicate that the pyrite-sericite seams belong to Stage II. Furthermore, native gold grains have been found in breccias in the Stage II veins, indicating brittle deformation, whereas the elongated pyrite crystals in the sericite-pyrite seams are interpreted to be the product of ductile deformation. Therefore, ductile and brittle deformation occurred over the same time interval, and both were coeval with gold mineralization.

**Systematic changes of the magnetic susceptibility**

Alteration haloes around the Bellerophon veins show a systematic variation in magnetic susceptibility. An example is illustrated in Figure 6a, which shows a ca. 15 cm-wide hematite alteration halo adjacent to an auriferous vein. Magnetic susceptibility gradually increases from <1.0 (the dimensionless ratio of the magnetic moment per unit volume to the magnetic field strength) in the hematite alteration to up to >10.0 in precursor rocks. The hematite-altered rock shown in Figure 6b has magnetic susceptibility values as high as 22.6, whereas it is only 0.6 in the adjacent chlorite alteration zone. The magnetic susceptibility for hematite in altered rocks can vary from ~1.0 to ~50; for precursor rocks, the values can be between 10 and >100. The albite dikes rocks are strongly hematite altered (Fig. 5h), and magnetic susceptibility values generally lie between 15 and ~30.

The hematite- and albite-altered samples in Figure 6c are in close proximity to each other and associated with similar Stage II mineralized veins. However, the magnetic susceptibility value of the hematite-altered rock is 25 times higher than that of the albite-altered rock. Generally, the magnetic susceptibility value in albite-altered rocks is <1, and the values for the precursor rocks to albite alteration are in the same range as that of albite alteration. For instance, the magnetic susceptibility values for both fresh and albite-altered monzonite rocks are ~0.5. Furthermore, the magnetic susceptibility values of albite+hematite-altered rocks generally have large overlaps with those of both the hematite alteration and albite alteration.
NATIVE GOLD, TELLURIDE, SULFOSALT AND 
CL-BEARING MINERALS

Thirteen telluride sulfosalt and chlorine-bearing minerals have been identified in the mineralized Stage II 
and III rocks, of which calaverite, petzite, tellurobis-
muthite, and altaite are the most common. Other miner-
als include melonite, telluroantimony, rucklidgeite, 
tetradymite, buckhornite, aikinite, and native tellurium. 
The Cl-bearing minerals are associated with native tel-
lurium or Te-oxide, and consist of Bi-Cl and Pb-Cl 
oxide or hydroxide minerals.

Native gold

Native gold has three different types of occurrence 
at Bellerophon. The first is found in mineralized 
veins, i.e., quartz-albite-carbonate-pyrite veins from 
Stage II and quartz-pyrite veins from Stage III. The 
second is in the sericite-pyrite seams. Finally, small 
grains of gold are locally enclosed by pyrite in inten-
sely altered wall rocks. Representative compositions 
of native gold are listed in Table 1.

Native gold is either intergrown with pyrite or 
quartz in the quartz-albite-carbonate-pyrite veins, or 
enclosed by pyrite and rarely by albite (Fig. 7a). 
Alternatively, it may fill fractures in pyrite (Fig. 8a). 
The size of the gold grains varies from several μm up 
to 100 μm. Some grains of pyrite have dusty textures 
with abundant inclusions that consist mainly of rutile 
with trace amounts of gold + calaverite (Fig. 7a). The 
finesse of native gold in the quartz-albite-carbonate-
pyrite veins is always higher than 900 (Table 1). In 
the quartz-pyrite veins, native gold is either enclosed 
by pyrite or disseminated in the veins with telluride 
minerals (Fig. 7b, 7f). The size of these gold grains is 
20–50 μm, and the finesse is highly variable, from as 
high as 960 down to as low as 800 (Table 1). Gold 
grains enclosed by pyrite have higher values of finesse 
that the free gold in the quartz-pyrite veins. Gold 
grains in the pyrite-sericite seams are enclosed by 
pyrite (Fig. 7c and 7d) or fill extension fractures in 
pyrite (Fig. 8b). Some gold grains form films that 
crystallized along the edges of pyrite grains in associ-
ation with sericite. Native gold in the sericite-pyrite 
seams is generally a few μm in size, and the finesse is 
mainly in the range of 900 ~ 950.

Fig. 5. (a) Native gold and telluride minerals in a quartz-albite-carbonate-pyrite vein with a hematite alteration halo. (b) A 
quartz-albite-carbonate-pyrite vein mantled by albite and hematite alteration. Note that the hematite alteration is associated 
with biotite-chlorite domains. (c) A quartz-albite-carbonate-pyrite vein with the albite+hematite alteration halo. (d) Quartz-
albite-carbonate-breccia texture in a Stage II vein. (e) A quartz-pyrite vein with albite+hematite and hematite alteration 
halos. Abundant gold and telluride minerals are found in this vein. (f) Quartz veins cutting the primary foliation. Native 
gold, albite, pyrite, muscovite, and scheelite have precipitated in the dark reaction zone. (g) A quartz-pyrite vein cutting a 
monzonite intrusion and forming a stockwork texture. The monzonite shows intense albite alteration, and gold grades can 
be as high as 16 ppm. (h) Stockwork texture in a quartz-pyrite vein within an albite dyke rock. (i) Stringers of sericite-
pyrite seams in a late white quartz vein. The stringer contains native gold and telluride minerals. (j) Curvilinear bound-
aries of pyrite in sericite-pyrite seams. Sericite inclusions and pressure shadows are found in association with these pyrite 
grains.
Calaverite was deposited in the Stage II and III veins, but was not found in the sericite-pyrite seams. It is bright and light yellow under reflected light and its shape varies from subhedral to anhedral. The size of the crystals is usually small, typically <20 μm. Calaverite is typically associated with albite and carbonate in the quartz-albite-carbonate-pyrite veins. Inclusions of calaverite have been found in pyrite (Fig. 7a) and rarely in albite (Fig. 7e) where they have curvilinear boundaries (Fig. 7e). In the quartz-pyrite veins, it is disseminated and associated with other precious minerals (Fig. 7b). Unlike native gold, the surface of calaverite is highly porous (Fig. 7b), which may explain why the totals for the SEM analyses are generally near 97%. The composition of calaverite is given in Table 1, and its formula is approximately AuTe(1.97−2.08). Apart from 0.4 to 0.9% Ag, no other trace element was detected.

Petzite has been found in both the quartz-pyrite veins and sericite-pyrite seams, but not in the quartz-albite-carbonate-pyrite veins. Two types occur in the quartz-pyrite veins based on size: the larger grains (near 20 μm) are usually subhedral, whereas smaller petzite crystals have a grain size of only a couple of microns (Fig. 7f). Furthermore, the small grains always form trails that fill fractures in quartz and are connected with larger grains of petzite (Fig. 7f). Petzite in the sericite-pyrite seams can be either

---

**TABLE 1. REPRESENTATIVE COMPOSITIONS OF NATIVE GOLD, CALAVERITE, AND PETZITE**

<table>
<thead>
<tr>
<th>Type</th>
<th>Vein 1</th>
<th>Vein 2</th>
<th>Vein 1</th>
<th>Vein 2</th>
<th>Vein 1</th>
<th>Vein 2</th>
<th>Vein 1</th>
<th>Vein 2</th>
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</table>

Vein 1 refers to the quartz-albite-carbonate-pyrite vein. Vein 2 refers to the quartz-pyrite vein. Ser-Py is the sericite-pyrite seams. They have the same meaning in the following tables.

Fig. 7. (a) Zoned pyrite crystal in a quartz-albite-carbonate-pyrite vein, with an inner zone of dusty pyrite that contains inclusions of rutile (Rt) + native gold (Au) + calaverite (Clv), and an outer zone of pyrite. (b) Mineral association of native gold, calaverite, petzite (Ptz), and tellurobismuthite (Tb) in a quartz-pyrite vein. Note the porous texture on the surface of calaverite, petzite, and tellurobismuthite grains. (c) A mineral assemblage of native gold, galena (Gn), tellurobismuthite, buckhornite (Buh), and chalcopyrite (Cp) enclosed by pyrite in a sericite-pyrite seam. (d) An association of native gold, petzite, melonite (Mlt), and rucklidgeite (Ruc) enclosed by pyrite in a sericite-pyrite seam; altaite (Alt) can fill fractures in pyrite. Note that melonite occurs in the middle of petzite. (e) Rounded grains of calaverite with curvilinear boundaries enclosed by albite in a quartz-albite-carbonate-pyrite vein. (f) Native gold and petzite in a quartz-pyrite vein. Note the trails of small petzite crystals associated with larger petzite grains. (g) Galena and aikinite (Aik) enclosed by pyrite in a quartz-albite-carbonate-pyrite vein. The relations between galena and aikinite indicate exsolution or an immiscibility reaction. (h) Buckhornite associated with chalcopyrite and galena in disseminated pyrite from an albite altered rock next to a mineralized vein.
FIG. 8. (a) Mineral assemblage consisting of Bi-Cl-Te, telluriobismuthite (Tb), and native gold filling fractures in pyrite in a sheared sericite-pyrite seam. Note that the Bi-Cl-Te area is a mixture of native Te and Bi-Cl minerals, which replaces telluriobismuthite (Tb). The microphotograph also shows native gold associated with Bi-Cl-Te. (b) Native gold and altaite (Alt) filling a fracture in pyrite from a pyrite-carbonate-quartz-albite vein. The area of Pb-Cl-Te, which replace altaite, is a mixture between native tellurium and Pb-Cl minerals. (c) An altaite and Pb-Cl-Te mixture filling a fracture in pyrite from a quartz-carbonate-albite-pyrite vein; the altaite was replaced by Pb-Cl-Te. (d) An assemblage of native gold, Pb-Cl minerals, Pb-Cl-Te mixture, and altaite filling fractures in pyrite in a quartz-albite-carbonate-pyrite vein. Note that the Pb-Cl mineral has precipitated next to native gold, and that the Pb-Cl-Te minerals occur between Pb-Cl and altaite. (e) Element spectra diagram of Bi-Cl-Te minerals from a sericite-pyrite seam. (f) Element spectra for the Pb-Cl-Te minerals from a quartz-albite-carbonate-pyrite vein.
intergrown with sericite or enclosed by pyrite in association with native gold (Fig. 7d). The surface of petzite is porous (Fig. 7b) and, as a consequence, the SEM analyses totals fall below 100%. It has low Ag but high Te (Table 1) with the approximate formula \( \text{Au}_{0.98-1.09}\text{Ag}_{2.87-2.95}\text{Te}_{2.03-2.07} \).

**Altaite**

Altaite is found in both the quartz-albite-pyrite-carbonate veins and sericite-pyrite seams. In the former, altaite fills fractures in pyrite and must therefore be later than pyrite and albite (Fig. 8a, c, d). Altaite in the sericite-pyrite seams fills fractures in pyrite together with petzite and native gold (Fig. 7d). Euhedral laths of altaite were also found in sheared seams intergrown with sericite, albite, and native gold. The atomic percentage of Pb is normally 1 to 5% less than Te. Bismuth can substitute Pb in altaite and the atomic percentage can be as high as 11% (Table 2).

**Tellurobismuthite**

Tellurobismuthite (\( \text{Bi}_2\text{Te}_3 \)) is the only Bi-telluride mineral found in the Bellerophon deposit. Subhedral-euhedral grains, with a size of around 20 \( \mu \text{m} \), are found associated with native gold, carbonate, and albite in the quartz-albite-carbonate-pyrite veins. It can also fill fractures in pyrite in association with altaite. Tellurobismuthite in quartz-pyrite veins is normally associated with gold and calaverite (Fig. 7b). It is also found intergrown with chalcopyrite as inclusions in quartz in quartz-pyrite veins. Tellurobismuthite in sericite-pyrite seams is either enclosed by pyrite (Fig. 7c) or fills fractures in pyrite. The analyzed tellurobismuthite has the formula \( \text{Bi}_{1.94-1.97}\text{Te}_{3.01-3.06} \) (Table 2).

Two other telluride minerals found in this study were melonite (\( \text{NiTe}_2 \)) and tellurantimony (\( \text{Sb}_2\text{Te}_3 \)). Melonite was found in both quartz-albite-carbonate-pyrite veins and sericite-pyrite seams (Fig. 7d). Melonite in Bellerophon contains Fe, with the composition of \( (\text{Ni}_{0.9},\text{Fe}_{0.1})\text{Te}_2 \). Tellurantimony was only found in the sericite-pyrite seams and it is enriched in tellurium relative to antimony compared with the stoichiometric formula.

**Sulfosalts**

The sulfosalts minerals found at Bellerophon include rucklidgeite (\( \text{Pb}_3\text{Bi}_2\text{Te}_5 \)), buckhornite \( [\text{Au}_1(\text{Pb}_2\text{Bi})_3] \text{Te}_2\text{S}_1 \), aikinite \( [(\text{Cu},\text{Pb},\text{Bi})_8]\text{S}_1 \), and tetradyrite \( (\text{Bi}_2\text{Te}_2\text{S}_1) \); the compositions are listed in Tables 2 and 3. Tetradyrite and buckhornite are the most common. Only one grain of rucklidgeite was found in a sericite-pyrite seam, and aikinite was only found in quartz-albite-carbonate-pyrite veins where it was
enclosed by pyrite and associated with galena (Fig. 7g).

Buckhornite inclusions in pyrite were found in quartz-albite-carbonate-pyrite veins, sericite-pyrite sheared seams, and mineralized wall rocks (Fig. 7h). Most buckhornite grains are rounded with curvilinear boundaries, but some are euhedral with rhombic shapes. The size of buckhornite grains can be up to \( \sim 30 \mu m \). Tetradymite, in association with altaite, fills fractures in pyrite from the mineralized veins. Small grains of tetradymite were also found in sericite-pyrite seams. As shown in Table 2, tetradymite from the Bellerophon deposit is enriched in tellurium and depleted in sulfur.

Native Te/Te-oxide and chlorine bearing minerals

Native Te/Te-oxide and Cl-bearing minerals have been found in the Bellerophon deposit (Fig. 8). The CI-bearing minerals are divided into two groups, those that belong to the Pb-CI system of minerals and those that belong to the Bi-CI system. The Pb-CI minerals have only been found in the mineralized quartz-albite-carbonate-pyrite veins (Fig. 8a, c, d), and the Bi-CI minerals in the sericite-pyrite seams (Fig. 8b). Native Te/Te-oxide occurs intergrown with both. Altaite has been found associated with Pb-CI minerals, and tellurobismuthite with Bi-CI minerals. Native gold locally occurs intergrown with both Bi-CI and Pb-CI minerals (Fig. 8). Chlorine-bearing minerals are among the last to crystallize, as they fill fractures in pyrite and replace other telluride minerals (Fig. 8). During the replacement process, invisible Au in telluride minerals was reactivated and precipitated with the Cl-bearing minerals in pyrite fractures. Tellurium/Te-oxide and Cl-bearing minerals are always mixed together, and cannot be distinguished under the microscope (Fig. 8a, c). Therefore, they are described together here.

The surface of these Bi-CI-Te and Pb-CI-Te minerals is porous, as shown in Figure 8, and the compositions are listed in Table 4. The atomic percentage of Te in the Pb-CI-Te system minerals lies within the range 19.5–84.0%, whereas the values for Pb and Cl lie between 4.6–36.3% and 10.0–41.4% respectively, which indicates a mixture between native Te/Te-oxide and Pb-CI. This is the same as for the Bi-CI-Te system minerals. Element spectra diagrams for Pb-CI and Bi-CI minerals show obvious oxygen peaks that can be higher than the Cl and Bi peaks (Fig. 8e, 8f), which could be the reason for the low analytical totals in Table 4. The Cl-bearing minerals are interpreted to be of the form PbCl(OH)\(_x\) and BiCl(OH)\(_x\). As shown in Table 4, the atom ratios of Pb/Cl and Bi/Cl range from 0.85 to 1.02 and 0.94 to 1.18, respectively. Therefore, the PbCl(OH)\(_x\) mineral can be a mixture between laurionite/paralaurionite [PbCl(OH)] and penfieldite [Pb\(_2\)Cl\(_3\)(OH)] and the BiCl(OH)\(_x\) mineral should be a mixture between bismoclite (BiOCl) and daubreeite [BiO(OH,Cl)]. After calculated (OH) is added to the SEM analytical total, the final results are still below 100\%, which suggests that there should be tellurite (TeO\(_2\)) instead of native Te. However, if all of the Te was present as TeO\(_2\) the totals would be over 100\%. We conclude that Pb-CI-Te-(OH) system minerals are mixtures of PbCl(OH),
Pb₂Cl₃(OH), native Te, and TeO₂; and the Bi-Cl-Te-(OH) minerals are mixtures of BiOCl, BiO(OH,Cl), native Te, and TeO₂.

**DISCUSSION**

**Hematite and albite alteration**

The two most prominent types of alteration found adjacent to the mineralized veins at Bellerophon are hematite and albite alteration, and both types can be mineralized. Hematite alteration is found in the albite dike rocks, and intense albite alteration has also been found in monzonite intrusions. Intense hematite and albite alteration occurs in both meta-sedimentary and volcaniclastic rocks. Both the nature of the hydrothermal fluid and precursor rocks affect the type of alteration. A study of the temporal and spatial relations between alteration and mineralization, and causes of the observed differences, is essential to understand the mineralization process.

We interpret the hematite and albite alterations to be coeval and derived from the same hydrothermal fluid. The evidence to support this interpretation can be summarized as follows: (1) quartz-albite-carbonate-pyrite veins can be associated with both intense albite alteration and hematite alteration in the meta-sedimentary rocks (Fig. 5a, 5b); (2) quartz-pyrite veins can be associated with albite alteration in monzonite intrusions and hematite alteration in albite dike rocks; (3) intense albite + hematite alteration, which has features of both the albite alteration and hematite alteration, has been observed; (4) hematite alteration implies that the hydrothermal fluid was highly oxidized (Neumayr et al. 2008). The sulfate minerals, baryte and celestite, have been found in the veins which are associated with albite alteration. Therefore, both types of alteration were caused by highly oxidized fluids, which suggests that the differences in alterations are due to differences in the precursor rocks, rather than differences in fluid composition. As described above (Fig. 6), the magnetic susceptibility values of hematite altered rock and their precursor rocks lie within the range 1.0 to 50 and 10 to >100, respectively. By comparison, the albite altered rocks and their precursors have values <1. Combining with the microscope observations that magnetite has been altered to hematite, the hematite in the alteration zones resulted from the oxidation of magnetite. Magnetite has been found in regional metamorphic veins, which suggests that it was probably an original mineral in the precursor rocks, or formed during regional metamorphism before gold mineralization. The sample shown in Figure 5b has both hematite and albite alterations surrounding the same vein. The hematite and albite alterations most likely occurred at the same time and, in both cases, alteration was produced by the same highly oxidized auriferous fluid. The difference between the hematite and albite alteration arises from variations in the magnetite content of the precursor rocks.

**Comparison with other deposits in the St. Ives camp**

The major gold deposits in the St. Ives camp include Victory-Defiance, Revenge, Argo, Intrepid, Athena-Hamlet, and Junction. Numerous detailed studies have been conducted on these deposits (Clark et al. 1986, 1989, Neumayr et al. 2008, Nguyen 1997), and a comparative study between these deposits with regards to host rocks, structure, and alteration was carried out by Watchorn (1998), but it does not include the more recently discovered Athena and Bellerophon deposits. Bellerophon is unique in terms of its wall rock alteration and ore mineral association.

### TABLE 4. REPRESENTATIVE COMPOSITIONS OF Bi-Te-Cl SYSTEM AND Pb-Te-Cl SYSTEM MINERALS

<table>
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</table>

Pb₅Cl₃(OH), native Te, and TeO₂; and the Bi-Cl-Te-(OH) minerals are mixtures of BiOCl, BiO(OH,Cl), native Te, and TeO₂.

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A wide range of rocks types can host gold deposits at St. Ives, but most of the economic mineralization occurs in mafic-ultramafic rocks. Felsic intrusions are the second most common lithologies to host mineralization.
e.g., Intrepid, Bellerophon, Flame porphyry from Victory, Bellesile, Beta porphyry. Furthermore, a significant fraction of the Bellerophon mineralization is hosted by the meta-sedimentary and felsic volcaniclastic rocks of the Black Flag Group, which was the first such discovery in St. Ives camp. Subsequently, other prospects have been found within the Black Flag Group and Merouqil Creek Beds sedimentary sequences.

Two zones of wall rock alteration can be recognized in the St. Ives camp; an outer zone of chlorite alteration with biotite contents gradually increasing towards the center of the mineralized lodes, and an inner bleached alteration zone adjacent to the quartz veins, with quartz, albite, carbonate, and pyrite forming the lode itself (e.g., Clark et al. 1989, 1986, Watchorn 1998, Neumayr et al. 2008). In contrast, Bellerophon is characterized by strong albite-hematite and muscovite alteration zones, with a distinct absence of biotite alteration. Sulfide minerals at Bellerophon consist mainly of pyrite with trace amounts of chalcopyrite + galena, which is the same as for the other deposits in the St. Ives camp, except that some deposits contain pyrrhotite and arsenopyrite (Neumayr et al. 2008). Two types of magnetite occurred at Bellerophon: (1) the pre-gold mineralization magnetite, which can be oxidized into hematite during the hydrothermal alteration; and (2) hydrothermal magnetite, which was precipitated at the later stage of the gold mineralization and is intergrown with chlorite-quartz in the Stage IV veins. In comparison with other gold deposits in this region, hydrothermal magnetite minerals mostly predate or are coeval with the early stage of gold mineralization. In addition, sedimentary magnetite has been found in the sedimentary layers, e.g., the Kapai Slate. The mineralization styles of the Bellerophon deposits are similar to the other St. Ives deposits, and consist of quartz veins, quartz vein stock-work, breccias, and shear zones.

Neumayr et al. (2008) has documented that both reduced and oxidized assemblages occur in the St. Ives camp. Pyrrhotite-pyrite occurrences have been discovered in the Argo and Junction deposits, and hematite-pyrite minerals have been found in the Revenge deposit. Magnetite in the Kapai Slate has been observed to be oxidized into hematite during the hydrothermal alteration in the Victory gold deposit (Xu 1999). The wide range of δ34S values for pyrite in the St. Ives camp also indicates variable redox conditions during mineralization. Accordingly, the mixing of two fluids with contrasting redox state has been proposed as the model for gold precipitation in the St. Ives camp (Neumayr et al. 2008). As stated above, all the mineral assemblages, e.g., the sulfate minerals and hematite, in Bellerophon represent oxidized conditions. The oxidation of magnetite into hematite during mineralization has been observed as well. Therefore, the fluid mixing theory cannot be applied to explain the mineralization in Bellerophon.

The most distinctive feature of the Bellerophon prospect is the abundance of telluride minerals, which account for about 15% of the Au. Telluride minerals have been reported at the Victory deposit by Clark et al. (1989), but in trace amounts. Blewett et al. (2008) and Clout (1989) argued that the gold mineralization that was coeval with D5 dextral deformation can contain significant amounts of base metals and tellurides. The low calcium granites in EGST crystallized within the same time period as the D5 deformation (2655 ~ 2630 Ma, Blewett et al. 2008), which is similar to the age of formation of the gold deposits at the St. Ives camp (e.g., Miller et al. 2008, Nguyen 1997, Clark et al. 1986). By analogy, it is inferred that the Bellerophon deposit may have formed during the D5 deformation at about 2630 Ma. The monzonite and syenite intrusions at Bellerophon occurred as dikes, which can be cross cut by mineralized veins. Native gold and telluride minerals have been observed in these veins, and the Au grade can be over 20 g/t. Disseminated telluride minerals have also been found in the adjacent alteration halos of monzonite.

The δS2, fTe2 conditions for Au and telluride minerals deposition

The telluride and sulfosalts mineral assemblages can be used to estimate the δS2 and fTe2 of the auroferous hydrothermal fluid (e.g., Afifi et al. 1988a, b, Cabri 1965, Zhang & Spry 1994). Different mineral assemblages correspond to distinct δS2, fTe2 values for the fluid at a given temperature. Based on where the sulfide-sulfate assemblages at Bellerophon plot on an fO2-T diagram (Ohmoto & Goldhaber 1997), the temperature of formation of these phases can be constrained to be around 300 °C. Two sphalerite grains with Fe contents of 1.2–1.3%, coexisting with pyrite in quartz-albite-carbonate-pyrite veins, requires a crystallization temperature of about 300 °C calculated from the pyrite-sphalerite geothermometer. Based on the above discussion we conclude that the temperature of the hydrothermal system was ~300 °C.

Gold and telluride minerals in the quartz-pyrite-carbonate-albite veins can be further divided into three stages: the earliest involves minerals which are enclosed by pyrite or disseminated in the veins, e.g., bismuthtelluride, gold, calaverite, melonite, buckhornite, galena, and aikinite; stage two consists of minerals that fill the fractures in pyrite and include gold, altaite, and bismuthtelluride; the final stage is the assemblage of native tellurium and Pb-Cl minerals that replaced altaite in the pyrite fractures. Intergrowth of gold and calaverite in the first stage can be used to constrain the upper boundary of fTe2. The coexistence of bismuthtelluride, tetradymite, and buckhornite suggests that the fTe2 and δS2 values should lie close to the boundary between Bi2S3 and Bi2Te3, which is consistent with the intergrowth between galena and aikinite in
some samples. As seen from the $f_{S_2}$ versus $f_{Te_2}$ diagram (Fig. 9), the first stage corresponds to a log$Te_2$ of $-8$, and a log$S_2$ of $-8$ to $-9$. The second stage mineral pairs of gold-altaiite and altaite-bismuthtelluride give a log$Te_2$ that is the same as the first stage, but log$S_2$ drops to $-11$. After the second stage, chlorine-bearing fluids have replaced altaite and led to the crystallization of mixtures of native Te/Te-oxide and Pb-Cl minerals, which require a log$Te_2$ of up to $-7.5$.

Native gold, telluride, and sulfosalt minerals are also found in the sheared sericite-pyrite seams, which can be again classified into the three stages recognized in the quartz-albite-carbonate-pyrite veins. The first stage involves native gold, galena, bismuthtelluride, buckhornite, and tetradymite, which are enclosed by pyrite. Altaite, gold, petzite, melonite, and rucklidgeite form the second stage, which fills in the pyrite fractures. In the third stage, chlorine-bearing minerals and native tellurium replaced bismuthtelluride. The association of bismuthtelluride and tetradymite constrains the crystallization conditions to lie close to the boundary between Bi$_2$S$_3$ and Bi$_2$Te$_3$. Co-existing native gold and buckhornite constrain the upper limit of $f_{Te_2}$. The assemblage of altaite, gold, and petzite defines the stability of the second stage. The variations in the log$Te_2$ and log$S_2$ of telluride minerals in the sheared sericite-pyrite veins are the same as those of the quartz-ankerite-pyrite-albite veins (Fig. 9), which suggests that the precipitation of native gold and telluride minerals in both systems was synchronous. Only native gold, calaverite, petzite, and bismuthtelluride were found in the quartz-pyrite veins. No temporal difference has been detected between these minerals. The values of log$Te_2$ and log$S_2$ of the quartz-pyrite veins are constrained to lie between $-8$ and $-11$, and $-9$ and $-11.5$, respectively.

Telluride minerals can also be used to determine the redox state of the hydrothermal fluid. The Bellerophon tellurides, with Bi/(Te + S) < 1, e.g., the tellurobismuthite (Bi$_2$Te$_3$) and tetradymite (Bi$_2$Te$_2$.1S$_1$), require oxidized conditions (Afifi et al. 1988a, b). Furthermore, Au-(Ag)-telluride and native tellurium or Te-oxide minerals also crystallize under oxidized conditions (Zhang & Spry 1994). McPail (1995) suggests that the stability of native Te broadly covers the pyrite field in $f_{S_2}$–$f_{O_2}$ space with a solubility minimum close to the Mt-Hem buffer at 300 °C.

In summary, both the quartz-albite-carbonate-pyrite veins and sericite-pyrite seams require log$S_2$ values to decrease from $-8$ to $-11$, with log$Te_2$ remaining at

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**Fig. 9.** Stability fields for selected telluride and sulfide phases on a plot of $f_{S_2}$ against $f_{Te_2}$ at 300 °C (Afifi et al. 1988a).
<table>
<thead>
<tr>
<th>Deposit</th>
<th>Location</th>
<th>Age</th>
<th>Te and Bi minerals</th>
<th>Mineralization type</th>
<th>Relations with intrusions</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sandaowanzi</td>
<td>The Northern Great Xing'an Range, China</td>
<td>125.3~116.6 Ma</td>
<td>Clv, Knn, Syl, Ptz, Emp, Hes, Stz</td>
<td>Low sulfidation epithermal deposit</td>
<td>Temporally related</td>
<td>Liu et al. (2011)</td>
</tr>
<tr>
<td>Cripple Creek</td>
<td>Colorado, USA</td>
<td>~30 to ~28 Ma</td>
<td>Clv, Knn, Syl, Ptz, etc.</td>
<td>Low sulfidation epithermal deposit</td>
<td>Genetically related</td>
<td>Kelley et al. (1998)</td>
</tr>
<tr>
<td>Panormos Bay</td>
<td>Tinos Island, Cyclades, Greece</td>
<td>~14 Ma</td>
<td>Hes, Alt, Syl, Te, Stz, Cv, Mit, Kos, Knn, Ptz, Clv</td>
<td>Low sulfidation epithermal gold deposit</td>
<td>Genetically related</td>
<td>Spry et al. 2006</td>
</tr>
<tr>
<td>Emperor and Tuvatu deposit</td>
<td>Fiji</td>
<td>Pliocene</td>
<td>Clv, Knn, Syl, Ptz, Stz, Hes, Alt, Clr, Bmt, Te, Me</td>
<td>Low sulfidation epithermal deposit</td>
<td>Genetically related</td>
<td>Scherbarth &amp; Spry (2006); Pals &amp; Spry (2003)</td>
</tr>
<tr>
<td>Sacarimb deposit</td>
<td>Golden Quadrilateral, Romania</td>
<td>Neogene</td>
<td>Ttd, Tb, Buh, Ngy, Syl, Knn, Ptz, Hes, Stz, Te</td>
<td>Low sulfidation epithermal deposit</td>
<td>Genetically related</td>
<td>Cook et al. 2005</td>
</tr>
<tr>
<td>Acupan gold mine</td>
<td>Baguio district, Philippines</td>
<td>0.65 Ma</td>
<td>Hes, Clv, Te, Ptz, Alt, Syl, Clr</td>
<td>Low sulfidation epithermal deposit</td>
<td>Genetically related</td>
<td>Cooke et al. (1996); Cooke &amp; McPhail (2001) Siddleley &amp; Araneda (1986); Jannas et al. (1990)</td>
</tr>
<tr>
<td>El Indio</td>
<td>The high Andes of North-central, Chile</td>
<td>Tertiary</td>
<td>Te, Clv, Syl, Ptz, Hes, Knn</td>
<td>High sulfidation epithermal deposit</td>
<td>Genetically related</td>
<td>Plotinskaya et al. (2006); Kovalenker et al. (2003)</td>
</tr>
<tr>
<td>Kochblak and Kairagach</td>
<td>Kurama Ridge, Middle Tien Shan, Uzbekistan</td>
<td>Late Paleozoic</td>
<td>Clv, Alt, Hes, Ttd; Bi-(S)-selenides and Pb-(S)-selenides in Kairagach; Bi-Te-(S)-selenides and Pb-Te-(S)-selenides in Kochbulak</td>
<td>Epithermal deposit</td>
<td>Genetically related</td>
<td></td>
</tr>
<tr>
<td>Golden sunlight</td>
<td>Whitehall, Montana, USA</td>
<td>Cretaceous</td>
<td>Clv, Ttd, Tb, Clr, Mit, Buh, Ptz, Knn, Syl</td>
<td>Epithermal gold deposit</td>
<td>Genetically related</td>
<td>Spry et al. (1997); Spry et al. (1996)</td>
</tr>
<tr>
<td>Furtei Au deposit</td>
<td>Sardinia, Italy</td>
<td>Tertiary</td>
<td>Hes, Stz, Syl, Ptz, Clr, Alt, Te</td>
<td>Epithermal gold deposit</td>
<td>Genetically related</td>
<td>Fadda et al. (2005)</td>
</tr>
<tr>
<td>Commoner Mine</td>
<td>Mildlands greenstone belt, Zimbabwe</td>
<td>2.7 Ga</td>
<td>Hes, Alt, Ptz, Syl, Mit, Kos, Gr, Stz, Te</td>
<td>Orogenic type deposit</td>
<td>Genetically related</td>
<td>Twemlow (1984)</td>
</tr>
<tr>
<td>Deposit</td>
<td>Location</td>
<td>Age</td>
<td>Te and Bi minerals</td>
<td>Mineralization type</td>
<td>Relations with intrusions</td>
<td>Reference</td>
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<tr>
<td>Macassa gold mine</td>
<td>Kirkland lake camp, Canada</td>
<td>Late Archean</td>
<td>Clv, Ptz, Alt etc.</td>
<td>Orogenic gold deposit</td>
<td>Spatially related</td>
<td>Tesfaye (1992)</td>
</tr>
<tr>
<td>Sunrise Dam</td>
<td>Laverton Tectonic Zone, Western Australia</td>
<td>2654±8 Ma</td>
<td>Ngy, Hes, Td, Alt, Ptz, Ti, Clv, kolarite</td>
<td>Orogenic type deposit</td>
<td>Genetically related</td>
<td>Brown et al. (2002); Sung et al. (2007)</td>
</tr>
<tr>
<td>Golden Mile</td>
<td>Kalgoorlie, Western Australia</td>
<td>2627±3 Ma</td>
<td>Clv, Ptz, Alt, Ti, Hes, Clv, Knn, Clr</td>
<td>Orogenic gold deposit</td>
<td>Spatially related</td>
<td>Shackleton et al. (2003)</td>
</tr>
<tr>
<td>Kutemajärvi deposit</td>
<td>Southwest Finland</td>
<td>Paleoproterozoic</td>
<td>Clv, Tb, Alt, Te, Hes, Ptz, Knn, Clr</td>
<td>Orogenic gold deposit</td>
<td>No</td>
<td>Poutiainen &amp; Grönholm (1996)</td>
</tr>
<tr>
<td>Potashnya</td>
<td>Kocherov zone, Ukrainian Shield</td>
<td>Precambrian</td>
<td>Hes, Alt, Mlt, Tb</td>
<td>Orogenic gold deposit</td>
<td>No</td>
<td>Bondarenko et al. (2005)</td>
</tr>
<tr>
<td>Maldon</td>
<td>Central Victorian gold province, Australia</td>
<td>365±2 to 373 Ma</td>
<td>Bi, Bmt, Mld, Js-B, Js-A, Hdl</td>
<td>Orogenic deposit overprinted by intrusive related gold system</td>
<td>Temporally related</td>
<td>Bierlein et al. (2001); Cobotanu et al. (2010)</td>
</tr>
<tr>
<td>Tissa-Sarkhoi ore cluster</td>
<td>Southeastern East Sayan</td>
<td>Late Riphean-Vendian</td>
<td>Ptz, Clv, Hes, Tb, Alt, Mlt</td>
<td>Porphyry mineralization system</td>
<td>Temporally and spatially related</td>
<td>Damdinov et al. (2007)</td>
</tr>
<tr>
<td>Kassiteres-Sappes area</td>
<td>Western Thrace, Greece</td>
<td>Tertiary</td>
<td>Hes, Ptz, Alt, Stz, Syl, Td, Te, Gl, Tn</td>
<td>Porphyry-epithermal system</td>
<td>Genetically related</td>
<td>Voudouris et al. (2006)</td>
</tr>
<tr>
<td>Western Thrace and Limnos island</td>
<td>Northeastern Greece</td>
<td>Tertiary-Quaternary</td>
<td>Hes, Stz, Syl, Ptz, Te, Al, Te</td>
<td>Porphyry system</td>
<td>Genetically related</td>
<td>Voudouris (2006, 2007)</td>
</tr>
<tr>
<td>Gyas, Safyanovsk, and Uzelginisk deposits</td>
<td>The Tagil-Magnitogorsk trough in the Urals orogenic belt, Russia</td>
<td>Lower Silurian-middle Devonian</td>
<td>Alt, Hes, Clv, Stz, Ptz, Knn, Tl, Aik, Te</td>
<td>VHMS deposits</td>
<td>No</td>
<td>Vikentyev (2006)</td>
</tr>
<tr>
<td>El Valle deposit, and Ortosar deposit</td>
<td>Rio Narsea gold belt, Asturias, Spain</td>
<td>300±5 Ma</td>
<td>El Valle: Bmt, Tl, Aik, Hes, Stz, Ptz, Syl, Clv, Glf, Tsm, Wtr, Bi; Ortosor: Mld, Hdl, Bi, Js-B, Js-A, Ik</td>
<td>Cu-Au calcic and magnesian skarn deposit and Au reduced skarn deposits</td>
<td>Genetically related</td>
<td>Cepedal et al. (2006)</td>
</tr>
</tbody>
</table>
about −8. The chlorine replacement reactions require log₂Te₂ to increase to the stable field of native tellurium. Log₂Te₂ and log₂S₂ in the quartz-pyrite veins should lie within the ranges −8 to −11 and −9 to −11.5, respectively. The decrease in log₂Te₂ values from the end of quartz-albite-carbonate-pyrite vein formation to the start of formation of the quartz-pyrite veins corresponds to the appearance of native Te. The associated decrease in fS₂ corresponds to pyrite crystallization, which may have caused gold precipitation. The telluride mineral assemblage indicates a highly oxidized environment, which is consistent with hematite alteration and sulfate minerals detected in the veins.

Comparison with other gold telluride deposits

Tellurides and sulfosalts are present as trace constituents in many gold deposits from different parts of the world. However, deposits in which telluride minerals are major Au reservoirs are rare. Examples include the Sandoowanzi gold deposit in China, Sacarimb in Romania, and the Golden Mile in Western Australia, with 100%, 50%, and >20% of the Au held in telluride minerals, respectively (Liu et al. 2011, Cook et al. 2005, Shackleton et al. 2003). Telluride minerals have been reported in the Yilgarn craton from the Golden Mile, Sunrise Dam, Victory, Jupiter, Chalice, Kanowna Belle, and Porphyry gold deposits (Clout 1989, Shackleton et al. 2003, Sung et al. 2007, Clark et al. 1989, Cleverly & Nugus 2008). However, only the first two of these and the Bellerophon prospect can be termed Au-tellurides deposits because of the abundant contents of telluride minerals. Representative Au-Te deposits are summarized in Table 5. These deposits cover a wide range in ages from late Archaean to Neocene.

Au-Te mineralization has been reported from a variety of types of gold deposits, with low sulfidation epithermal being the most important (Table 5, e.g., Cripple Creek, Colorado; Emperor, Fiji; El Indo, Chile). Low sulfidation epithermal gold mineralization is interpreted to be related to alkaline-calc to alkaline arc magmatism (e.g., Kelley & Ludington 2002, Cooke & McPhail 2001, Cook et al. 2009). Gold-telluride deposits from porphyry, skarn, intrusion related, and VMS types of mineralization are also interpreted to be genetically related to magmatism. All these examples suggest that the Te-enriched auriferous fluid is derived from a magmatic system. A positive correlation between Au and Te, Bi, W, As, and Pb has also been identified at the Bellerophon prospect, which is consistent with the magmatic source hypothesis (Baker et al. 2005, Meinert 2000, Tooth et al. 2008), as is the spatial relation between gold mineralization and alkaline intrusions in Bellerophon. However, further studies, including
geochronology and radiogenic isotope analyses, are necessary to support this point for Bellerophon. Another notable feature of Bellerophon mineralogy is the presence of Cl-bearing minerals. These minerals have only previously been reported from the Kolar gold deposit in India and the Sunrise Dam gold-telluride deposit in Western Australia (Table 5, Genkin et al. 1985, Sung et al. 2007, Sung 2008), where kolarite (PbTeCl$_2$) and radhakrishnaite [Pb$_2$Te$_5$(Cl,S)$_2$] have been found. However, the Cl-bearing minerals in the Bellerophon deposit are PbCl(OH) mixed with Pb$_2$Cl$_3$(OH) and BiOCl mixed with BiO(OH,Cl), which have not previously been reported from gold deposits. The gold grains in association with Cl-bearing minerals in the Bellerophon deposit are interpreted to have precipitated from a later and lower-temperature hydrothermal event.

**Conclusion**

Gold at the Bellerophon gold prospect is hosted mainly by the meta-sedimentary rocks of the Black Flag Group, but the spatially associated monzonite and albite dikes also carry significant mineralization. Four stages of mineralization have been identified: Stage I, quartz-carbonate veins; Stage II, quartz-albite-carbonate-pyrite veins and sericite-pyrite seams; Stage III, quartz-pyrite veins; and Stage IV, carbonate-chlorite/carbonate veins. Gold-bearing minerals, including native gold and Au-tellurides, crystallized during Stages II and III. Extension veins and breccia textures and sericite-pyrite seams provide evidence for brittle and ductile deformation, respectively, and both are synchronous with gold mineralization. Intense hematite and albite alterations, which envelop the mineralized veins, are interpreted to be coeval and derived from the same hydrothermal fluid that deposited the ore minerals. Thirteen telluride and sulfosalt minerals have been identified from the Bellerophon deposit, and these carry about 15% of the Au. The most common telluride minerals include calaverite, petzite, altaite, and tellurobismuthite. The mineral associations of PbCl(OH)-Pb$_2$Cl$_3$(OH)-Te$_2$O$_2$ and BiOCl-BiO(OH,Cl)-Te$_2$O$_2$ from Stage II were formed by replacement of telluride minerals during reactions with a Cl-bearing fluid. The hematite alteration, sulfate minerals in the veins, and the telluride minerals assemblage at Bellerophon all indicate a highly oxidized auriferous fluid. The telluride and sulfosalt mineralogy suggest the log$_f$S$_2$ value during Stage II decreased from about −8 to −11, while log$_f$Te$_2$ was initially about −8, before rising into the stability field of native tellurium. During Stage III, the log$_f$Te$_2$ and log$_f$S$_2$ were about −8 to −11 and −9 to −11.5, respectively. The highly oxidized nature of the tellurium-enriched auriferous fluid is consistent with the involvement of magmatic fluid during mineralization.

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