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GENETIC MODELS FOR EPITHERMAL GOLD DEPOSITS AND APPLICATIONS TO EXPLORATION

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ABSTRACT

Epithermal gold deposits are the product of large-scale hydrothermal systems in tectonically active regions. They form at shallow crustal levels where the physico-chemical conditions change abruptly.

Two major groups of epithermal gold deposits can be distinguished based on their genetic connection with: A) Copper-molybdenum porphyry systems and B) Geothermal systems related to volcanic centres and calderas.

Epithermal gold deposits connected with geothermal systems encompass three major types: adularia-sericite, acid-sulphate and disseminated replacement (the Carlin-type). Their essential ingredients are: high heat source which leads to convection of groundwater in the upper crust; source of hydrothermal fluid, metals and reduced sulphur; and high-permeability structures which allow fluid convection and metal deposition. Mixing of these ingredients leads to the formation of epithermal gold deposits throughout crustal history, without any restriction on age.

The ores were deposited from near-neutral (adularia-sericite type and some of the Carlin-type) to acidic (acid-sulphate type and porphyry-related epithermal gold deposits), low-salinity, high CO_2 and high H_2S fluids, which were predominantly meteoritic in origin. The transport capability of deep fluids in epithermal hydrothermal systems may be shown to be dependent largely on their H_2S content and, through a series of fluid mineral equilibria, on temperature and on CO_2 content. The most common mechanisms of ore deposition are boiling (phase separation), mixing of fluids of different temperatures and salinities, reaction between them and wall rocks, dilution and cooling.

An understanding of genetic models for epithermal gold deposits provides the basis for the selection of favourable areas for regional to prospect-scale exploration.

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1.0 INTRODUCTION

The term "epithermal" was introduced by Lindgren (1922, 1933) to define a class of hydrothermal metalliferous deposits (precious metals, base metals, mercury and stibnite deposits) formed at a depth of less than 1500 m and at temperatures of less than 200°C, within, or in association with, subareal volcanic rocks. According to recent geochemical studies,carried out using techniques such as fluid inclusion and stable isotope analysis, it is generally accepted that hydrothermal fluids at higher temperatures (up to 350°C) and at pressures of less than a few hundred bars were present in these deposits (Heald et al., 1987; Henley, 1991). Recent geologic, chemical and thermal studies have led to a better understanding of the epithermal environment, by stressing its relationship to currently active hydrothermal systems (White, 1981; Henley and Ellis, 1983; Henley, 1985; Berger and Henley, 1989). Epithermal deposits are the products of geothermal systems which occur in the region of active or recently active volcanic centres, where magmatic heat in the upper 5-10 Km of the Earth's crust leads to the generation of a hydrothermal convective cell (Henley and Ellis, 1983). Some epithermal gold deposits can also be localized at shallow levels above or at the margins of intrusive-related copper-molybdenum porphyry systems (Sillitoe, 1991; Mason, 1991b).

The majority of epithermal deposits are of Cenozoic and, to a lesser extent, of Mesozoic age. However, newly discovered Paleozoic epithermal gold deposits and occurrences in north-eastern Queensland (Wood et al., 1990) and in New South Wales (Bottomer, 1986; Thompson et al., 1986) have been reported. Several Archean gold deposits have been recognized as having similar features to those found in active and fossil geothermal systems, suggesting their classification as epithermal (in Rossiter, 1984). Some of these deposits have been localized in Archean rocks in Canada and Australia (in Henley, 1991). However, the number of pre-Tertiary deposits remains small (Mosier et al., 1986). Unfortunately, a comprehensive review of epithermal deposits through geologic time is not possible in the space available and mainly Cenozoic and Mesozoic (epithermal) deposits will be considered in this dissertation.

Epithermal gold deposits play an important role in the world economy, as they represent one of the principal sources of gold. The western United States, western and south-western Pacific and Central Andean deposits are the major examples of long lasting gold mining and production from lode-type and base-metal deposits (Henley, 1991; Mason, 1991b). Bingham copper porphyry deposit contained at least 854 metric tons of gold (as a by-product) and was scheduled to produce 8241 Kg per year when production resumed in 1987 (Bonham,1989). Most of the new epithermal discoveries in the western United States are small and of low grade. For example, Pinson Mine contained about 4 Mt of ore at 4 g/t gold (some epithermal deposits contain even lower grade ore; Mason, 1991b). The recently discovered Ladolam (Papua, New Guinea) and Hishihari (Japan) deposits have estimated reserves of 167 Mt of ore at 3.43 g/t gold and 1.4 Mt of ore at 70 g/t gold respectively (Henley, 1991). Fig 1.1 is a summary of ore reserves and grades for vein-style, disseminated, or breccia-hosted deposits and porphyry-related copper-gold deposits.

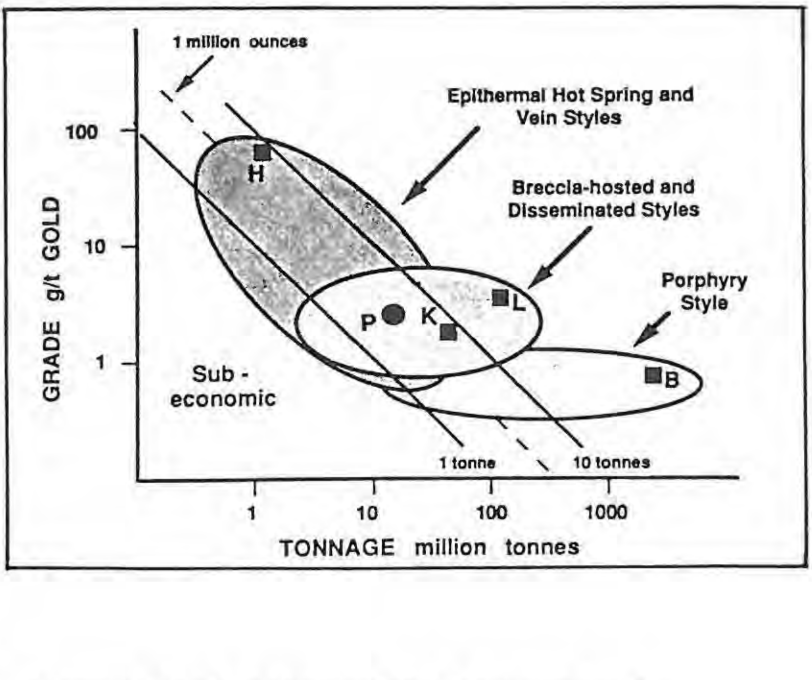


Fig. 1.1 *Summary of grade and tonnage data for epithermal- and porphyry- style gold deposits (H = Hishikari, B = Bougainville, L = Ladolam, K = Kelian, P = Paradise Peak; after Henley, 1991).*

The possibility of open-pit mining and the availability of modern techniques such as cyanide extraction and heap leaching make epithermal gold deposits economically attractive.

This dissertation will first review the chemistry and geochemistry of gold, then briefly discuss the hydrothermal geochemistry of gold and proceed to analyse the geological and geochemical environments for ore formation and implications for exploration.

2.0 CHEMISTRY AND GEOCHEMISTRY OF GOLD

Gold (Au) has an atomic number of 79 and, together with copper (Cu) and silver (Ag), belongs to group Ib of Mendeleev's periodic system. Its atomic weight is 196.967 and its specific gravity 19.3. Gold is a bright yellow, soft metal with a high melting point (1063 ^{o}C), above which it releases a violet vapour. Its boiling point is 2860 ^{o}C (Bache, 1987).

Gold has a number of oxidation states, of which the principal are zerovalent (native), monovalent (aurous) and trivalent (auric). The cycle of gold interconversions in nature is shown in Fig. 2.1.

The aqueous chemistry of gold is dominated by the auric Au(III) and to a lesser extent by the aurous Au(I) states. Higher oxidation states (Au(IV), Au(V), Au(VII)) are also known to occur, but are rare (Seward, 1991). Some of their compounds are summarized in Table 2.1.

In the compounds with Au(I) and Au(III), gold exibits an almost covalent character because of the high ionization potential of these two oxidation states. Native gold is inert and not affected by water, oxygen, sulphur or by most acids.

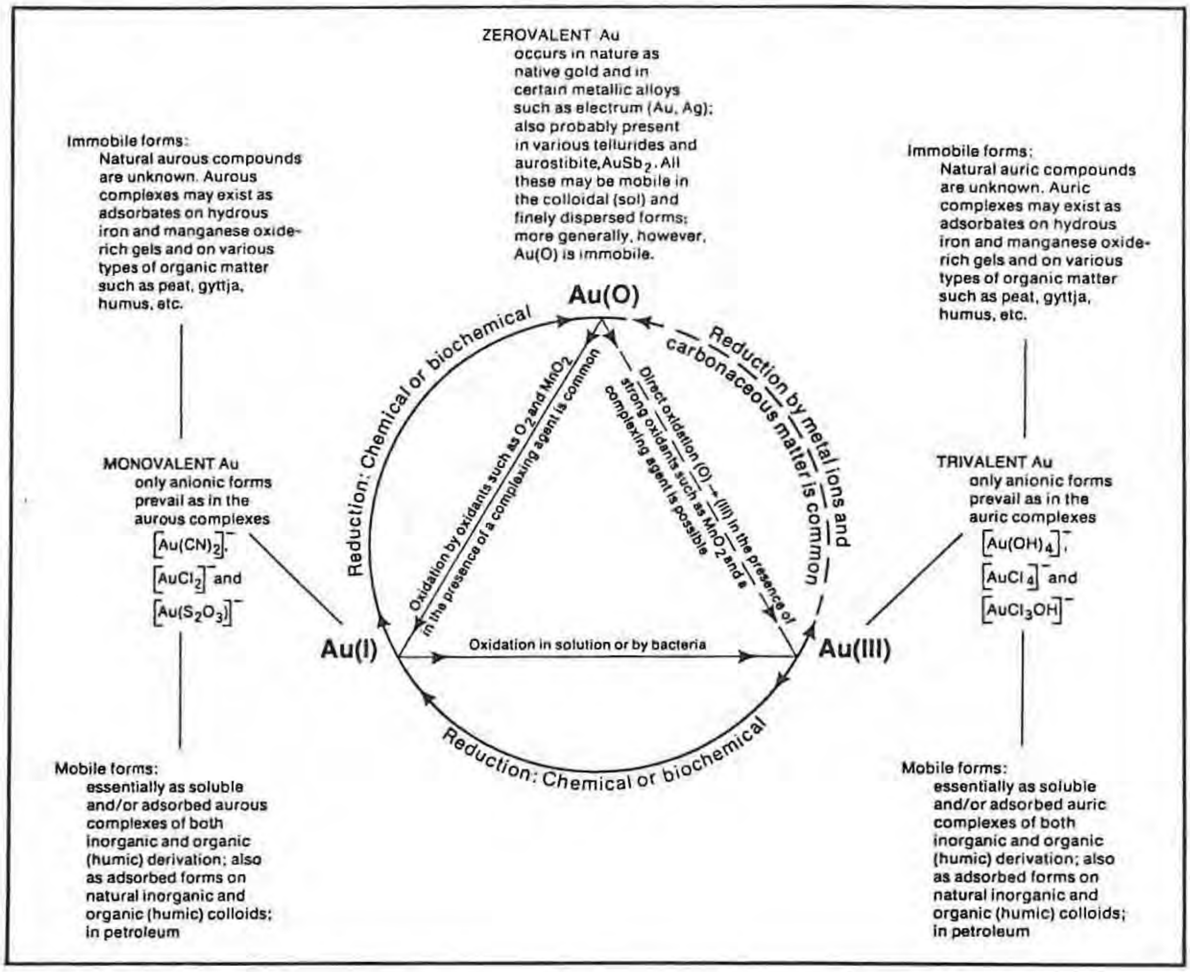


Fig 2.1 *Cycle of gold interconversions in nature (after Boyle, 1979).*

Oxidation state	Example of compounds
+2	$Au[R_2(NCS_2)]_2$
+4	$(NO_2)AuF_6$
+5	AuF_5, $CsAuF_6$
+7	AuF_7

Table 2.1 *Some oxidation states of gold (modified after Seward, 1991).*

However, it can be readily dissolved in the following solutions: alkali cyanide (in the presence of air), alkali sulphide, selenic acid, solutions containing telluric, sulphuric or phosphoric acid and sulphuric and hydrochloric acid solutions containing an oxidizing agent (i.e. MnO_2). The element is also sparingly soluble in solutions containing HCl and $CuCl_2$ (in the presence of an oxidizing agent) and in ferric chloride and ferric sulphate solutions containing an oxidizer (Boyle, 1979).

Only one stable isotope of the gold atom exists in nature: ^{197}Au. Three short-lived isotopes are known, ^{196}Au,^{198}Au,^{199}Au (Bache, 1987).

Gold is not lithophile and is more siderophile than chalcophile. It is a trace constituent of meteorites with the highest content in siderites (1.2 ppm) and the metal phase of chondrites (0.22ppm) (Boyle, 1979).

Acccording to Crocket (1991), igneous rocks containing gold can be divided into orogenic, non-orogenic and Precambrian greenstone. The average gold content of the igneous rock groups within these environments are summarized in Table 2.2.

ENVIRONMENT	ROCK GROUP	GOLD CONTENT (ppb)
OROGENIC: Plate convergent (compressive stress); island arcs; active continental margins; obductive emplacement of ophiolite.	Ophilite harzburgite	2.8
	Mafic volcanic	2.2
	Felsic volcanic	1.55
	Granitic plutons	2.6
NON OROGENIC: Stable craton settings; divergent plate margins (extensional stress); rifting.	MORB	1.2
	Intraplate basalt	2
	Flood basalt	3.5
	Initial magma; Layered gabbroic complex	4.6
	Kimberlites	3.1
	Eclogite	3.4
	Spinel lherzolites	0.5
	Alkaline plutons	2.8
PRECAMBRIAN GREENSTONE without tectonics implications	Komatiitic basalt	12.4
	Tholeiitic basalt	5.7
	Granitic plutons	1.5

Table 2.2 *Average gold content of igneous rock groups (modified after Crocket, 1991).*

Some authors have recognized a weak trend of decreasing gold content from mafic to felsic rocks, related by magma differentiation (Shilin, 1980; Korobeynikov, 1981) consistent with the decrease of gold content from early crystallizing mafic silicates and Fe-Ti-oxides to late crystallizing quartz and feldspar. As accessory (magnetite, sphene, zircon) and ferromagnesian (biotite, hornblende) minerals carry higher gold content in comparison to their host rocks, they may be gold concentrators, in particular during crystallization (Crocket, 1991). However, other authors do not support the existence of the mafic-to-felsic trend as a direct consequence of fractional crystallization. For example, according to Lee Tredoux (1986), there are no systematic variations of gold with mafic-ultramafic stratigraphy of the Bushveld Complex.

According to Mints (1975), gold variations in volcanic rocks in the USSR are related rather to source region and magma generation than to magma fractionation. Some of the recent estimations of gold in mantle rocks show variability in gold content in garnet peridotite on a regional scale and indicate that sulphide and platinum group minerals are also concentrators in mantle rocks. Hence, the upper mantle has been suggested to be a heterogeneous reservoir with respect to gold (Mitchell and Keays, 1981).

In the case of MORB magma, sulphide globules are of critical importance as they carry the highest gold content (up to 13 ppm) (in Crocket, 1991).

Volcanic rocks in a convergent environment are represented by calc-alkaline rocks of the basalt-andesite-rhyolite association, island-arc boninites and continental, bimodal, basalt-rhyolite assemblages (Crocket, 1991). The differences in gold content between felsic and mafic types (1.55 and 2.2 ppb respectively) are not well understood. They could reflect regional variations.

Average gold contents in some sediments and sedimentary rocks are shown in Table 2.3.

GROUP	GOLD CONTENT (ppb)	COMMENT
SEDIMENTS		
Clay bearing (deep sea)	3.0	generally
Globigerina and siliceous oozes	1.5	
Terrigenous sediments < 500 m water deep	3.2	
SEDIMENTARY ROCKS		
Clastic rocks (conglomerate, sandstone, siltstone)	8.1	Undino-Dainsk basin (USSR)
Chert	16.7	
Shale	2.3	
Carbonaceous shale	6.7	
Carbonate rocks, associated evaporites	1.9	
Precambrian iron formation:		
Archean (volcanic-hosted)	38	Canadian shield
Proterozoic (sediment-hosted)	19	Canadian shield
Ordovician (sediment-hosted)	7.5	Bathurst district, Canada

Table 2.3 *Average gold content in sediments and sedimentary rocks (modified after Crocket, 1991 and Boyle, 1979).*

Average gold contents in regionally metamorphosed pelites and granulites are summarized in Table 2.4.

FACIES	GOLD CONTENT (ppb)	COMMENT
Greenschist metapelites	7.1	Kamchatka, Nova Scotia, Kola, Yenisey, Platom Uplands
Amphibolites	5.9	Kamchatka, Kola, Yenisey, Platom Uplands
Granulites	2.2	Nova Scotia, Kola, Yenisey, Tien Shan

Table 2.4 *Average gold content in regionally metamorphosed pelites and granulites (modified after Crocket, 1991).*

2.1 GOLD MINERALS

Gold is usually found in the free state. The purity of native gold is expressed in terms of "fineness", defined as the proportion of pure gold in the sample in parts per thousand (native gold that is 900 fine contains 90% gold) (Boyle, 1979). Gold is usually alloyed with silver and copper and more rarely with bismuth, antimony, platinum, palladium, rhenium or iridium. It also occurs combined with tellurium or selenium. Gold can contain traces of one or several of the following elements: Li, Na, K, Be, Mg, Ca, Sr, Ba, Zn, Cd, Hg, B, Al, Ga, In, Sc, Si, Ge, Sn, Pb, Ti, Zr, As, V, Cr, Mo, W, Mn, Co, Ni, U, Th and rare-earths.

Gold is also found as microscopic inclusions in a large number of sulphides (arsenopyrite, pyrite and pyrrhotite are the most common sulphide hosts) and sulphoarsenides (Bache, 1987).

According to Boyle (1979), microscopic amounts of gold also occur in some native elements, such as arsenic, bismuth, copper, silver and platinum-group elements.

The most common auriferous minerals are presented in Table 2.5.

Name	Formula	Crystal system	Hardness (Mohs' scale)	Relative density	Au content (%)	Colour
Gold	Au	Cubic	2.5–3	19.3	100	Golden yellow
Electrum	Au–Ag	Cubic	2.5–3	12.5–15.5	55–80	Pale yellow
Maldonite	Au_2–Bi	Cubic	1.5–2	15.5	65	Silver white
Calaverite	$(Au, Ag)Te_2$	Monoclinic	2.5	9–9.5	39.5	Bronze yellow
Krennerite	$(Au)Te_2$	Orthorhombic	2.5	8.35	39.5	Yellowish white
Sylvanite	$(Au, Ag)Te_4$	Monoclinic	1.5–2.5	7.9–8.3	24.5	White
Petzite	Ag_3Au_2Te	Pseudocubic	2.5–3	8.7–9	18–25	Grey
Nagyagite	$Pb_5Au(TeSb_4)S_{5.8}$	Tetragonal	1.0–1.5	7	6–13	Grey
Fischesserite	Ag_3AuSe_2	Cubic	2.5–3	8	18–25	White

Table 2.5 *Main auriferous minerals (after Bache, 1987).*

3.0 THE HYDROTHERMAL GEOCHEMISTRY OF GOLD

The deposition of gold and auriferous minerals depends on changes in the physico-chemical conditions of the environment through which the hydrothermal fluids migrate. The precipitation of gold is due to the stability of the gold complex ions in the new conditions. It can take place over a wide range of temperatures and pressures, extending from amphibolite facies metamorphism, through lower temperature epithermal systems, to the sea-floor massive sulphide environment (Seward, 1991).

The purpose of this section is to review briefly the chemistry of gold in hydrothermal ore-forming fluids, emphasizing the role of gold complex ions in gold transport and deposition.

3.1 GOLD COMPLEXING AND SOLUBILITY IN HYDROTHERMAL SOLUTION

Ore-forming fluids are characterized by low oxidation potentials such that the dominant oxidation state of dissolved gold is +1 (Henley et al., 1985). It occurs in several Au(I) complexes which are important in gold transport and deposition. The complexation depends on temperature, pressure, pH, redox state of the fluid and ligand concentration of the hydrothermal fluids (Hayshi and Ohmoto, 1991). Some ligands which can be of importance in forming stable complexes with Au(I) in hydrothermal fluids are summarized in Table 3.1.

Cl^-	Br^-	I^-		
HS^-	S^{2-}	S_n^{2-}	$S_2O_3^{2-}$	$S_nO_6^{2-}$
$A_3S_6^{3-}$,	$Sb_3S_6^{3-}$	Te_2^{2-}		
NH_3	OH^-			
CN^-	SCN^-			

Table 3.1 *Ligands of potential importance in hydrothermal gold transport (after Seward, 1991).*

Most Au(I) complexes are linear, tri-atomic molecules (structural formula XMX). The stability of tri-atomic Au(I) and Ag(I) XMX complexes at room temperature is illustrated in Fig. 3.1.

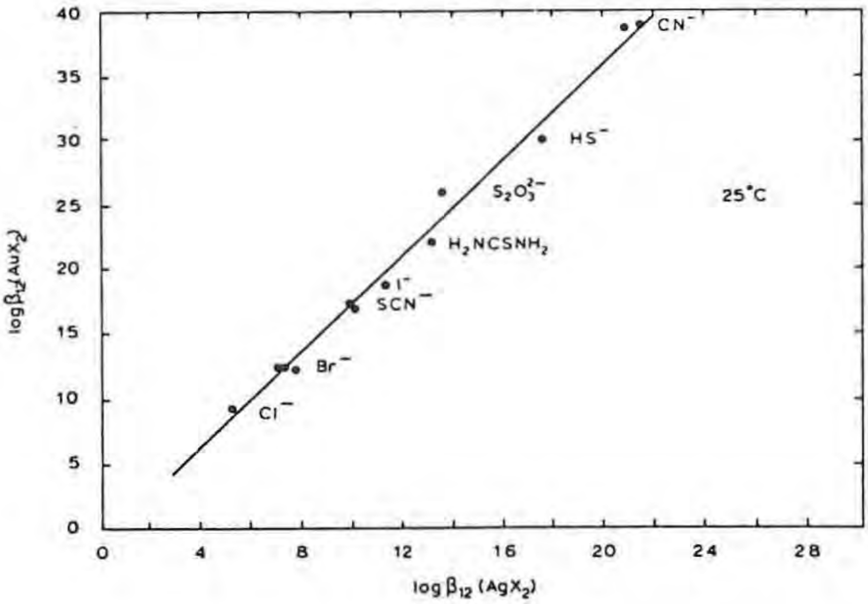


Fig. 3.1 *The change in stability of linear XMX complexes (after Renders and Seward, 1989b).*

It can be seen that the stability of Au(I) and Ag(I) XMX complexes increases from Cl^- to CN^- as the softness of X increases for the series (Renders and Seward, 1989b):

$Cl^- < Br^- < SCN^- < I^- < SC(NH_2)_2^0 < S_2O_3^{2-} < HS^- < CN^-$

For example, the gold(I) iodide complex, AuI_2 is much more stable than $AuCl_2^-$, but its concentration in hydrothermal fluids is very low and therefore it does not play an important role in gold transport in comparison with $AuCl_2^-$ (Seward, 1991). For this reason much consideration has been given to the role of chloride and reduced sulphur ligands: they both form stable complexes with gold and are sufficientely abundant in hydrothermal ore fluids (Shenberger and

Barnes, 1989). Fig. 3.1 also shows a much larger magnitude of $Au(HS)_2^-$ stability than of $AuCl_2^-$ stability, indicating the importance of the gold(I) complex in the mobilization of gold in epithermal systems (Renders and Seward, 1989b).

3.1.1 Halide complexes of gold

Gold solubility in chloride solutions at elevated temperatures has been measured by Oqryzlo (1935) who found that gold was soluble in acid chloride solutions up to 300°C. Henley (1973) also determined the solubility of gold in chloride-bearing solutions (KCl solutions) at 300°-500°C in the presence of the mineral assemblage hematite+ magnetite+ K-feldspar+ muscovite+ quartz. Anderson and Burnham (1967) demonstrated that gold was soluble in chloride solutions (HCl, KCl, NaCl) at temperatures up to 800°C and pressures up to 3 kbar. Most authors suggested a high solubility of gold as $AuCl_2^-$, based on gold solubility experiments in aqueous solutions containing KCl, NaCl and HCl at 450°C, and concluded that $AuCl_2^-$ is more important than gold-sulphide complexes under most hydrothermal conditions. Unfortunately, the redox state (e.g. f_{H2}) of these early experiments was not controlled and these data are only of qualitative importance and do not allow the determination of the stoichiometry and stability constants of the gold complexes involved. The value of these constants was first determined by Seward (1973), who experimented with aqueous sulphide solutions containing both chloride and sulphide at 160°-300°C and 1000 bar. He concluded that gold-chloride complexes are not as important as sulphide complexes in solutions of neutral to alkaline pH.

Recently, Hayashi and Ohmoto (1991) determined the solubility controlling reactions and the stability constants of gold complexes in solutions containing NaCl and H_2S at temperatures of 250°, 300° and 350°C and at pH levels from 1.9 to 5.5 and all types of redox equilibria (i.e., among aqueous species and between solution and solid phase) have been established. The results of the experiments show that gold solubility is not dependent on the activity of Cl^- and H^+ in the solutions, indicating that chloride complexes are not important, unless the fluid is H_2S^- poor, Cl^- rich and of low pH (less than 4.5).

The experiments show also that gold solubility increases with increasing $H_2S_{(aq)}$ activity, indicating that gold dissolved largely as a bisulphide complex which plays therefore a very important role in gold transport. This can be shown by the reaction (Hayashi and Ohmoto, 1991):

$$Au_{(s)} + 2H_2S_{(aq)} = HAu(HS)_2^{o} + (1/2)H_2\ _{(aq)}$$

3.1.2 Hydrosulphide and sulphide complexes

As has already been mentioned, soft ligands, such as HS^- and S^{2-} also form extremely stable complexes with Au^+ (Fig. 2.1.1) The $Au(HS)_2^-$ and $Au_2S_2^{2-}$ species, as well as the cyanide complex $Au(CN)_2^-$ are the most stable known inorganic complexes of Au^+ (Seward, 1989).

Seward (1973) was the first to demonstrate that sulphide complexes of gold(I) played a fundamental role in gold transport and deposition in hydrothermal fluids. His experiments (conducted at temperatures between 160°C and 300°C) showed that three complexes were present in the solution: $Au(HS)^o$ (acid pH), $Au(HS_2)^-$ (near neutral to weakly alkaline pH) and $Au_2(HS)S^-$ (alkaline pH). Recent work by Renders and Seward (1989b) has confirmed the stoichiometry of $Au(HS)^o$ (acid) and $Au(HS)_2^-$ (pH between 6 and 9) at 25°C and suggests that these species may be of considerable importance in the hydrothermal transport of gold at a temperature greater than 300°C. In more alkaline solutions, the species $Au_2S_2^{2-}$ was identified. Shenberger and Barnes (1989) have also recently measured the solubility of gold in aqueous sulphide solutions up to 350°C at equilibrium vapor pressures for the system. They also concluded that $Au(HS)_2^-$ is the dominant complex in the near-neutral region of pH at a wide range of temperatures. The region where gold is highly soluble as $Au(HS)_2^-$ coincides with the pyrite region, which means that gold can be transported and deposited in equilibrium with pyrite. This would explain the common association of gold with pyrite.

Fig. 3.2 shows how the stability region, and therefore the solubility maximum for $Au(HS)_2^-$ changes as a function of pH and temperature (Seward, 1991).

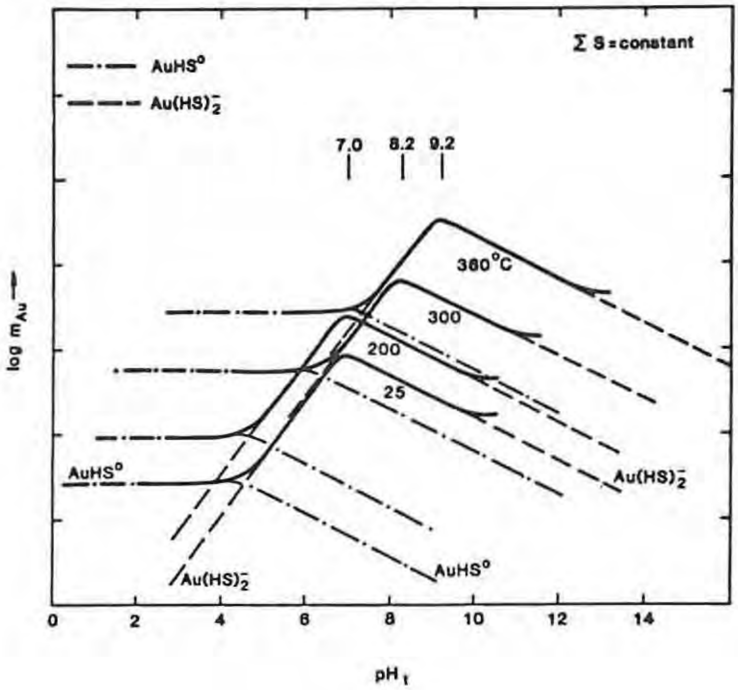


Fig. 3.2 *The stability field for hydrosulphide gold complexes (after Seward, 1991).*

It can be noticed that the maximum in solubility (and the region of stability) shifts from pH 7 to pH 9.2 as the temperature increases from 25°C to 360°C. As the pH of hydrothermal ore fluids is between 5 and 6.5, at high temperatures (greater than 300°C), $Au(HS)_2^-$ will play a reduced role in gold transport (Seward, 1991). This does not mean, however, that the sulphide complexes have lost their importance. Fig 3.2 illustrates that, for example, at a pH of about 5 and temperature of 360 °C the concentration of $AuHS^0$ will be higher than $Au(HS)_2^-$.

Complexes of gold(I) with polysulphide and the oxyanions of sulphur (e.g. $S_2O_3^{2-}$ and a polythionate such as $S_4O_6^{2-}$) can be important for gold transport from the near-surface environment to thermal springs and pools (Hedenquist and Henley, 1985; Krupp and Seward, 1987). These sulphur species are not important in high-temperature hydrothermal systems, because they disproportionate in these conditions.

3.1.3 Other ligands

Other ligands which could be important in gold transport are those involving tellurium (e.g. Te_2^{2-}, TeS^{2-}), NH_3 and CN^-. Unfortunately, there is little published information on gold solubility in high-temperature aqueous solutions containing tellurium. Smith (1944) reported that gold is soluble in high-temperature aqueous solutions containing tellurium, dissolved tellurium and gold in aqueous sulphide/polysulphide solution at 250°C, precipitating crystals of $AuTe_2$ (calaverite).

Other complexeswhich could be important in gold transport are $Au(NH_3)_2^+$, which is quite stable at 25°C, and $Au(CN)_2^-$, which is very stable at 25°C. Unfortunately there are no high-temperature thermodynamic data pertaining to the stability of these complexes (Seward, 1991).

3.2 BOILING AND GOLD DEPOSITION

There are several processes (e.g. fluid-rock interaction, boiling, mixing) which are important for the chemical evolution of fluids and the final deposition of gold. In this section the most important characteristics of the boiling process will be reviewed, to give a basis for the understanding of the fundamental process responsible for gold precipitation in epithermal systems.

A hydrothermal fluid ascending buoyantly to the surface reaches a level where boiling and/or phase separation take place (Seward, 1991). The depth at which boiling commences varies greatly with the gas content of the ascending fluid (Seward, 1989). The boiling starts at the point where the confining hydrostatic pressure is equal to the equilibrium-saturated vapour pressure. At this point, two-phase conditions predominate. When an aqueous solution boils, volatile components are forced into their vapour phase. Gaseous components, e.g. H_2S and CO_2 (which precipitate in hydrolitic equilibria), are more soluble in the liquid phase, whereas gases such as H_2 and CH_4 partition into the gas phases during boiling (Seward, 1989). The loss of three important volatile components (H_2S, CO_2, H_2) during this adiabatic, closed-system boiling has a major effect on the chemistry of the residual liquid phase (Seward, 1989; Krupp and Seward, 1987). Some of these changes, which are important in determining the transport and deposition of gold, are shown in Fig. 3.3.

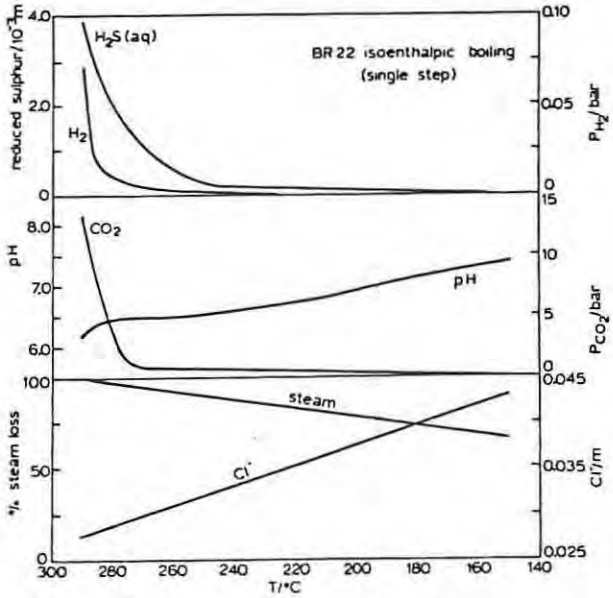


Fig. 3.3 *Changes in several important chemical parameters during adiabatic, closed-system boiling (after Seward, 1989).*

There is a significant loss of reduced sulphur and hydrogen resulting in diminition of the oxidation potential of the residual liquid phase. The concentration of CO_2 falls drastically, causing a significant increase in pH and an associated increase in the stability of $Au(HS)_2^-$. The increase in chloride concentration results from water particles lost to the steam phase. Fig 3.4 shows the changes in gold solubility in the residual liquid as boiling proceeds.

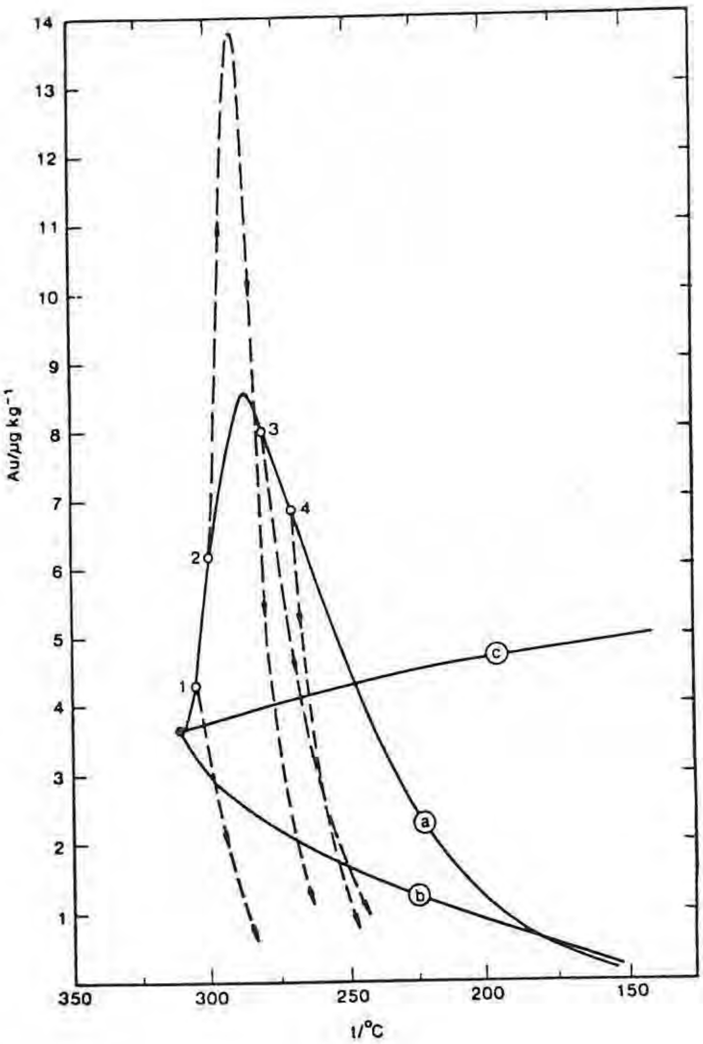


Fig. 3.4 *The effect of boiling on the solubility of gold (as* $Au(HS)_2^-$*) in Rotokwa deep fluid; curve a = adiabatic closed-system boiling; curve b = conductive cooling; curve c = increase in initial gold concentration, due to steam loss. The broken line departing from points 1,2,3 and 4 (305°, 300°, 280° and 270°C) represents open-system boiling following adiabatic, closed-system boiling to those points (after Seward, 1991).*

The gold concentration increases due to the steam loss (along curve c). Precipitation commences at 250°C, where curves 'a' and 'c' intersect. Most of the gold will precipitate within the temperature interval 250°-180°C. Curves 'a' and 'b' do not play an important role in gold precipitation. The gold concentration of the residual liquid will not increase along curve 'a' and gold would not precipitate. Curve 'b' represents the decrease in gold solubility as a result of conductive cooling, which is not of great importance in this process (Seward, 1991).

Apart from the closed-system boiling discussed above, open-system boiling also occursin active hydrothermal systems. Fig. 3.4 also illustrates the effect of multi-step open-system boiling on gold solubility. Adiabatic boiling, together with discontinuous (open-system) boiling gas-phase removal (points 1,2,3,4), leads to the precipitation of virtually all gold in solution over a small temperature interval (Seward, 1989).

Apart from boiling, some other processes such as mixing, oxidation and sulphide mineral deposition can decrease the activity of reduced sulphur, leading to gold precipitation. These processes are only important when considering $Au(HS)_2^-$ and $AuHS^o$.

The processes of adsorption, absorption, surface precipitation or reduction are also important in controlling gold concentration at lower temperatures. One of the first studies which investigated the adsorption of gold onto colloidal As_2S_3 and Sb_2S_3 up to 90^oC has been carried out by Renders and Seward (1989a). They concluded that gold (as $AuHS^o$) can be removed in large quantities from solution by arsenic and antimony sulphide colloids within a specific range of pH. If a sulphide colloid forms in a hydrothermal ore-depositing environment (e.g. at higher temperatures), it can crystallize to give a compound. Scavenging onto the charged surface of the colloid can be very important in the deposition of gold from the solution. Precipitation onto mineral surfaces such as pyrite at elevated temperatures could also be significant (Seward, 1991).

4.0 EPITHERMAL DEPOSITS

Epithermal gold deposits are diverse, reflecting the variety of geological environments (igneous, tectonic and structural) in which they occur, the many processes involved in their formation and prevailing physico-chemical conditions. However, there is one feature common to all of these deposits, namely their association with fault zones and centres of magmatism in tectonically active regions (Mason, 1991b). Regional faults control epithermal deposits by guiding the emplacement of the magmatic heat source and influencing subsequent hydrothermial activity (Hedenquist, 1986).

Epithermal gold deposits form at shallow crustal levels where the physico-chemical conditions change abruptly. Some factors influencing these conditions include: geology (structure, stratigraphy, intrusions and rock type), which affects the style and degree of permeability; pressure and temperature; hydrology; chemistry of the mineralizing fluid (White and Hedenquist, 1990).

There are various classifications of epithermal gold deposits, based on depth relation, mineralogy and alteration, host rocks, genetic models, etc. Jiles and Nelson (1982; Heald et al.,1987) proposed a depth-related classification, focussing on the relationship between epithermal districts and convergent plate boundaries. They divided these deposits into: near-surface hot spring deposits, disseminated replacement deposits and deeper bonanza systems. Based on mineralogy and alteration, epithermal deposits can be classified into: high and low sulphur bonanza and high and low sulphur hot spring types (Silberman and Berger, 1986), high sulphur and low sulphur (Bonham, 1986,in White and Hedenquist, 1990), acid-sulfate types and adularia-sericite types (Hayba et al., 1985). Commonly, epithermal gold deposits are classified on the basis of their host rocks e.g. volcanic-hosted, sediment-hosted, carbonate-hosted (Bagby and Berger, 1985; Hayba et al., 1985; Berger and Henley, 1989; Henley, 1991).

Various genetic models have been constructed. The classic model proposed by Buchanan (1981) for epithermal vein deposits involves vertical distribution of alteration with the mineralogy related to the depth of first boiling. Sillitoe and Bonham (1984) suggested that shallow precious and base metal deposits in volcanic rocks can be classified into four genetic subtypes characterized by four volcanic landforms: stratovolcano, flow dome complex, caldera and maar.

The possible relationships between epithermal deposits and porphyry systems have been discussed by some investigators (Sillitoe,1973, 1983, 1989, 1991; Mason,1991a, 1991b). Many major gold deposits in the western American Cordillera and the western Pacific region are associated with porphyry systems centred on felsic intrusive centres (Sillitoe,1979, 1987). According to Mason (1991a), many similar features can be observed in hydrothermal (mesothermal and epithermal) gold deposits of all ages and porphyry copper and molybdenum deposits. They include large zones of hydrothermal alteration (potassic, phyllic, argillic and propylitic) which define mineralized areas. Another important feature is the extent of intense fracturing at scales from several kilometres down to microscopic, which is often directly related to the size of an individual hydrothermal system and its ore deposits. These characteristics combined with the common association with epizonal felsic intrusions, breccia pipes and breccia dikes, suggest the examination of epithermal gold deposits in terms of magmatic hydrothermal models for porphyry deposits rather than geothermal models, when considering them as offspring of such systems (Mason, 1991a).

Therefore two major groups of epithermal gold deposits can be distinguished (and will be described in this dissertation) based on their genetic connection with:

A) Copper-molybdenum porphyry systems and
B) Geothermal systems related to volcanic centres and calderas.

Geothermal systems related to volcanic centres and calderas encompass three major types of deposits : acid-sulphate (also known as kaolinite-sulphate), adularia-sericite and disseminated replacement type of deposits. In each case the magmatism tends to be felsic but may be bimodal, calc-alkaline or alkaline (Mason, 1991a).

Major epithermal and epithermal-porphyry deposits are located along the Cordilleran-type central Andean margin and in the western Pacific island arcs. Figures 4.1, 4.2, 4.3, 4.4 and 4.5 show the locations of the principle epithermal and gold-rich porphyry deposits in the western Pacific, the western United States, Chile and Queensland, Australia.

In this chapter general characteristics rather than descriptions of individual epithermal gold deposits will be presented.

4.1 EPITHERMAL GOLD DEPOSITS RELATED TO PORPHYRY (Cu-Mo) SYSTEMS

Many stockwork porphyry copper-molybdenum deposits are major producers of by- or co-product gold and silver. Some of the most significant are Bingham, Butte and Bisbee in the western United States (Fig. 4.4 A); Panguna, Lepanto, Ok Tedi, Santo Tomas II and Dizon in the western Pacific (Fig. 4.1, 4.2); Andacollo and El Hueso in Chile (Fig. 4.3); Mt Leyshon and Highway (Fig. 4.5) in Queensland, etc. (Sillitoe, 1979,1991; Mason, 1991b; Morrison, 1988).

4.1.1 Geotectonic settings

Most intrusion-related Cu and Au deposits form a series of linear belts generated at Phanerozoic convergent plate margins above zones of active subduction (Sillitoe 1987, 1991). Fig. 4.6 shows the western Americas, southwest Pacific and Alpine porphyry belts in relation to Mesozoic-Cenozoic orogenic belts and accreting and consuming plate boundaries.

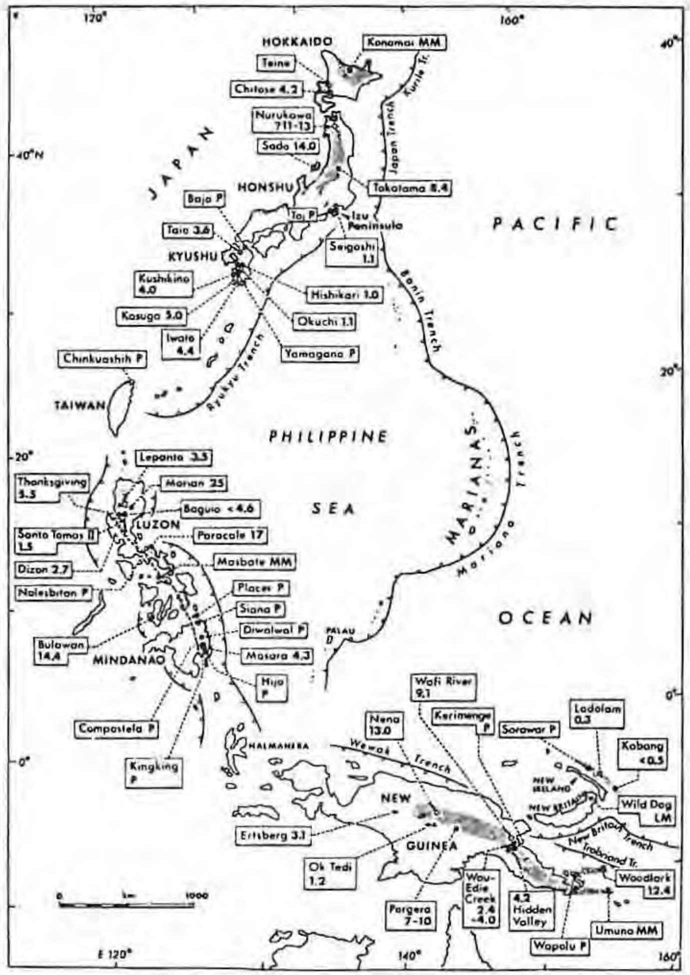


Fig. 4.1 *Principal gold deposits in western Pacific island arcs: Japan to Papua New Guinea. See Fig. 4.2 for legend (after Sillitoe, 1989).*

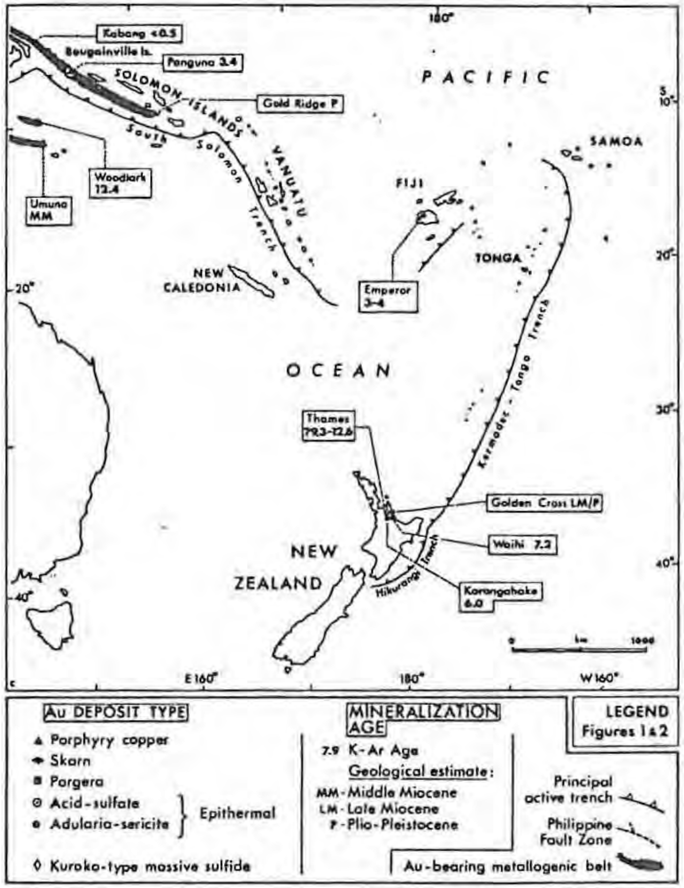


Fig. 4.2 *Principal gold deposits in western Pacific island arcs: Solomon Island to New Zeland (after Sillitoe, 1989).*

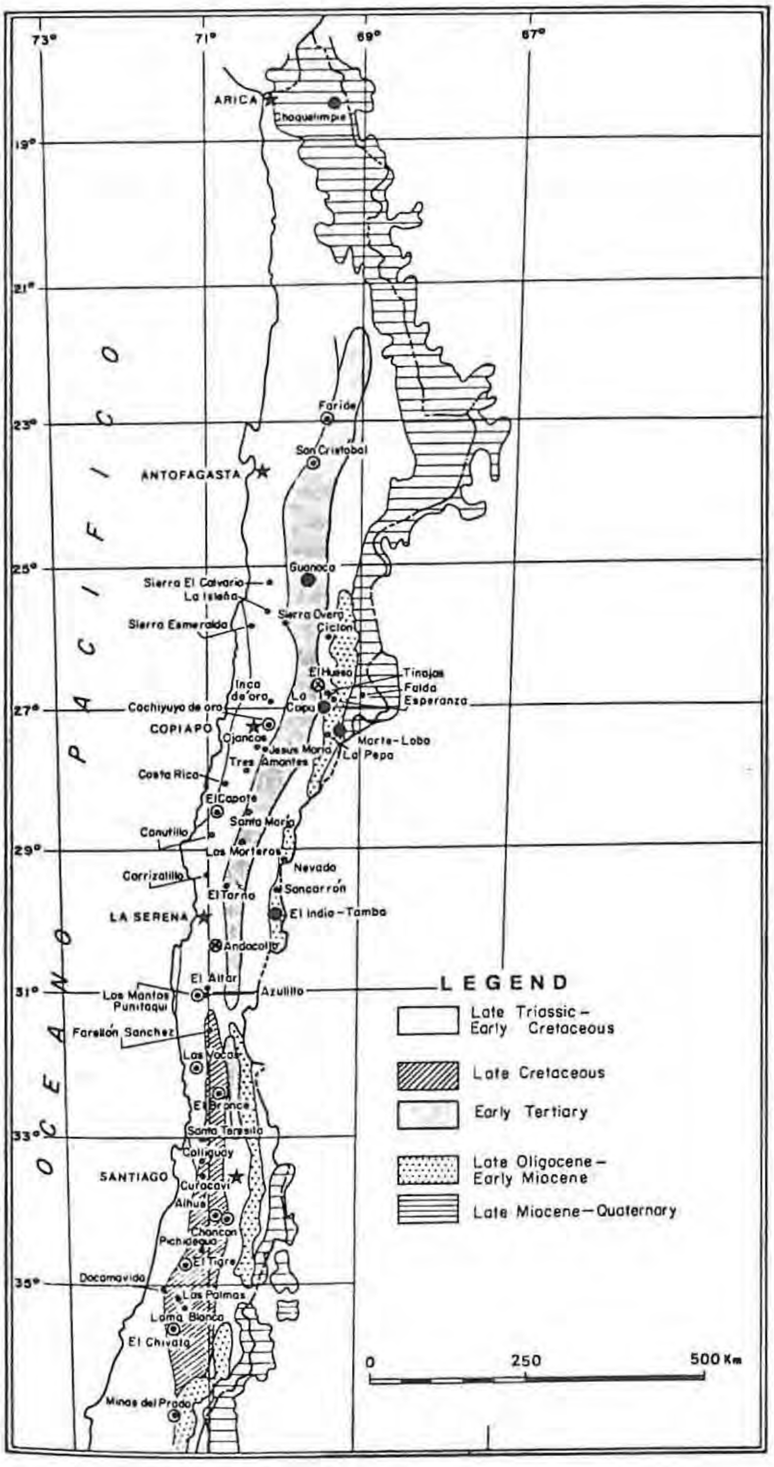


Fig. 4.3 *Location map of Chilean hydrothermal gold deposits. Simbols: gold deposits; adularia-sericite type; acid-sulphate type; porphyry-related gold deposits (modified after Camus, 1990).*

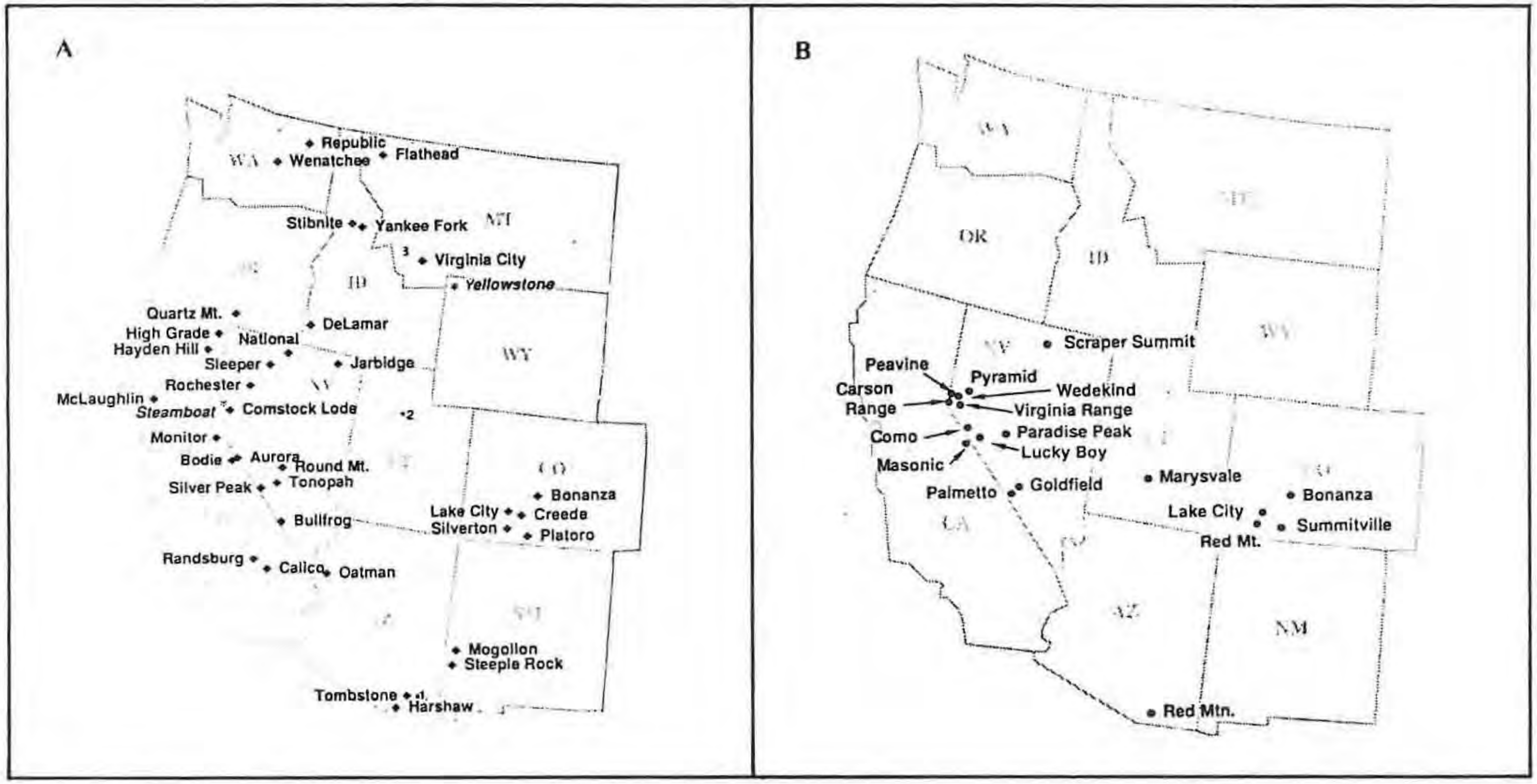


Fig. 4.4 *Generalized location map of A) adularia-sericite type deposits; porphyry copper deposits: 1 Bisbee, 2 Bingham, 3 Butte; and B) acid-sulphate type deposits in the western United States (modified after Berger and Henley, 1989).*

Fig. 4.5 *Distribution of gold deposits in north Queensland, Australia. Provinces: G = Georgetown; GR = Greenvale; CT = Charters Towers; H = Hodgkinson; BR = Broken River; DR = Drummond; CRIP = Coast Range Igneous Province (modified after Morrison, 1988).*

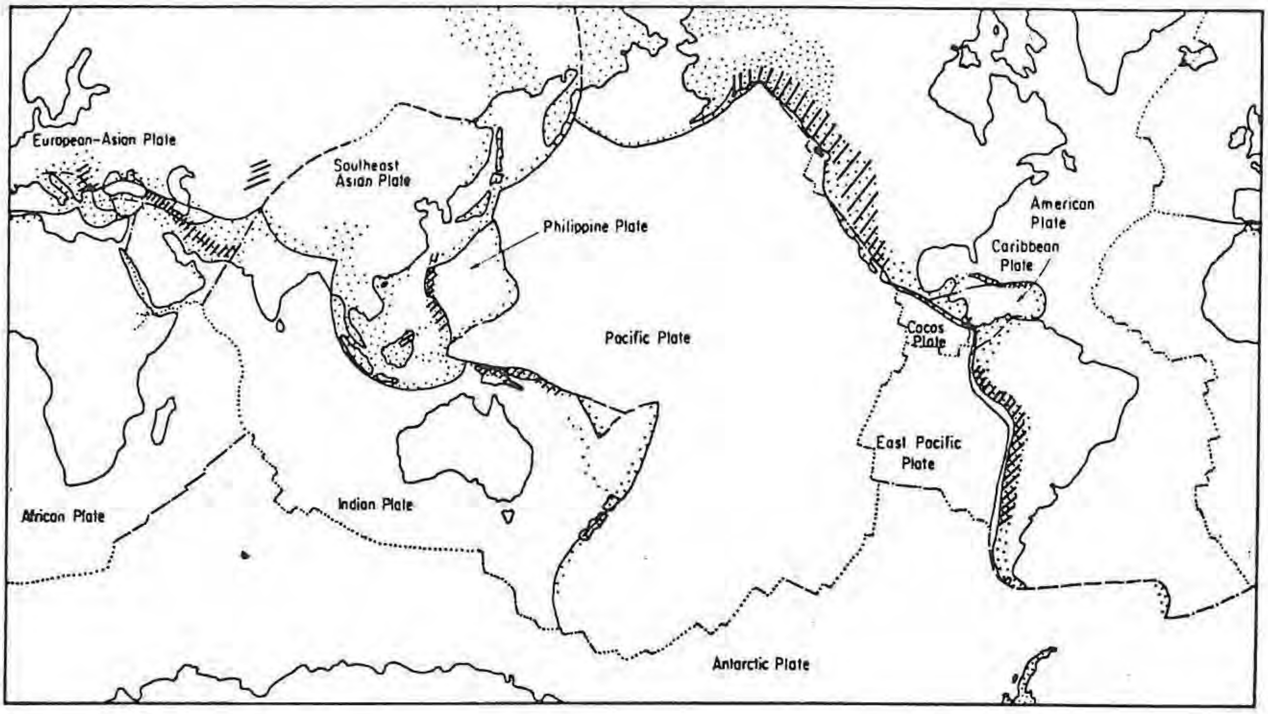
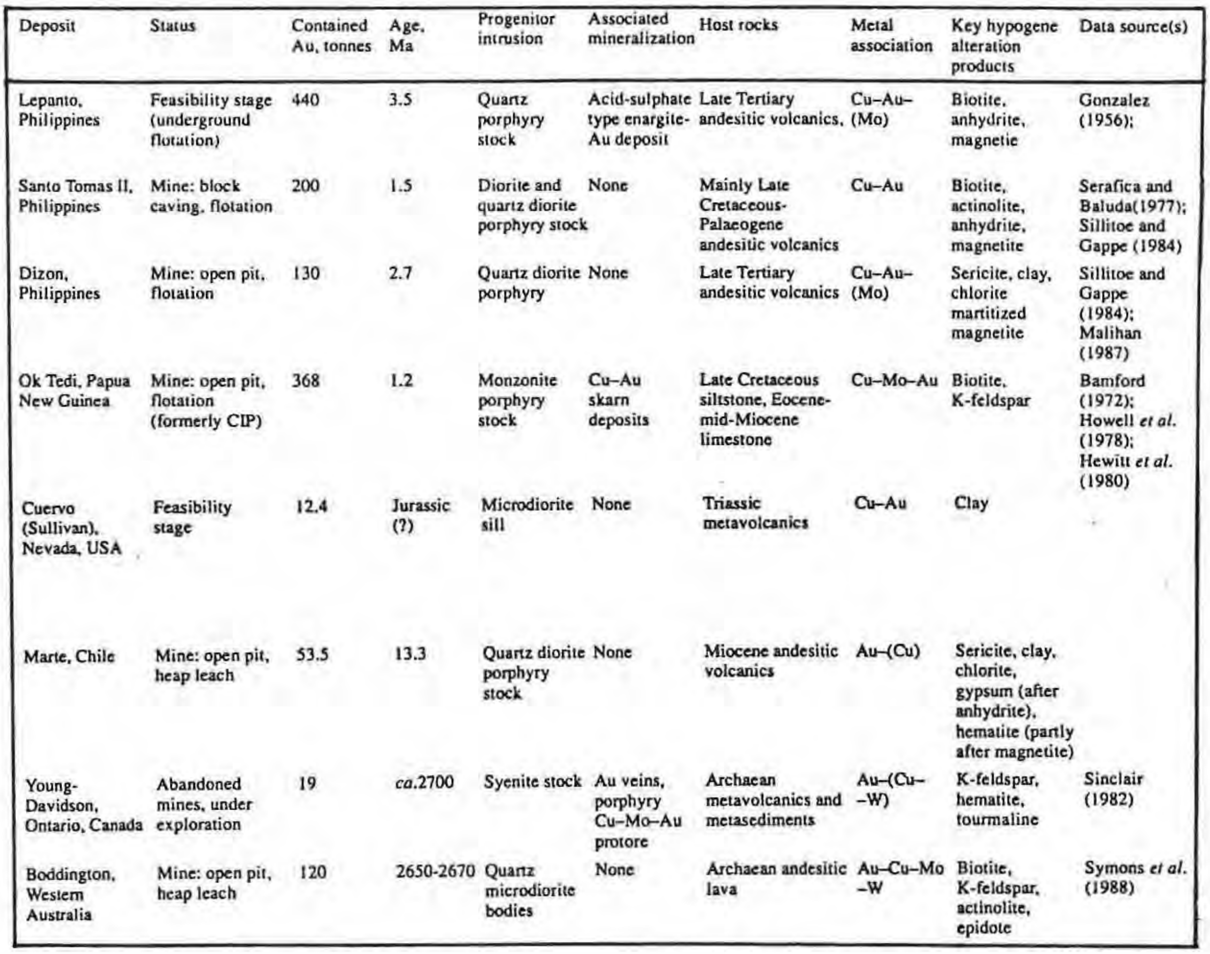


Fig. 4.6 *Major Mesozoic-Cenozoic porphyry belts in the world (after Sillitoe, 1972)*

Ideal tectonic settings for the generation of volcano-plutonic-related gold deposits are represented by the convergent margins ranging from primitive, through mature island arcs to continental margins (Sillitoe, 1987, 1991,; Mason, 1991b). For example, along the eastern Pacific margin volcano-plutonic mineralized arcs are mainly of the Cordilleran type, while they form island arcs along the western Pacific. For the gold metallogeny in the western Pacific, the most critical plate tectonic process was the subduction of the Pacific and marginal basin lithosphere and the consequent construction of volcano-plutonic arcs (Sillitoe, 1989). The arcs were constructed by I-type magnetite-series felsic magmas generated as a result of hornblende melting (Ishihara, 1981). According to Gustafsen (1979) and Sillitoe (1987), copper and gold deposits are associated with a wide variety of felsic magma composition. The felsic magmatism in extensional arcs often exibits a bimodal character because of the presence of small amounts of coeval basaltic rocks (Sillitoe, 1989). Bimodality is present in north- eastern Japan and northern Queensland, Australia. Advanced extensional back-arc environments are characterized by alkaline magmatism.

Some porphyry Cu(-Mo)-Au deposits and their characteristics are summarised in Table 4.1.



Deposit	Status	Contained Au, tonnes	Age, Ma	Progenitor intrusion	Associated mineralization	Host rocks	Metal association	Key hypogene alteration products	Data source(s)
Lepanto, Philippines	Feasibility stage (underground flotation)	440	3.5	Quartz porphyry stock	Acid-sulphate type enargite-Au deposit	Late Tertiary andesitic volcanics,	Cu–Au–(Mo)	Biotite, anhydrite, magnetie	Gonzalez (1956);
Santo Tomas II, Philippines	Mine: block caving, flotation	200	1.5	Diorite and quartz diorite porphyry stock	None	Mainly Late Cretaceous-Palaeogene andesitic volcanics	Cu–Au	Biotite, actinolite, anhydrite, magnetite	Serafica and Baluda(1977); Sillitoe and Gappe (1984)
Dizon, Philippines	Mine: open pit, flotation	130	2.7	Quartz diorite porphyry	None	Late Tertiary andesitic volcanics	Cu–Au–(Mo)	Sericite, clay, chlorite martitized magnetite	Sillitoe and Gappe (1984); Malihan (1987)
Ok Tedi, Papua New Guinea	Mine: open pit, flotation (formerly CIP)	368	1.2	Monzonite porphyry stock	Cu–Au skarn deposits	Late Cretaceous siltstone, Eocene-mid-Miocene limestone	Cu–Mo–Au	Biotite, K-feldspar	Bamford (1972); Howell *et al.* (1978); Hewitt *et al.* (1980)
Cuervo (Sullivan), Nevada, USA	Feasibility stage	12.4	Jurassic (?)	Microdiorite sill	None	Triassic metavolcanics	Cu–Au	Clay	
Marte, Chile	Mine: open pit, heap leach	53.5	13.3	Quartz diorite porphyry stock	None	Miocene andesitic volcanics	Au–(Cu)	Sericite, clay, chlorite, gypsum (after anhydrite), hematite (partly after magnetite)	
Young-Davidson, Ontario, Canada	Abandoned mines, under exploration	19	*ca.*2700	Syenite stock	Au veins, porphyry Cu–Mo–Au protore	Archaean metavolcanics and metasediments	Au–(Cu–W)	K-feldspar, hematite, tourmaline	Sinclair (1982)
Boddington, Western Australia	Mine: open pit, heap leach	120	2650-2670	Quartz microdiorite bodies	None	Archaean andesitic lava	Au–Cu–Mo–W	Biotite, K-feldspar, actinolite, epidote	Symons *et al.* (1988)

Table 4.1 *Characteristics of some of the Cu(-Mo)-Au porphyry deposits (modified after Sillitoe, 1991).*

4.1.2 Hydrothermal alteration and type of mineralization

Gold-rich copper- or molybdenum- dominated porphyry stockworks are present within prophyry stocks and in their immediately adjoining wall rocks. The stocks are commonly subvolcanic in character and range in composition from diorite and quartz diorite porphyry at Lepanto, Santo Tomas II, Dizon and Marte through syenite at Young Davidson to monzonite porphyry at Ok Tedi (Table 4.1) Gold-copper (Mo) mineralization was introduced with K-silicate alteration, rich in biotite and anhydrite, which in some deposits are accompanied by K-feldspar (Ok Tedi) or actinolite (Santo Tomas II) (Table 4.1) (Sillitoe, 1991).

Gold is present with a low pyrite content chalcopyrite-bornite assemblage at Lepanto and Santo Tomas II and a similar assemblage is supposed to have existed prior to supergene oxidation at Ok Tedi. A close association between gold and copper-bearing sulphides and a positive correlation between them indicate their co-transport and co-precipitation, providing evidence against the supposition that gold is introduced later than the K-silicate-copper association (Sillitoe, 1991). The pyrite content of gold-copper ore at Dizon was increased during overprinting of K-silicate alteration by a sericite-illite-clay-chlorite assemblage of epithermal character (Sillitoe, 1989), which induced little modification of gold and copper grades. Five types of hydrothermal alteration have been recognised in the Dizon gold-rich porphyry copper deposits, developed within the stock and the volcanic wallrocks (Malihan, 1987). They are represented by 1) centrally located K-silicate alteration, developed largely within the intrusion; 2) a phyllic-argillic or quartz-sericite-clay-chlorite assemblage overprinting the former; 3) an advanced argillic alteration; 4) an outer propylitic zone and 5) a sericitic alteration (Fig. 4.7). Ore- grade copper-gold mineralization is largely found in the K-silicate and the phyllic zones and to some extent in the advanced argillic zone.

Detailed information about these deposits can be obtained from references listed in Table 4.1.

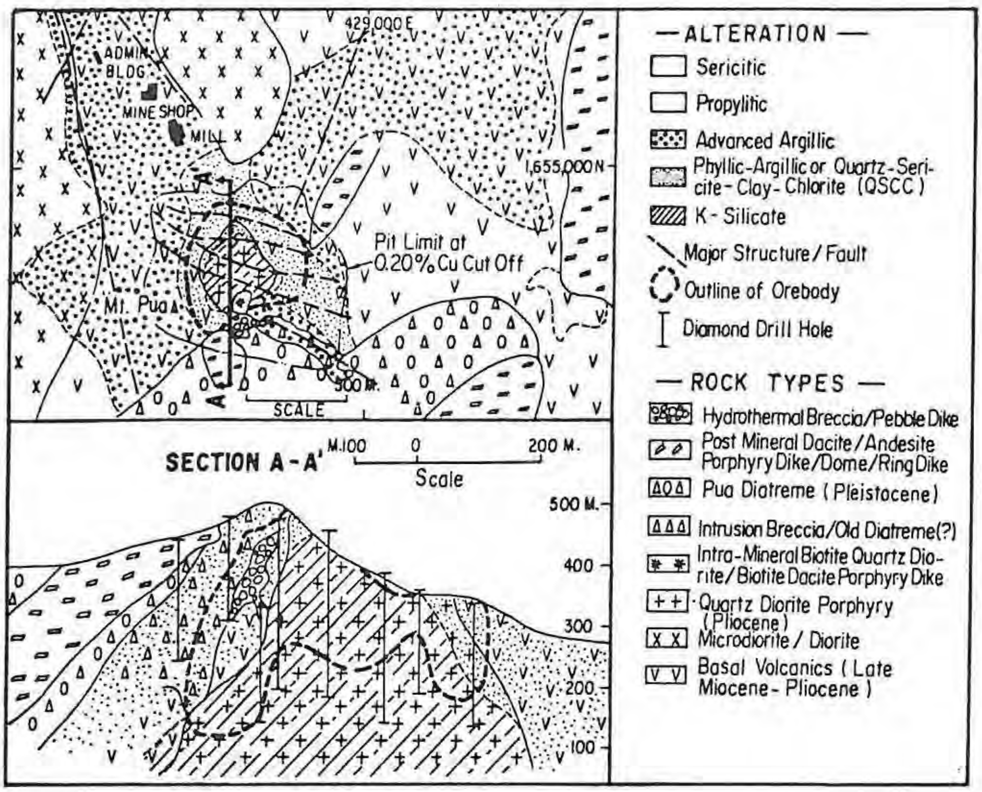


Fig. 4.7 *Geological and alteration map of the Dizon porphyry Cu-Au deposit (after Malihan, 1987).*

At Lepanto, the enargite-bearing massive sulphide deposit occurs within zones of advanced argillic alteration in the upper volcanic portions of, but marginal to, the porphyry copper system. Lense-like ore bodies occur in andesitic (and subordinate dacitic) volcanic rocks. The deposit was generated by replacement of andesitic host rocks and the pyroclastic fill of the diatreme. The prinicipal constituents of the ore bodies are massive fine-grained pyrite and subordinate marcasite. Enargite and luzonite are typically coarse-grained and commonly either replaced earlier iron sulphides as veinlets and patches, or filled in remnant open spaces within them. Gold is present as the native metal, but tellurides are also present. Minor metals are tin and bismuth. Massive chalcedonic silica and subordinate crystalline (hypogene) alunite occur as a component of massive ore and in a halo around it. Silicification tends to grade outward to intermediate argillic alteration beyond the ore zones (Sillitoe, 1983). Hydrothermal breccias are a widespread part of the massive enargite-bearing deposit and host significant amounts of ore.

Other massive sulphide deposits with similar characteristics to Lepanto are: Recsk in Hungary, Frieda River in Papua and Bor in Yugoslavia (Sillitoe, 1983).

The alteration and mineral assemblages of the above mentioned deposits suggest that they are equivalents of theacid-sulphate type of epithermal deposit (Heald et al., 1987, Henley 1991; Sillitoe 1989, 1991), the genesis of which will be discussed in chapter 5.0.

4.1.3 Relationships between porphyry systems and epithermal style of gold mineralization

The genetic relationships between porphyry Cu and/or Mo gold-rich deposits and other styles of gold mineralization can be reconstructed (Fig. 4.8).

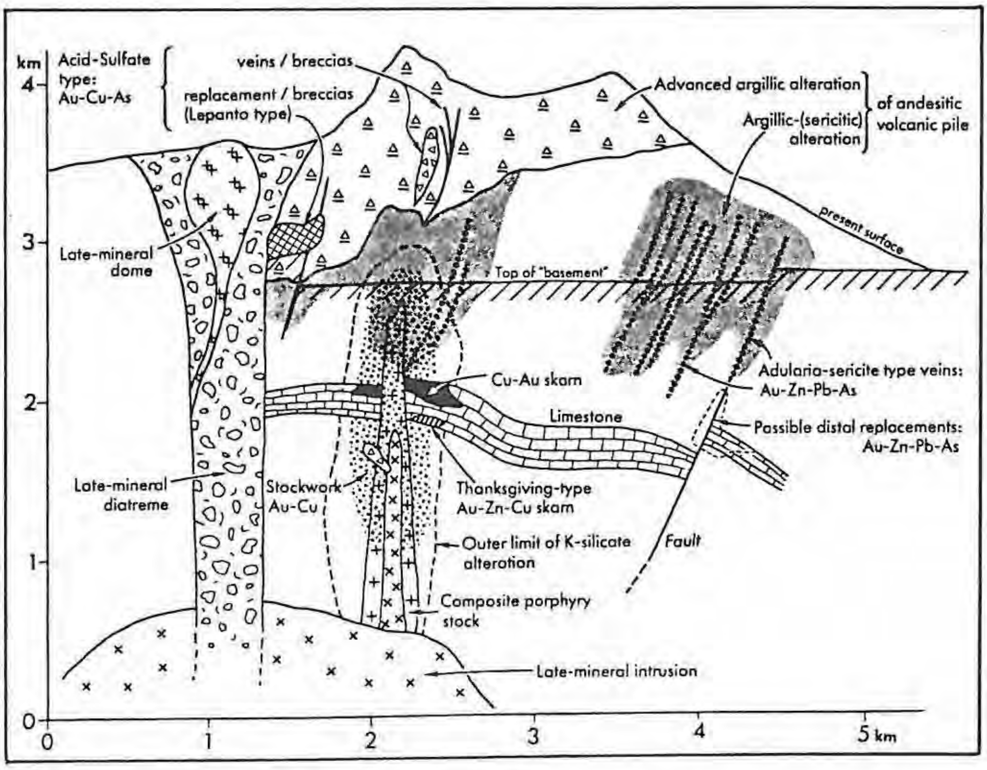


Fig. 4.8 *Schematized relationship between porphyry systems and other styles of gold mineralization (after Sillitoe, 1989).*

Very common relationships exist between porphyry-type Cu(-Mo)-Au deposits within stocks and Cu-Au skarn deposits, which in turn give way to carbonate-replacement-type Pb-Zn-Ag mineralization with or without gold (Sillitoe, 1991).

Porphyry-type Cu(-Mo)-Au deposits can be spatially and genetically associated with both acid-sulphate and adularia-sericite types of epithermal gold deposits (Fig. 4.8). Acid-sulphate gold deposits (Lepanto, Frieda River, Equity Silver) were generated as late-stage events in porphyry systems in association with advanced argillic alteration. They occur above the porphyry stock or alongside intrusions, as at Porgera, Gilt Edge, Kori Kollo etc.

Adularia-sericite epithermal deposits can be peripheral to, or up to 4 km from, mineralized porphyry stocks. Examples of the latter are found in the western Pacific e.g. Marian, Paracele and Umuna.

At Ladolam (New Guinea), anadularia-sericite-type gold deposit is superimposed on the upper parts of a mineralized porphyry stock (Sillitoe, 1991). The Ladolam deposit is located within the Luise collapse caldera, in active geothermal systems with boiling hot springs, solfatara and submarine discharge into Luise Harbour (Fig. 4.9).

Extensive breccias, derived from volcanic, intrusive and hydrothermal events, host the deposit in volcanics and monzonitic intrusivons. Alteration is represented by early porphyry-style potassic and propylitic assemblages, overprinted by advanced argillic and phyllic alteration associated with gold mineralization. Mineralization is represented by pyrite-marcasite with minor arsenopyrite, sulpho-salts and gold-silver tellurides.

This deposit is characterized by early, subvolcanic porphyry- style alteration and associated mineralization leading to the subsequent development of a late evolutionary stage represented by current hydrothermal activity, following caldera collapse. This resulted in telescoped alteration zoning and overprinting of alteration assemblages (Sillitoe, 1991).

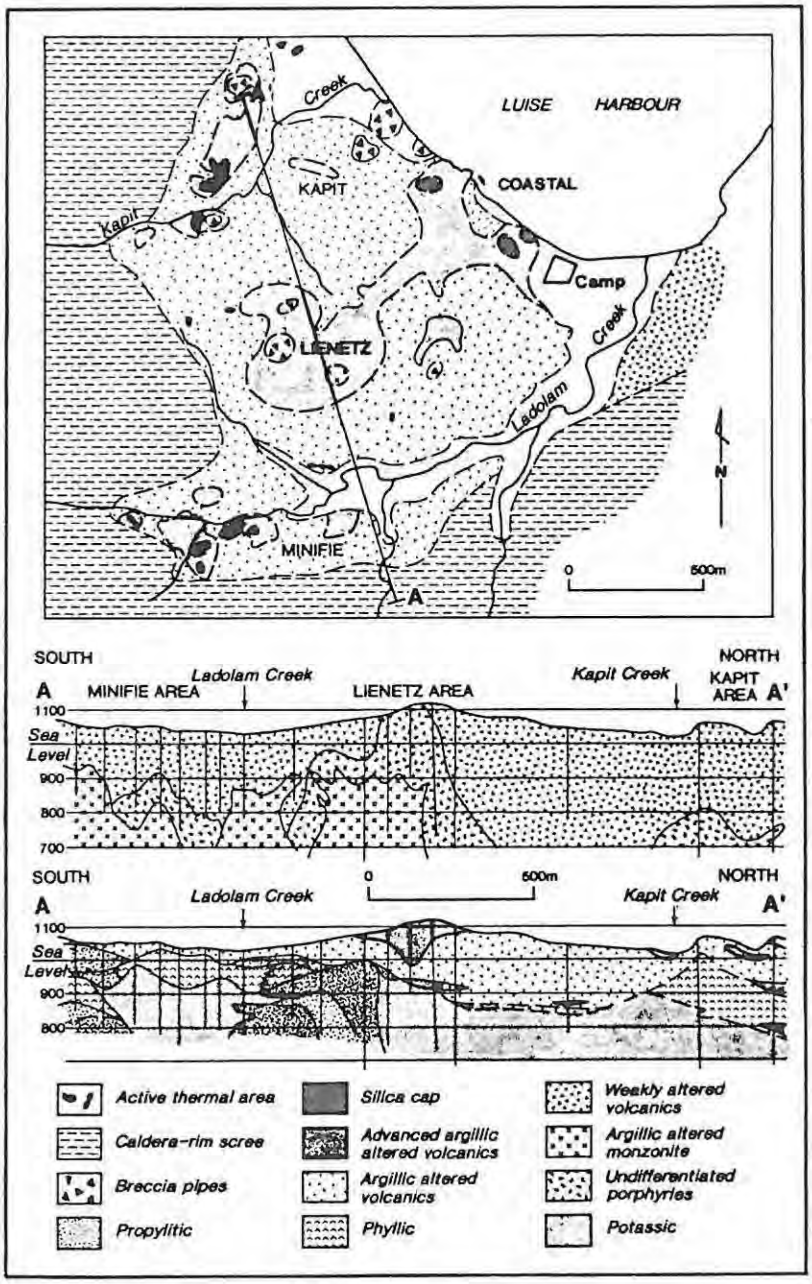


Fig. 4.9 *Ladolam deposit, Lihir Island, Papua, New Guinea (After Sillitoe, 1991).*

4.1.4 Genetic model for epithermal gold deposit related to porphyry Cu(-Mo)-Au systems

For the genesis of porphyry Cu(-Mo)-Au deposits and other styles of gold mineralization related to theformer (discussed above), three hydrothermal fluid regimes were recognised by Henley and McNabb (1978) and Eastoe (1982): relatively dense magmatic-hydrothermal brines which tend to reflux within the cooling intrusions; lower -density ascendent plumes, containing sulphur-rich magmatic volatiles which condense upward and mix with meteoritic fluids and marginally located meteoritic convective cells that "collapse" progressively onto the stock as magmatic-hydrothermal fluid decreases (Fig. 4.10).

Magmatic-hydrothermal brines are responsible for the formation of porphyry Cu(Mo)-Au mineralization, characterised by K-silicate alteration. According to Burnham and Ohmoto (1980), the formation of hydrothermal ore deposits, whether of the geothermal, vein, skarn or porphyry type, is dependent upon processes that operate during the late stages of felsic magmatism. The effectiveness of these processes depends on : magma composition including metal, sulphur, chloride and especially H_2O content; geological environment, including depth of magma emplacement. The content of H_2O in the initial melts is important as it controls formation of the aqueous phase and enhances the solubility of metal sulphides in initial melts and thereby provides the hydrous magmas with a high sulphur-carrying capacity. It also leads to the crystallization of hornblende from the magmas, upon cooling at depths greater than about two kilometres, and therefore provides a possible exploration guide for those intrusive bodies with which hydrothermal ore-forming processes are associated.

The initial H_2O content of magma is approximately 2 - 4 wt% which is enough to result in second boiling. Second boiling releases large amounts of mechanical energy which can produce explosive volcanism or extensive fracturing and also produce an aqueous phase (hydrothermal fluid) into which chlorides and sulphur are strongly partitioned (Burnham and Ohmoto, 1980). Chlorides occur as neutral complexes with alkalies, iron, hydrogen and calcium, while sulphur exists as both H_2S and SO_2.

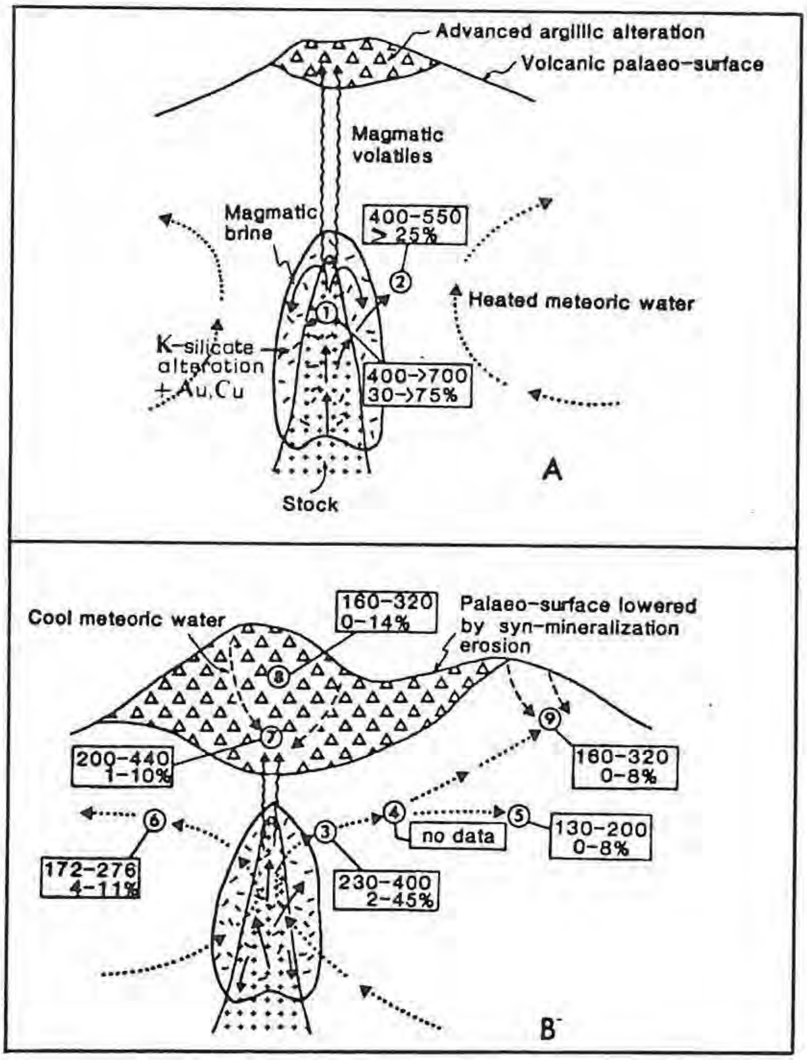


Fig. 4.10 *Genetic model for the acid-sulphate and adularia-sericite types of gold deposit related to an intrusion-centred gold district: (A) early stage dominated by magmatic-hydrothermal fluids; (B) late stage with "collapse" of the meteoritic system. Numbers in boxes represent homogenization temperatures (^{o}C) and salinities (wt % NaCl equivalent) (1) porphyry-type deposit (2) prograde skarn (3) retrograde skarn (4) carbonate replacement Au-base metals (5) carbonate replacement Au-As-Sb (6) non-carbonate disseminated, stockwork and replacement (7) deep acid-sulphate (8) acid-sulphate epithermal (9) adularia-sericite epithermal (after Sillitoe, 1989, 1991).*

Gold and copper in the porphyry system were probably introduced as chloride complexes and precipitated with the K-silicate assemblage when high temperature chloride- and sulphur-bearing fluids escaped into cooler wallrocks. Precipitation of metal sulphides from the chloride complexes in porphyry deposits resulted through hydrolysis of SO_2 to H_2S and H_2SO_4. HCl produced in this hydrolysis promotes additional phyllic alteration (Burnham and Ohmoto, 1980).

Magmatic-hydrothermal brines can also be responsible for the formation of skarn Cu-Au mineralization (Fig. 4.10).

Advanced argillic alteration and associated acid-sulphate-type gold deposits occurred in the upper part of porphyry copper systems (Lepanto enargite-gold deposit), where ascending magmatic volatiles undergo condensation and dissolution, resulting in low-pH meteoritic fluid. Some of the metals, including Au and Cu in these fluids could have been contributed as magmatic volatile complexes from the subjacent porphyry environment (Sillitoe, 1983). The temperatures (less than 320°C) and salinities (0-14 wt.% NaCl eq.) obtained from several acid-sulphate gold deposits indicate that meteoritic fluids, under boiling conditions, were responsible for gold deposition (Sillitoe, 1989). According to Stoffregen (1987), hydrothermal activity in the Summitville (Colorado) acid-sulphate type gold deposit, can be divided into two stages. An early period of acid leaching, under the influence of magmatic volatiles, resulted in the formation of "vuggy" silica and quartz-alunite alteration, at a pH of 2.0 and temperature near 250°C. At a later period, copper, gold and silver precipitated in a high sulphidic assemblage in a meteoritic water-dominated environment (Fig. 4.10A,B).

The vuggy silica rock is mineralogically similar to acid-leached zones formed in many modern hot springs. These areas contain quartz and anatase or other titanium oxides, along with opal, cristobalite and minor amounts of pyrite and/or iron oxide. They are formed by the surficial oxidation of H_2S , producing the H_2SO_4 necessary for intense leaching (Berger and Eimon, 1983). However, the vuggy silica and adjacent alunite zone at Summitville are believed to have been formed by the effect of magmatic gas (SO_2) ascending into a relatively near-surface environment and not from oxidization of H_2S above the water table. There is evidence to support this opinion, such as : alunitic alteration in the deposit is contemporaneous with the crystallization of the quartz latite of the area. This intrusion represents the source for the magmatic SO_2 necessary to produce acid-sulphate alteration in the hypogene setting. Hypogene acid formation, resulting from the effects of the SO_2, produces isotopically light sulphide and heavy sulphate, whereas oxidation of H_2S in a near-surface environment should not result in isotopic fractionation of sulphate and sulphide (Stoffregen, 1987).

This early magmatic-vapour dominated alteration stage caused intense leaching but did not contribute significant gold or other metals. The deposition of gold and sulphide is believed to result from a decreasing magmatic-vapour-dominated phase and its replacement by a system containing significant meteoritic water (Fig. 4.10B). The meteoritic fluids flowed laterally for several kilometres beyond the acid-sulphate type deposit to give rise to pyritic, advanced argillic alteration (Fig. 4.8).

These meteoritic fluids were also able to flow downward, causing overprinting of K-silicate alteration by advanced argillic alteration during late-stage activity in porphyry systems (Henley and Ellis, 1983) (Fig. 4.10 B).

Adularia-sericite-type gold deposits were also formed by cool (160-320 ^{o}C), dilute (0-8 wt.% NaCl eq.) fluids dominated by meteoritic water. However, these fluids were nearly neutral and more reduced than those responsible for acid-sulphate-type deposits (Heald et al., 1987). These deeply circulating meteoritic fluids that constitute the collapsed meteoritic cells (Fig. 4.10B), have transported gold as bisulphide complexes. Gold is believed to have precipitated with K-silicate alteration through destabilization of chloride complexes (introduced by magmatic fluid), partially remobilized in bisulphide form, under the influx of meteoritic hydrothermal fluids. Meteoritic hydrothermal fluid can circulate through fault systems at distances of 4-5 km from the porphyry copper deposit such as at the Andacollo deposit in Chile (Reyes, 1991), entering the more permeable and reactive dacite and andesite flow breccias where potassic alteration and deposition of gold-bearing pyrite and minor base metals take place. The Andacollo porphyry copper system may also be divided into two stages : an early mineralizing event responsible for deposition of gold in copper-gold porphyry stockwork, and the epithermal,distal adularia-sericite type of mineralization, precipitated during late events. Therefore the "two-stage"model, schematically represented in Fig. 4.10A/B can be accepted as a general genetic model for acid-sulphate and adularia-sericite types of epithermal gold deposit related to porphyry Cu(-Mo) systems.

4.2 EPITHERMAL GOLD DEPOSITS IN CONNECTION WITH GEOTHERMAL SYSTEMS RELATED TO VOLCANIC CENTRES AND CALDERAS

In the context of the recent, well-established understanding of geothermal systems and volcanic-hosted gold and silver deposits, Hayba et al. (1985) provided a classification of epithermal gold deposits into two types, based on alteration and mineral assemblages: acid-sulphate types such as Summitville and Red Mountain, Colorado and Goldfield and Paradise Peak, Nevada, and adularia-sericite(illite) types such as Creede, Colorado and Tonopah and Virginia City, Nevada.

The distribution and genesis of volcanic-hosted epithermal Au-Ag deposits is commonly tied to the plate-tectonic history of the region where they occur and the accompanying processes are volcanism and plutonism (Berger and Bonham, 1990). However, in areas where subvolcanic processes are absent, but where shallow plutonism or magmatic underplating of the crust is present, hydrothermal systems may occur with mixed meteoritic, connate and magmatic fluid sources, forming the so-called "Carlin-type" sedimentary-rock-hosted deposits.

Many diffferent classifications for the Carlin-type deposits have been proposed, such as mesothermal (Lindgren, 1933) and hot spring (Joralemon, 1951). Bagby and Berger (1985) accepted the epithermal classification for Carlin-type deposits, but included them along with a wider range of gold and silver replacement deposits termed "sediment-hosted, disseminated precious-metal deposits." There are attributes that indicate an environment of deposition different to most epithermal-type deposits, suggesting that Carlin-type deposits should be considered separately (Berger and Bagby, 1991). However, there are also similarities between Carlin-type deposits and epithermal systems such as: possible magmatic heat source that drives hydrothermal cells into the continental environment (Berger and Henley, 1989), high permeability structures through which meteoritic hydrothermal fluids circulate (Berger and Bagby, 1991), characteristics of the hydrothermal fluids, etc, indicating that this type of deposit should belong to the group of epithermal gold deposits.

Therefore two major groups of epithermal gold deposits connected with geothermal systems can be distinguished:

A) VOLCANIC-ROCK-HOSTED-DEPOSITS (acid-sulphate and adularia-sericite (illite) types) and
B) SEDIMENT-HOSTED and DISSEMINATED DEPOSITS (Carlin type)

The purpose of this section is to review the general geological characteristics of these deposits, whereas the geochemical characteristics of these and active geothermal systems will be discussed in chapter 5.0 of this dissertation. Detailed descriptions of the geology and geological environment of individual deposits are available elswhere in the recent literature (Bagby and Berger, 1985; proceedings of conferences such as Pacrim '87 [Aust. I.M.M., Gold Coast, Queensland, 1987]; Gold '88 [Geological Society Australia, Melbourne, 1988]; Brazil Gold '91;Hedenquist, White and Siddeley, 1990; Berger and Bagby, 1991).

4.2.1 Volcanic-rock-hosted epithermal gold deposits

Volcanic-rock-hosted epithermal gold deposits are widely distributed in tectonically active environments (Fig. 4.1-4.5). They may occur in a continuum of types ranging from shallow quartz-pyrite stockworks and breccias (the hot spring environment) to relatively deep veins and fissures (the bonanza environment). A schematic cross-section showing the form, hydrothermal alteration, ore and gangue mineralogy in a bonanza-type depositional environment which is transitional into a hot spring-type environment is shown in Fig. 4.11.

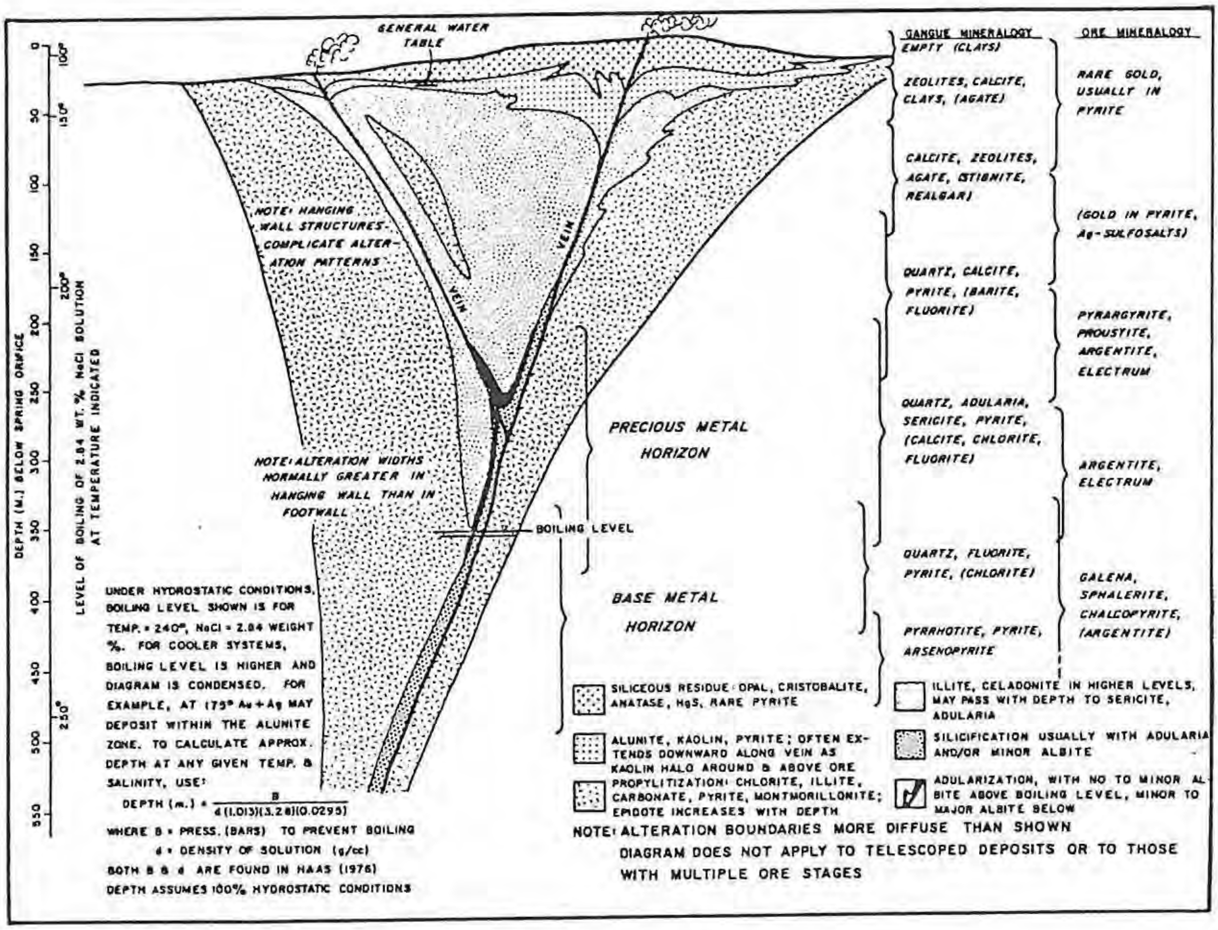


Fig. 4.11 *Hot spring and bonanza (open-vein) type epithermal deposits (after Buchanan, 1981).*

It also represents the "classic" classification of epithermal gold deposits based on depth relation. However, according to Hayba et al. (1985), Heald et al. (1987) and White and Hedenquist (1990), the most useful classification schemes for the explorationist are those based on the mineralogy, alteration and form of the deposit. Hydrothermal alteration and ore mineralogy are good indicators of fluid chemistry and temperature and in epithermal systems they are used to estimate these variables (White and Hedenquist, 1990). Hydrothermal alteration related to a near-neutral pH fluid system, which is analogous to those in active geothermal systems (Henley and Ellis, 1983), is described as the adularia-sericite type, whereas that related to acid pH systems is described as the acid-sulphate type (Heald et al., 1987). The characteristic features of the acid-sulphate (also called alunite-kaolinite) and adularia-sericite type epithermal gold deposits are shown in Table 4.2.

Characteristic	Adularia–sericite	Alunite-kaolinite
Structural setting	Structurally complex volcanic environments, commonly in calderas	Intrusive centers, 4 out of the 5 studied related to the margins of calderas
Size length : width ratio	Variable; some very large usually 3 : 1 or greater	Relatively small equidimensional
Host rocks	Silicic to intermediate and alkalic volcanics	Rhyodacite typical
Timing of ore and host	Similar ages of host and ore	Similar ages of host and ore (< 0.5 m.y.)
Mineralogy	Argentite, tetrahedrite, tennantite, native silver and gold, and base-metal sulphides. Chlorite common, selenides present, Mn gangue present, no bismuthinite	Enargite, pyrite, native gold, electrum, and base-metal sulphides. Chlorite rare, no selenides, Mn minerals rare, sometimes bismuthinite
Production data	Both gold- and silver-rich deposits variable base-metal production	Both gold- and silver-rich deposits, noteworthy Cu production
Alteration	Propylitic to argillic Supergene alunite, occasional kaolinite, abundant adularia	Advanced argillic to argillic (± sericitic) Extensive hypogene alunite, major hypogene kaolinite, no adularia
Temperature	100–300°C	200–300°C[1]
Salinity	0–13 wt.% Nacl equiv.[3]	1–24 wt.% NaCl equiv.[2]
Source of fluids	Dominantly meteoric	Dominantly meteoric, possibly significant magmatic component
Source of sulphide sulphur	Deep-seated magmatic or derived by leaching wallrocks deep in system	Deep-seated, probably magmatic
Source of lead	Precambrian or Phanerozoic rocks under volcanics	Volcanic rocks or magmatic fluids

Table 4.2 *Characteristics of adularia-sericite and acid-sulphate (kaolinite-alunite-type) epithermal gold deposits.*

a) Geotectonic settings of volcanic-rock-hosted epithermal deposits

Both the acid-sulphate and adularia-sericite types of epithermal deposit occur in similar tectonic settings associated with subduction zones at plate boundaries.The Western Pacific island arcs represent favourable tectonic settings for these types of deposits. The western Pacific terrain underwent a complex Cenozoic evolution resulting from the convergence of the major Eurasian and Indo-Australian plates with the Pacific plate. This evolution involved subduction of the Pacific basin lithosphere, dilation and closure by subduction of numerous marginal ocean basins, collision between island arcs and continents and intraoceanic and intracontinental faulting (Sillitoe, 1989). Subduction of the Pacific and marginal basin lithosphere was the most important process in the formation of volcanoplutonic arcs and associated epithermal gold deposits.

Many epithermal gold deposits were deposited in an extensional regime environment in the Basin and Range region of the western United States. Mesozoic-Cenozoic back-arc magmatism in the western United States consisted primarily of continental andesitic-dacitic suites. The back-arc volcanism, formed in the Eocene, was represented by andesite-rhyolite suites (Berger and Henley, 1989). An extensional regime in the Basin and Range region was accompanied by bimodal basalt-rhyolite volcanism. Therefore, epithermal gold deposits in these tectonic settings are related to alkalic, andesitic or felsic volcanism (Berger and Henley, 1989).

The Chilean continental margin is also known as a favourable tectonic setting, hosting numerous epithermal volcanic-hosted gold deposits. These deposits, which formed during Mesozoic-Cenozoid (Andean cycle) time spans, are related to intrusive-extrusive magmatism and constitute metallogenic belts along the western boundary of the South American plate. An early stage of the Andean orogen, when a magmatic back-arc basin setting had developed, is represented by calc-alkaline and local tholeiitic magmatism. Intermediate and K-rich calc-alkaline magmatism characterizes a later stage of the Andean evolution, when the magmatic arc accreted to the continental margin. Important north-south-trending structures and volcanic centres developed along this margin. Gold mineralization occurs throughout the whole Andean cycle (Camus, 1990).

Some subareal volcanic settings are not prospective for epithermal deposits, such as regions of continental flood basalts, whether tholeiitic or alkaline. This is probably because their magma chambers are too deep or too small, with narrow conduits, resulting in small near- surface heat anomalies (White and Hedenquist, 1990). Oceanic ridge settings have not appeared to be favourable for epithermal gold deposits, probably because they are typically submarine (White and Hedenquist, 1990).

b) Structural setting and control of volcanic-hosted epithermal gold deposits

Gold deposits, in general, display strong structural control, due to the permeability enhancement caused by fractures in the near-surface environment (Berger and Bonham, 1990). Many epithermal deposits are regionally associated with concentric and radial volcanic-related structures. A caldera, particularly its margin, is the most common structural setting for volcanic-hosted epithermal deposits. The importance of the caldera setting is that it can provide an excellent plumbing system later for the hydrothermal fluid necessary for the formation of adularia-sericite type deposits in particular (Heald et al., 1987).

Calderas may be subdivided into small (up to 10 km wide), roughly circular collapse structures commonly centred on the summit regions of individual volcanic edifices, and large (10-30 km wide) Valles-type calderas (Sillitoe and Bonham, 1984). Volcanic-hosted gold deposits can be related to Valles-type calderas, particularly to their ring fracture zones. Some of these deposits are integral products of caldera resurgence, especially of high silica alkali rhyolitie magmas, and are related to ring fracture domes, as at Lake City, Colorado, and Mc Dermitt, Oregon-Nevada (Sillitoe and Bonham, 1984). A classic example of an ore deposit related to a small summit caldera is provided by the Vatukoula epithermal gold district on Viti Levu, Fiji, where collapse was centered on the shoshonitic basalt Tavua shield volcano.

Geothermal systems centred on partly eroded andesitic stratovolcanoes are believed to be the active analog of porphypy copper systems (Henley and Ellis, 1983). The geothermal systems may underlie the flanks of simple or multiple andesitic cones in various states of preservation.

Acid-sulphate type deposits, apart from margins of calderas, are also related to silicic domes at the intersections of major faults (Hayba et al., 1985). The association of most of these deposits with intrusive centres, particularly with domal structures on the margins of calderas suggests that this environment is highly condusive for their genesis (Hayba et al., 1985).

Sillitoe and Bonham (1984) also recognized an association of gold mineralization with large maar-diatremes in volcanic fields, e.g. Cripple Creek (Colorado), Wau (New Guinea) and Baguio (Phillipines).

In volcanic terranes, combination of two or more volcanic landforms are common. For example, stratovolcanoes may possess central or adventive domes or maars, or may undergo late-stage collapse to form small summit calderas (Fig. 4.12); Valles-type calderas may partly destroy preexisting stratovolcanoes and during resurgence undergo addition of flow domes, either centrally located or localized along ring fracture zones (Sillitoe and Bonham, 1984).

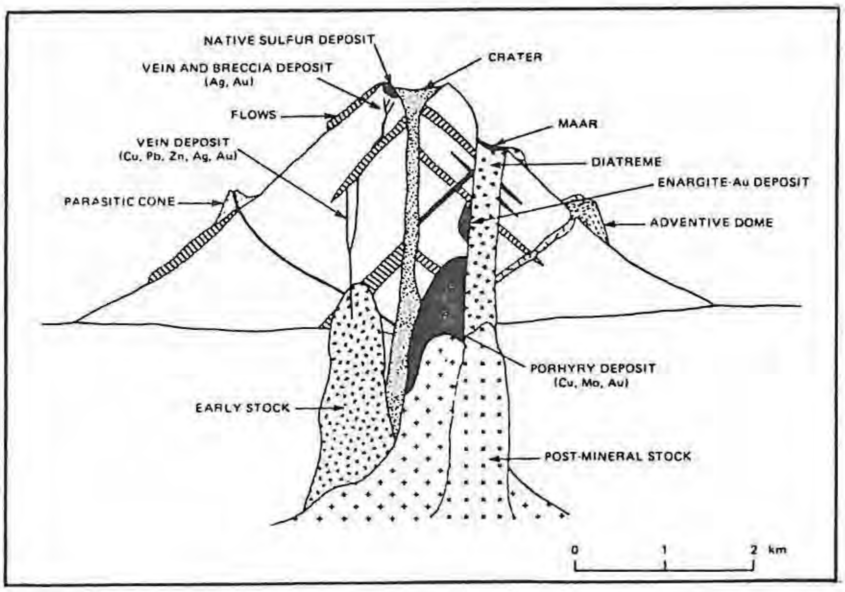


Fig. 4.12 *Idealized model of possible ore deposit types related to a stratovolcano (after Sillitoe and Bonham, 1984).*

In addition, deep-seated extensional fault zones and rifts play an important role in controlling the epithermal deposits as they guide the emplacement of magmatic heat (Berger and Henley, 1989). The volcanic eruptions may occur along normal faults, without the formation of any significant depressions or grabens (Berger and Bonham, 1990). However, graben commonly develops in many volcanic fields as volcanism progresses and epithermal gold deposits usually develop along the graben margin (Berger and Bonham, 1990).

c) Mineralogy of volcanic-rock-hosted epithermal deposits

The the major distinguishing features between adularia-sericite and acid-sulphate types of epithermal deposits are the different mineral assemblages and the presence or absence of specific minerals (Heald et al, 1987). Fig. 4.13 shows the mineralogical characteristics of some of the most significant adularia-sericite and acid-sulphate types of epithermal gold deposits.

The presence or absence of the assemblage enargite + pyrite +/- covellite and adularia, is one of the major characteristics for classification of these deposits.

Acid-sulphate type deposits are characterized by the occurence of the vein mineral assemblage enargite + pyrite +/- covellite and an advanced argillic assemblage including hypogene alunite and a major abundance of kaolinite (Heald et al., 1987). The mineralogy of the adularia-sericite type deposits is represented by the vein adularia and sericite and by the absence of both hypogene alunite and the mineral assemblage enargite + pyrite +/- covellite (Hayba et al., 1985). Adularia is absent in acid-sulphate type deposits and chlorite is rarely present, whereas chlorite is commonly present in adularia-sericite type deposits.

Selenides, rhodochrosite and fluorite are more common in adularia-sericite deposits than in acid-sulphate types, whereas bismuthinite has been identified only in acid-sulphate type deposits (Heald et al., 1987). The occurence of barite is recognized in almost all acid-sulphate-type deposits, whereas it is less common in adularia-sericite deposits (Heald et al., 1987).

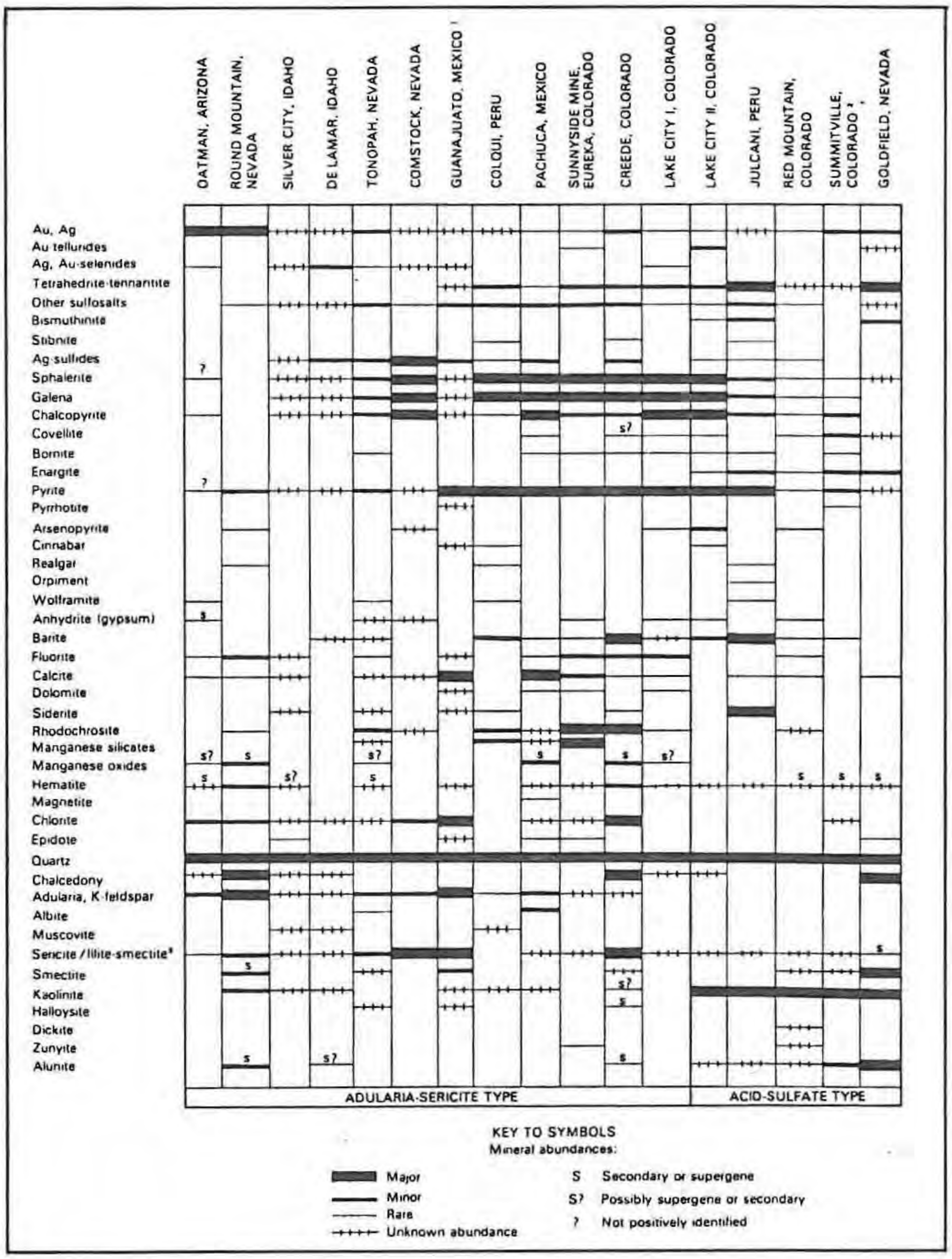


Fig. 4.13 *Schematic representation of the mineralogy of volcanic-hosted epithermal gold deposits (after Heald et al., 1987).*

Precious metals in these deposits are present mainly as native gold and silver, electrum, silver sulphides and silver sulpho-salts. Most of the adularia-sericite type deposits have a high silver-to-gold ratio. However, districts such as Round Mountain (Nevada) and Oatman (Arizona) have low silver-to-gold ratios and precious metals are present mainly as native gold, silver and electrum. The silver-to-gold ratio fluctuates for acid-sulphate deposits. Summitville and Goldfield have low silver-to-gold ratios (2:1) whereas Red Mountain has high ratios (10:1) (Hayba et al., 1985).

Base-metal production is usually lower for the gold-rich deposits and higher for the silver-rich deposits. In the acid-sulphate deposits, copper constitutes the major proportion of the base-metal production, whereas among the adularia-sericite type deposits, its production is highly variable and in some of them it has not even been reported (Heald et al., 1987).

Mineralization occurs in stratiform, disseminated stockwork, cross-cutting and breccia form. Steep veins or vein systems up to several kilometres in length are the most common form of adularia-sericite type deposits.

d) Alteration assemblages and the spatial zonation

Alteration assemblages are another key feature in distinguishing the two types of volcanic-rock-hosted epithermal deposits.

ADULARIA-SERICITE type deposits are characterized by the dominance of sericitic alteration, consisting of a mica-type mineral (e.g. illite) + quartz + pyrite (Fig. 4.14), and often border silicified zones near veins (Hayba et al., 1987).

Near the veins, K-feldspar and chlorite can often be disseminated in the wall rock. The sericitic alteration grades outward into intermediate argillic and propylitic zones (Heald et al., 1987). In deposits where erosion has not occured for more than a few tens of metres below the paleosurface, an advanced argillic alteration assemblage may be preserved (Fig. 4.14). Auriferous hot spring sinter may occur on the top of the adularia-sericite system,as in the Kabang system in the western Pacific (Licence et al., 1987), at Mc Laughlin (California) and Borealis and Sulphur in Nevada (Silberman and Berger, 1985).

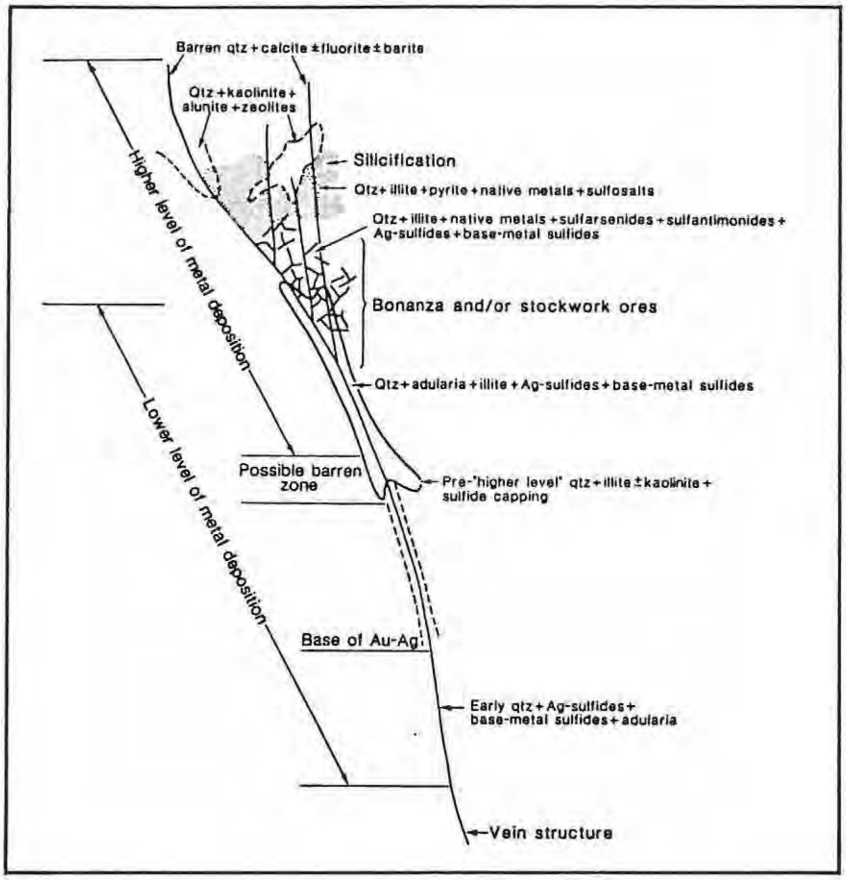


Fig. 4.14 *Schematic cross-section of an adularia-sericite-type deposit ("bonanza"-model of Berger and Eimon, 1983, in Silberman and Berger, 1985).*

These hot spring deposits are often associated with hydrothermal eruption breccias (Fig. 4.15) and laucostrine sediments (Henley, 1991). Alteration assemblages are represented by high silicification with adularia and/or albite, variable calcite or dolomite. This grades outwards to argillic and/or zeolitic alteration, then to propylitic assemblages (Fig. 4.15).

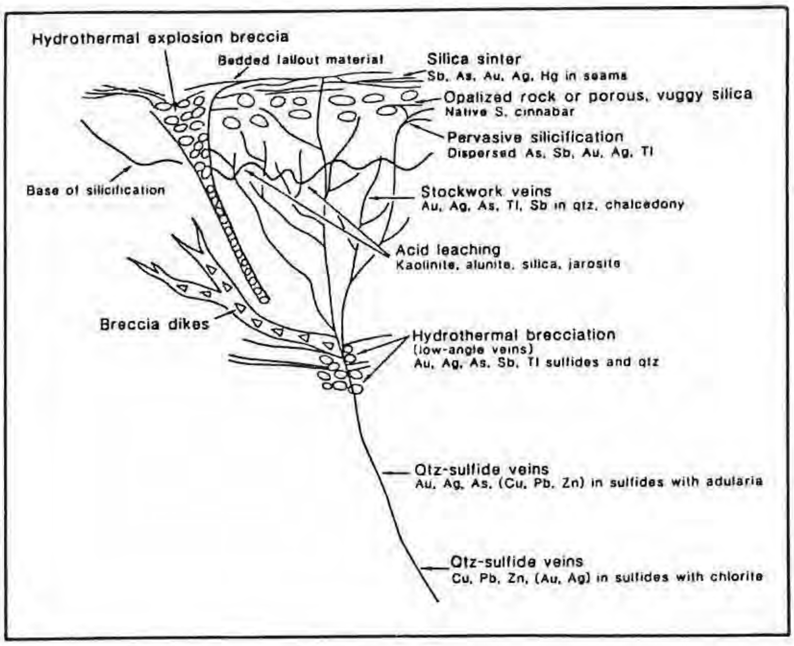


Fig. 4.15 *Schematic cross-section of adularia-sericite-type deposits ("hot spring-style") (after Hayba et al., 1987).*

Mineralization is due to repeated, episodic and explosive stock-work veining and hydrothermal brecciation and is represented by native gold and electrum with associated pyrite, marcasite and silver sulpho-salts (Silberman and Berger, 1985). According to Hayba et al. (1985) such hot spring mineralization forms in the surficial parts of hydrothermal systems similar to those which deposited the Creede (Colorado) ores.

The relationship of fluids to mineralization has been studied for the past 30 years in the adularia-sericite type Creede district. The deposit occurs above the projection of the Rio Grande rift fault zone in the San Juan volcanic field (Fig. 4.16).

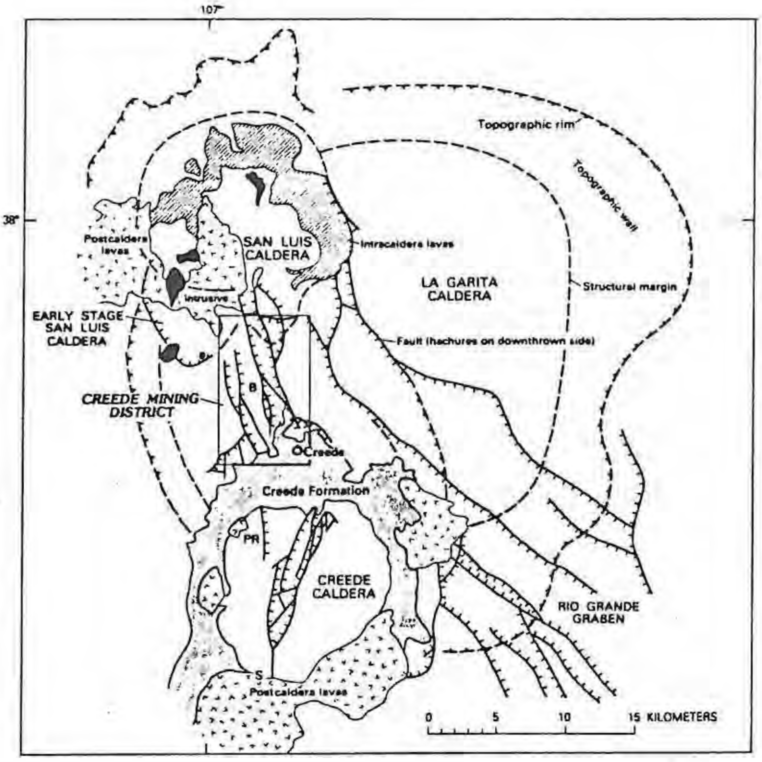


Fig. 4.16 *Generalized geology of the Creede and San Luis calderas in relation to remnants of the Bachelor (B) and La Jarita calderas and to the Creede mining district (shown in box) (after Hayba et al., 1985).*

The base- and precious-metal mineralization fills open fractures (Hayba et al., 1985). The vein mineralization was deposited from fluids flowing along Creede graben faults which originated along the structural margin of the La Garita caldera (Fig. 4.16) and the keystone graben of the resurgent Bachelor caldera (Berger and Henley, 1990).

Mineralization in the Creede veins is strongly zoned from an association in the north: chlorite + hematite + quartz + adularia + sphalerite + galena + chalcopyrite + pyrite +/- fluorite and tetrahedrite to a barite + rhodochrosite + quartz + adularia + galena + sphalerite + fluorite + tetrahedrite + silver sulpho-salt + native silver association to the south (Hayba et al., 1985).

Wallrock alteration is represented by sericite alteration (illite/smectite clays) on the top of the orebodies. The alteration is strongly fracture-controlled and the intensity of argillization decreases away from the veins. The most intense alteration forms a clay-cap at the upper limit of mining. In the upper parts of the vein systems, wallrock adjacent to the veins is strongly silicified. The sericitic alteration often borders the silicified zones.

Below the zone of intense sericitic alteration, the wallrocks are generally fresh, but occasionally show enrichment in potassium which originated from hydrothermal alteration that occured 2 Ma prior to ore deposition and does not show an association with any earlier period of mineralization (Hayba et al., 1985). Some of the general characteristics of the geochemical environment of the Creede ore-forming system are shown in Table 4.3, but will be discussed further in the following chapter of this dissertation.

Parameter	Range observed	Reference environment	Source of information
Temperature	190 - 285°C	250°C	Fluid inclusions
	40 - 50 bars	50 bars	Evidence of boiling in fluid inclusions
Depth	450 - 600 m	500 m	Estimated from pressure and geologic reconstr.
Salinity	4 - 12 wt%	6 wt%	Fluid inclusions
Na:K	7.4 - 9.9	9	Analyses of fluid inclusions
Total S	0.018-0.30 molal	0.02 molal^2	Analyses of fluid inclusions
pH		5.4	Calculated

Table 4.3 *General environmental parameters for the Creede district, Colorado (after Hayba et al., 1987).*

ACID-SULPHATE type deposits are characterized by the association of advanced argillic alteration with ore (Fig. 4.17). Kaolinite and alunite of the advanced argillic assemblage occur close to the vein and are often coextensive with silicification (Hayba et al., 1985). Sometimes sericitic alteration surrounds the advanced argillic alteration and is intermixed with argillic alteration which occurs farther from the vein (Fig. 4.17).

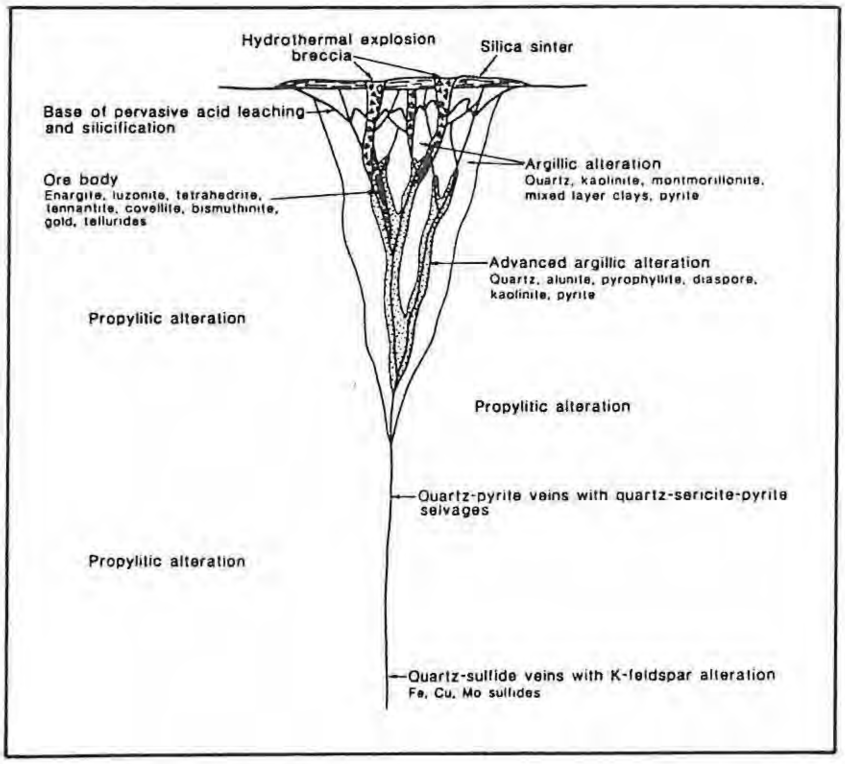


Fig. 4.17 *Schematic cross-section of the acid-sulphate-type deposit (after Silberman and Berger, 1985).*

Argillic alteration is often zoned itself, with kaolinite near the vein and smectite farther away (Heald et al., 1978). The outermost alteration zone is represented by propylitic alteration. This style of deposit is exemplified by Goldfield (Nevada), Summitville (Colorado) and Iwato and Kasuga (Japan). Spectacularly high grade ore (225g/t Au, 104g/t Ag and 2.4% Cu) is observed in El Indio (Chile) (Henley, 1991).

The Summitville mining district is the best documented acid-sulphate-type epithermal system and some of its most important geologic characteristics will be reviewed below. Certain of the geochemical characteristics and the hydrothermal activity have already been mentioned in section 4.1.4., when thegenetic connection between epithermal and porphyry systems was discussed. More information about the genetic aspects of acid-sulphate type deposits will be given in chapter 5.0.

The Summitville mining district is located on the edge of the Platoro caldera and the Summitville caldera in the southeastern San Juan Caldera complex, Colorado (Fig. 4.18).

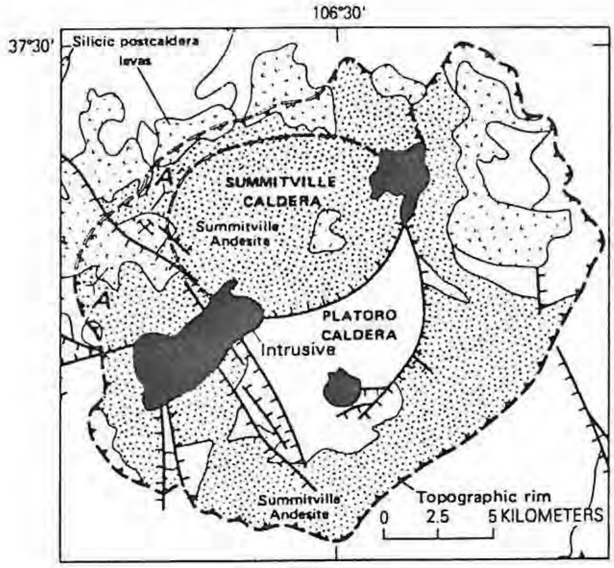


Fig. 4.18 *Generalized geology of the Summitville caldera. Location of the Summitville mining district is shown by pick-and-hammer (after Hayba et al., 1985). A - A1 marks location of cross-section shown in Fig. 4.19.*

Mineralization in the Summitville mining district was formed in association with a quartz-latite porphyry, emplaced 23.4 Ma (Berger and Bonham, 1990), which is bordered by older andesite (on the north, east and south) and rhyodacite (on the west).

The mineralization is of the alunite-kaolinite +/-pyrophyllite type (Hayba et al., 1985) and occurs along the South Mountain fault zone above the western ring-fracture zone of the Summitville caldera (Fig. 4.18, Fig. 4.19). The ore bodies are localized along the south western margin of the porphyritic core of the dome. The ore occurs in a series of irregular pipes and vein-like masses of quartz and alunite (Hayba et al., 1985).

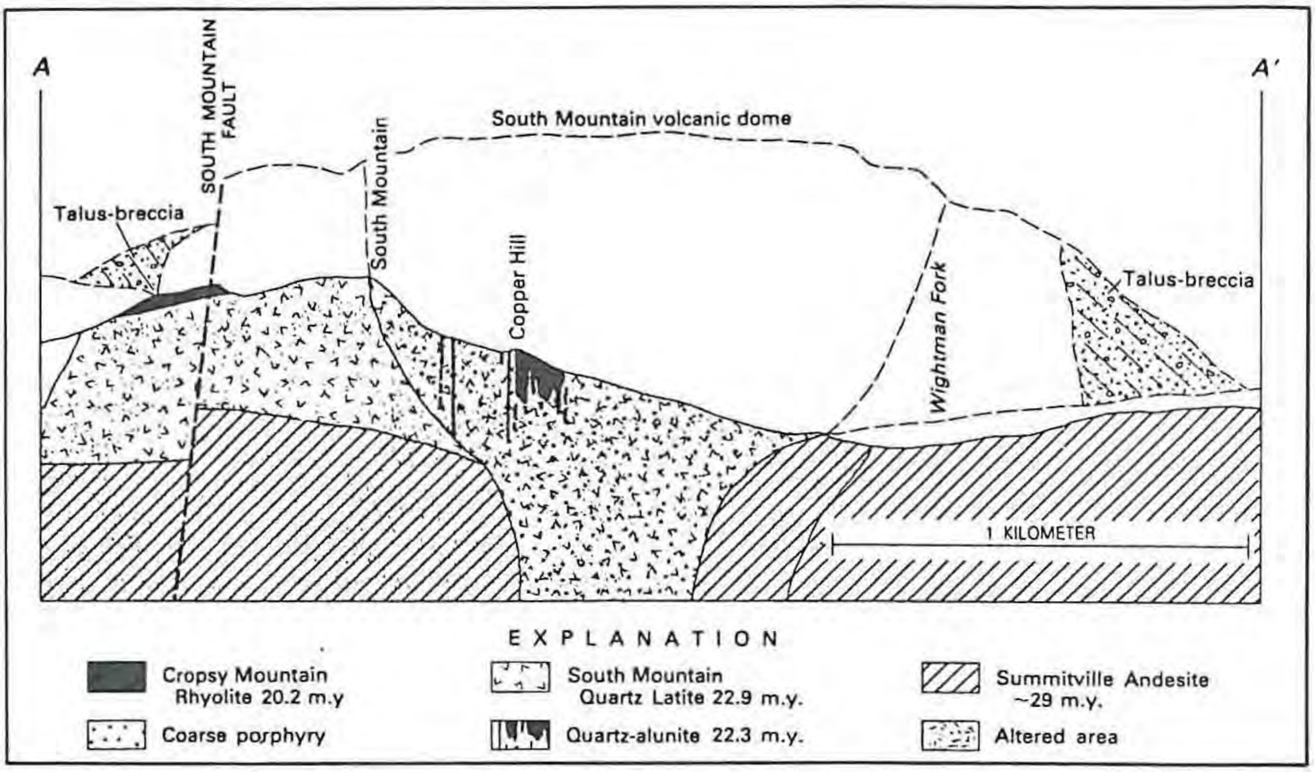


Fig. 4.19 *Geologic cross-section of the South Mountain volcanic dome. Fault is shown with heavy line; contacts with thin line; both are dashed where appropriate (after Hayba et al., 1987).*

The earliest stage of mineralization consisted of advanced argillic alteration with a core that is represented by vuggy, leached silica and a halo of quartz-alunite +/- pyrophyllite followed outwardly by envelopes of quartz-kaolinite and illite-kaolinite which then grades into montmorillonite (Stoffregen, 1987) (Fig. 4.20). Most of the ore occurs in the vuggy silica alteration zone which is quite extensive at surface, but at a depth of 300m it becomes much more restricted (Fig. 4.21).

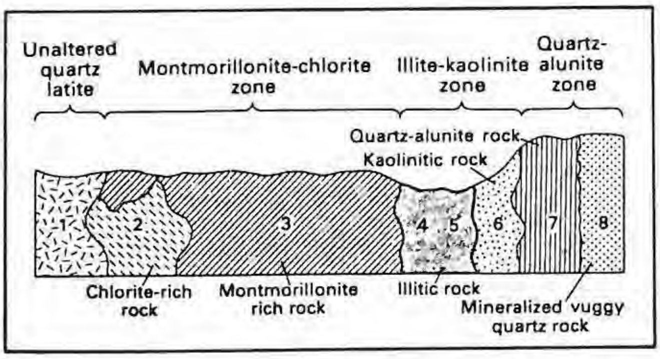


Fig. 4.20 *Hydrothermal alteration in the Summitville deposit (after Hayba et al., 1985).*

The upper part of the deposit contains economically significant precious-metal mineralization and the mineral assemblage is represented by covellite, luzonite, enargite, pyrite, marcarite, chalcopyrite, sulphur and trace amounts of sphalerite and native gold. The character of mineralization changes with depth and a chalcopyrite-tennantite assemblage predominates (Fig. 4.21). According to Stoffregen (1987), this mineralization is described as main stage mineralization. Late stage is represented by the assemblage barite, jarosite, goethite and gold, found in the uppermost levels of the deposit.

In the formation of the Summitville acid-sulphate type deposit, two stages of hydrothermal activity were recognised by Stoffregen (1982). An early hydrothermal fluid, derived from the mixing of an SO_2-rich magmatic vapor (see section 4.1.4) and a convecting meteoritic water system were responsible for intense leaching. As the magmatic plume collapsed at a later stage, the convective meteoritic water system interacted with the advanced argillic alteration and deposited gold and silver in association with enargite, covellite and luzonite (Stoffregen, 1987).

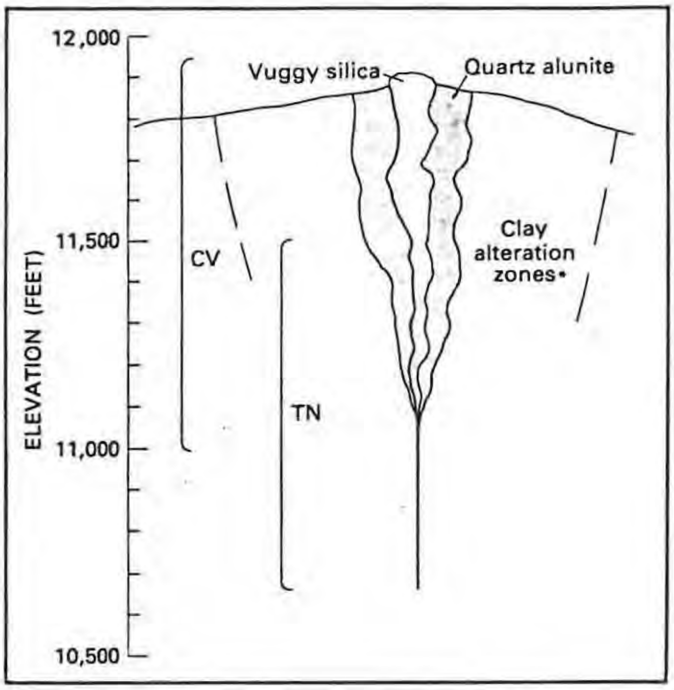


Fig. 4.21 *Schematic cross-section of the alteration patterns and mineral zonation of the Summitville deposit. The clay alteration zones refer to zones 3 - 6 in Fig 4.20. CV = covellite, luzonite, enargite, pyrite, marcasite, chalcopyrite, trace sphalerite, sulphyr and gold assemblage; TN = chalcopyrite, tennantite, pyrite, and minor sphalerite, trace galena (after Hayba et al., 1985)*

4.2.2 Sediment hosted, disseminated gold deposits

Sediment-hosted, disseminated precious-metal deposits are formed in carbonaceous, silty dolomites and limestones or calcareous siltstone and claystones (Bagby and Berger, 1985), which are intensely fractured and silicified in the mineralizing zones. Gold mineralization is disseminatead in the host rocks and often associated with trace elements As, Ba, Hg, Sb and Tl. Deposits of this type are referred to as "Carlin-type" deposits on the basis of their similarities to the Carlin deposit in Nevada, which has been extensively studied since its discovery in 1962 (Berger and Henley, 1989).

The majority of the Carlin-type deposits occur in thewestern United States, predominating in Nevada (Fig. 4.22).

Fig. 4.22 *Location of the Carlin-type deposits in western United States (after Berger and Henley, 1989).*

They vary in size and grade from low-grade and high tonnage to high-grade and low tonnage, with the most significant of these deposits in terms of grade being Carlin and Jerritt Canyon with 42 Mt, about 7 g/t

Au, and 2-20 Mt, about 7 g/t Au, respectively (Mason, 1991b). Carlin-type deposits have recently been discovered in southern China (Fig. 4.23) in a terrane which appears to be analogous to parts of western North America (Berger and Bagby, 1991).

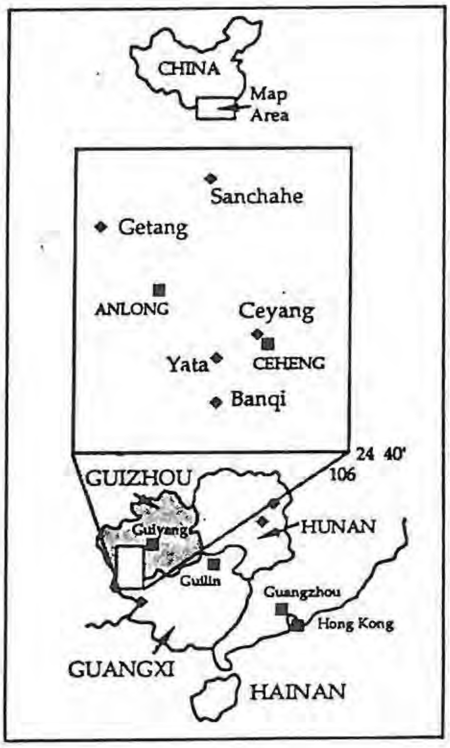


Fig. 4.23 *Location of the Carlin-type deposits in south-eastern China (after Cunningham et al., 1988).*

Sediment-hosted, disseminated precious-metal deposits vary considerably in their geological characteristics. Some of them are characterized by fine-grained disseminated gold ore that is not easily distinguishable from unaltered rock (e.g. the Carlin deposit), whereas gold in other deposits is associated with intense silicification (jasperoids) and quartz veining (e.g. Preble, Nevada). They can be gold-rich (e.g. Carlin and Preble) or silver-rich (e.g. Taylor deposit, Nevada). They also differ individually in host lithology, ore types and shapes (Bagby and Berger, 1985).

Despite all these differences, the presence of a large number of common features justifies the consideration of all Carlin-type precious-metal deposits as a single category.

Some geological and geochemical characteristics of the most significant Carlin-type deposits in the western United States are summarized in Table 4.4.

a) Geotectonic setting, magmatism and nature of host-rocks

The Carlin-type deposits of the western United States and southern China occur in Paleozoic and Mesozoic overthrust belts related to continental margin tectonics (Berger and Bagby, 1991). The belts in the western United States were formed during three major orogenic events, beginning with the Antler event, which occurred in the Late Devonian to Early Mississippian. Allochthonous, siliciclastic rocks were emplaced along the Roberts Mountains thrust during the Antler Orogeny, and represent the earliest allochthonous terranes accreted to the western margin of the United States (Bagby and Berger, 1985). The Carlin-type deposits within the Carlin mineral belt (Rain, Maggie Creek, Gold Quarry, Carlin, Bullion Monarch, Blue Star, Gold Strike, Bootstrap and Dee) occur in the Roberts Mountains allochthon (Fig. 4.24) which in turn overlies a Precambrian crystalline basement. Deposits within the Cortez trend (Gold Acres, Cortez, Horse Canyon and Tonkin Springs) occur within windows of the Roberts Mountains allochthon.

Igneous rocks occur in many of the Carlin-type districts in the western United States and are usually of Cretaceous or early to mid-Tertiary age (Table 4.4). Some may be of Permian/Triassic age. Igneous rocks are represented by intermediate to silica dikes, plugs, domes and stocks. Granodioritic stocks, andgranodiorite to granite dikes are found in the Getchell, Gold Acres, Goldstrike and some other deposits (Berger and Bagby, 1991). Genetic relationships between these rocks and gold mineralization is not well documented. However, Berger and Henley (1989) and Berger and Bagby (1991) propose that there is a strong link between magmatism (probably as a heat source) and metal source and tectonism in the formation of these deposits. They can occur very close to their igneous source, as at Getchell, or more distal, as at Jerritt Canyon.

Deposit	Host rocks			Igneous rocks		Structure		
	Formation	Age	Lithology	Age	Composition and occurrence	Faults	Folds	Mineralization age
Carlin	Roberts Mountains Formation	Silurian to Early Devonian	Laminated, silty to sandy, carbonaceous, dolomitic limestone	Cretaceous 130 m.y.	Altered quartz diorite and diorite dikes	Devonian-Mississippian thrust fault (Roberts Mountains thrust), high angle faults trending E, N, and NE and NW. Faults are pre-mineral with some post-mineral movement.	NW directed folds (antiforms and synforms). Major NW-trending antiform in the district. Age of folding is Mesozoic.	No direct date of mineralization. Post-mineral rhyolite lavas and domes dated at 14 m.y. provide upper age constraint and altered Cretaceous dikes (130 m.y.) provide lower age constraint.
Cortez	Roberts Mountains Formation	Silurian to Early Devonian	Laminated, silty, argillaceous, carbonaceous, pyrite-bearing limestone with dolomite	Oligocene 33-35 m.y. Jurassic 150 m.y.	Altered dikes and sills of quartz latite; Mill Canyon granodiorite stock	N, NW, and EW trending, high angle, normal faults. Roberts Mountains thrust surrounds district.	Drag folds associated with faults sympathetic to Roberts Mountains thrust. NW directed regional folds.	Altered and mineralized 33-35 m.y. old quartz latite dikes place a lower age constraint.
Getchell	Preble Formation Comus Formation	Cambrian Ordovician	Phyllitic shale with interbedded limestone	Cretaceous	Osgood Mountains pluton; Granodioritic with associated intermediate porphyritic dikes. All are altered and in part mineralized.	N. trending Getchell fault zone includes several strands. Inception of fault Late Cretaceous.	Fold axis plunges 45° NE. On southern limb sediments strike N and dip SE.	Sericite in mineralized granodiorite dated by K-Ar between 87-92 m.y. This is inferred age of gold mineralization.
Gold Acres	Roberts Mountains Formation Valmy Formation Wenban Formation	Silurian to Early Devonian Ordovician Devonian	Carbonates, argillites, and siltstones: All mineralized rocks occur as fault blocks low in the upper plate of the Roberts Mountains thrust.	Cretaceous Tertiary(?)	Altered and mineralized dikes of intermediate composition; Quartz latite sills that are altered.	N, NE faults dip steeply west. Roberts Mountains thrust.	NW directed antiforms and synforms of presumed Mesozoic age.	Not directly dated. Altered Tertiary(?) sill places a lower age constraint.
Jerritt Canyon	Hanson Creek	Ordovician	Carbonaceous, shaly limestones with chert dolomite, and bioclastic limestones	Mid-Tertiary(?)	Small dikes and plugs of diorite and a small rhyodacite flow 2.4 km SW and 3 km NE of mineralization, respectively	Normal faults striking E-W, N-S N. 20-30 E. Roberts Mountains thrust fault. NS faults post-mineral and high angle.	Regional E-W trending folds.	Unknown
Preble	Preble Formation	Cambrian	Phyllitic shale and interbedded limestone	Cretaceous(?)	Altered dike of intermediate composition	N and NE to EW trending high angle faults. Cretaceous(?).	NE directed antiforms and synforms.	Not dated but presumed to be the same as Getchell (87-92 m.y.).
Taylor	Guilmette Formation	Devonian	Limestone and shaly limestone	Tertiary (35 m.y.)	Rhyolitic dikes altered but not mineralized	N, NW, NE trending high angle faults of Tertiary age.	Possible S plunging antiform.	The age constraints are provided by post mineralization, mid-Tertiary dikes, and early Tertiary, pre-mineralization faults.

Table 4.4 *Characteristics of some of the Carlin-type gold deposits*

(modified after Bagby and Berger, 1985).

	Alteration		Ore bodies	Ore bodies		
Deposit	Hypogene	Supergene	Form	Mineralogy	Gold or silver site	Veins
Carlin	Decarbonization, silicification, calcification, carbon remobilization, acid leaching with oxidation(?)	Oxidation, calcite remobilization, clay formation	Pods localized along high angle faults and extending into sediments	Oxidized: metallic gold, goethite, illite, kaolinite, barite, anhydrite, alunite, dolomite, calcite, quaartz, schuetteite, cinnabar, arsenolite, scorodite, stibiconite, avicennite, and various lead, zinc, and copper oxides. Unoxidized: quartz, calcite, dolomite, illite, pyrite, realgar, orpiment, stibnite, cinnabar, base-metal sulfides, and rare Tl-As-Sb-S minerals.	Oxidized: metallic gold. Unoxidized: gold with mercury, arsenic, tin, and thallium form thin films on pyrite. Gold is locally associated with carbonaceous matter.	Quartz-calcite-orpiment, barite-galena, quartz-pyrobitumen, quartz-pyrite, and calcite veinlets. Calcite zoned away from deposit in general with quartz veins occurring within main ore deposit area.
Cortez	Silicification, acid leaching and oxidation(?), decalcification, and dedolomitization	Deep oxidation resulting in some redistribution of gold	Elongated zones paralleling faults and dikes and notably localized in breccia zones associated with folds	Oxidized: quartz, clays, iron oxides, metallic gold, and calicite. Unoxidized: quartz, illite, dolomite, calcite, pyrite.	Oxidized: clusters of particles between silt grains, metallic gold grains in quartz veinlets, grains in limonite pseudomorphs after pyrite. Unoxidized: associated with As and pyrite.	Quartz-pyrite veinlets and post-mineralization calcite veinlets. Calcite veinlets zoned away from ore zone may have formed contemporaneous with decalcification within the ore zone.
Getchell	Decarbonatization with silicification and argillization; early calc-silicate skarnification	Minor oxidation	Sheet-like zones localized along strands of Getchell fault and pods in fold hinges	Oxidized: metallic gold(?) Unoxidized: pyrite, pyrrhotite, arsenopyrite, marcasite, stibnite, orpiment, realgar, ilsemannite, cinnabar, magnetite, and metallic silver.	Oxidized: no published data. Unoxidized: metallic gold encapsulated by quartz. Gold also associated with pyrite, arsenopyrite, carbonaceous matter, and magnetite. Silver associated with sulfides.	Calcite veins in limestone and quartz veins in phyllitic shale. Stockwork quartz veins cutting igneous rocks are in turn cut by calcite-dolomite veins.
Gold Acres	Silicification, early contact metamorphism	Deep oxidation	Tabular, dipping to SW parallel to thrust faults	Oxidized: quartz, kaolinite, iron oxides, sericite, jarosite, gypsum, hexahydrite, dolomite, and calcite. Unoxidized: no published data available.	Oxidized: gold associated with iron oxide-clay fracture coatings.	Unreported.
Jerritt Canyon	Decalcification, silicification, remobilization of organic matter	Oxidization and leaching	Conformable to bedding as pods near faults	Unoxidized: pyrite, realgar, orpiment, arsenopyrite, cinnabar, and organic matter. Stibnite, barite, and quartz occur near ore bodies.	Gold 1-4 micron; mineral association inknown.	Quartz, stibnite, and barite veins occur in jasperoid near ore bodies. Calcite and arsenic minerals occur as veins in unoxidized ore.
Preble	Silicification, argillization, dolomitization	Oxidation to about 200 ft	Ore body localized along shear zone	Oxidized: iron oxides, minor pyrite, and metallic gold. Unoxidized: pyrite, carbonaceous matter, and minor arsenopyrite, marcasite, chalcopyrite, and sphalerite.	Gold encapsulated in quartz and associated with pyrite. Metallic, micron-size gold in oxidized ore.	Quartz, dolomite, jasperoid breccia, and calcite veins.
Taylor	Silicification, late argillization	Significant oxidation resulting in silver enrichment blanket	Tabular silicified bodies on top of Guilmette Formation; localized by faults	Oxidized: iron oxides, with remnant stibnite, sphalerite, tetrahedrite, chalcopyrite, galena, pyragyrite. Unoxidized: none reported.	Metallic silver.	Quartz and late stage calcite veinlets.

Table 4.4 *Characteristics of some of the Carlin-type gold deposits (modified after Bagby and Berger, 1985).*

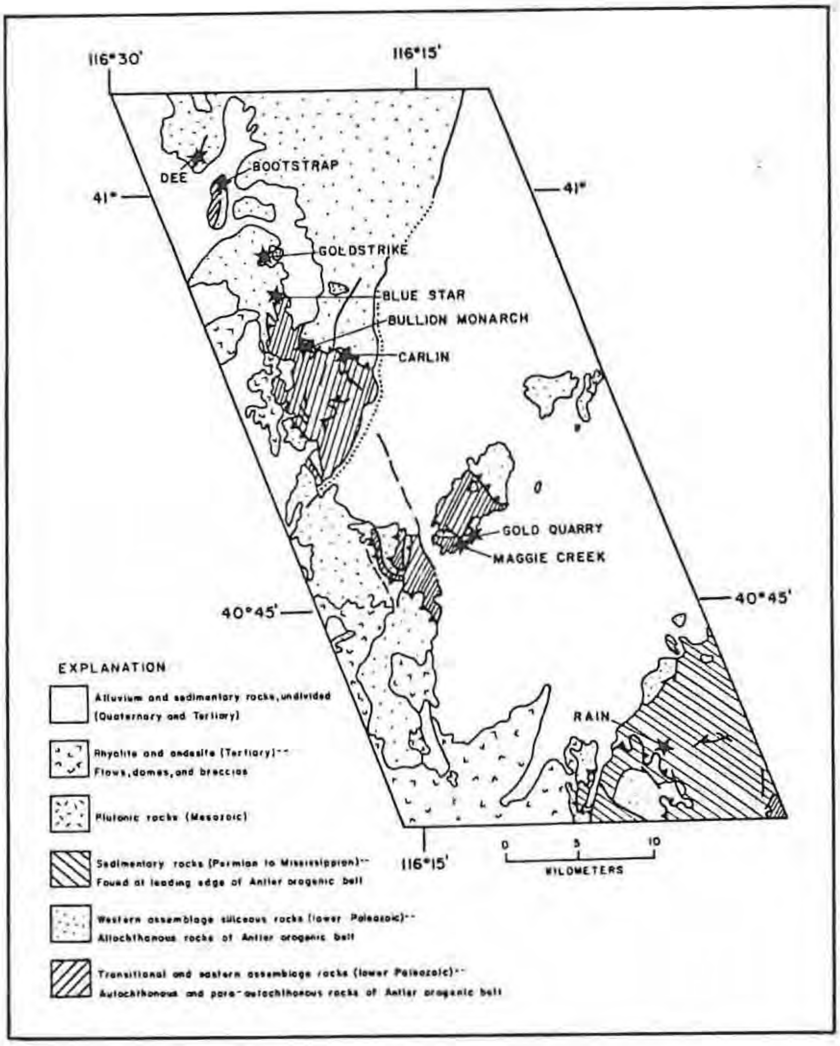


Fig. 4.24 *Regional geology of the Carlin trend (after Bagby and Berger, 1985).*

The host rocks of Carlin-type deposits are represented by a number of different lithologies (Table 4.4). The most favourable rocks are bedded silty or shaly, carbonaceous limestones and dolomites, calcareous, carbonaceous shales and siltstones (Bagby and Berger, 1985). They vary in age from Cambrian to Triassic.

In the Carlin mine area, the host Roberts Mountains Formation consists of dark-grey, fine-grained, laminated, silty, carbonaceous dolomitic limestone. It consists of calcite, dolomite, quartz and lesser amounts of: illite, kaolinite, chlorite, epidote, hornblende, plagioclase, K-feldspar and tourmaline. The total organic carbon content varies from about 0.4 to 1.6% and occurs as patches filling open spaces in limestone, disseminated in siltstone, chert and dolomitic limestone and lenticular seams along bedding planes in dolomitic siltstone (in Berger and Bagby, 1991).

In China, theCarlin-type deposits occur in a Cambrian stratigraphic succession, consisting of carbonaceous dolomite and limestone, with intercalated black shale, and in laminated argillaceous limestone with interbedded shale and sandstone (Cunningham et al., 1988). This sequence also overlies Proterozoic crystalline rocks as does the Robert Mountain Formation in the western United States.

b) Structural setting of the Carlin-type deposits

The presence of high-angle normal and strike-slip faults is important for the formation of Carlin-type deposits, as they represent the channels for hydrothermal fluid flow. Different structures in some of the Carlin-type deposits in the western United States are summarised in Table 4.4.

Intersections between these high-angle faults are of particular importance in controlling mineralization. They increase the host-rock permeability and form breccias, which often focus ore deposition. Ore bodies at the Taylor deposit were formed in and near fault breccia zones (Bagby and Berger, 1985).

Antiformal structures are a significant control on the locus of mineralization, as many deposits have been recognised to occur near the crest of regional antiforms (e.g. Carlin deposit).

Although thrust faults are not believed to be the major controlling structure for ore deposition, imbrication and folding associated with thrusting at Carlin and Jerritt Canyon have opened up the bedding planes of the bedded silty carbonates and micritic limestones and thus prepared favourable environments for circulation of hydrothermal fluids (Mason, 1991b).

c) Alteration and mineralization

The primary alteration associated with gold mineralization in Carlin-type deposits is characterized by dissolution of carbonate minerals and the precipitation of silica (Table 4.4). In many deposits, jasperoid replaced carbonate along or near faults (e.g. the Taylor, Dee andCortez deposits). Replacement can occur during the pre-gold-ore stage, as well as during the gold deposition stage (Berger and Bagby, 1991), which is characterized by the formation of quartz veins and silicified breccias that cross-cut the earlier, pre-ore silicification. Post-ore veins are usually represented by calcite which can be associated with barite, realgar, stibnite, cinnabar and complex thallium minerals. Anhydrite and kaolinite veins can also form during this post-ore stage (Radtke et al., 1980). The variable paragenesis and textures of silica result from deposition during different stages of alteration and can therefore serve as an indication of pre-ore, ore-stage and post-ore events during the life of a hydrothermal system (Berger and Bagby, 1991). They are summarized in Table 4.5.

A general alteration zonation has been recognized in some Carlin-type deposits. At Carlin, silicification is followed outwardly by quartz+ dictite+ kaolinite, then by quartz+ dolomite+ illite/K-mica and through quartz+ calcite+ dolomite+ illite+ K-feldspar into unaltered rock (Kuehn and Rose, 1987). The highest gold grades are associated with quartz+ dolomite+ illite/K-mica assemblage.

Hofstra et al. (1988) recognized alteration zonation in the Jerritt Canyon deposit from the ore outwards as: quartz, quartz+ dolomite, dolomite+ calcite, and calcite. The ore is associated with intermediate zones of carbonate dissolution and silicification.

Paragenesis	Deposit							
	Mercur		Pinson		Jerritt Canyon		Carlin	
	Replacement and veining	Textures of silica	Replacement and veining	Texture of silica	Replacement and veining	Texture of silica	Replacement and veining	Texture of silica
Pre-gold-ore stage	*Silver chert mineralization* Jasperoid; silica-replacement of silty ls and cal siltst preserving some sedimentary textures	Xenomorphic and jigsaw-puzzle texture	Jasperoid; silica-replacement of silty dolomite preserving some sedimentary textures	Xenomorphic and jigsaw-puzzle texture	Jasperoid; silica replacement of silty ls and chert preserving some sedimentary textures; quartz veins	Xenomorphic in jasperoid; jigsaw-puzzle in chert; grandular quartz in veins	Jasperoid; silica replacement of silty ls; abundant sericite and carbonaceous material; quartz microveinlets	Jigsaw-puzzle texture
	Hydrothermal solution brecciation at base of jasperoid		Radial masses of silica in jasperoid	Jigsaw-puzzle texture				
	Cementation of jasperoid and jasperoid breccia with quartz	Xenomorphic and reticulated textures						
	Silica veinlets[5]	Chalcedonic						
	Quartz in vugs and interstices	Euhedral to subhedral coxcomb overgrowths						
	Gold mineralization Jasperoid; silica replacement of silty ls and cal siltstone preserving sedementary textures	Jigsaw-puzzle texture						
Gold-ore stage	Disseminated silica-replacement of carbonate minerals	Euhedral to subhedral qtz	Multiple episodes of quartz veins	Reticulate texture	Multiple episodes of quartz veins	Xenomorphic to reticulate texture	Silica-replacement of silty ls; quartz veinlets	Silicification is jigsaw-puzzle texture; veinlets xenomorphic to reticulate texture
	Silica veinlets	Subhedral quartz			Quartz in open spaces	Granular to euhedral		
Post-main stage	Calcite veins often with barite, orpiment, or realgar		Calcite veins		Calcite veins often with realgar, orpiment or barite; stibnite + barite veins		Chalcedony veins; quartz in vugs	Quarts in vugs is xenomorphic

Table 4.5 *Paragenesis and textures of the silicification during different stages of alteration (after Berger and Bagby, 1991).*

Gold in the Carlin-type deposits occurs in the native form (Table 4.4) and is extremely fine-grained. Native silver and electrum have also been recognized (Berger and Bagby, 1991). The principal ore-stage sulphide is pyrite and marcasite and arsenopyrite are common (Table 4.4). Realgar, cinnabar, stibnite and barite are common in post-main ore stage (Radtke et al., 1980).

According to Radtke et al. (1980), five primary unoxidized ore types in the Carlin deposit can be distinguished: (1) normal, composed of dolomite, illite and quartz, with pyrite occurring as the most common sulphide, and gold as a coating on sulphides; (2)siliceous, characterized by silica replacement of carbonate with gold occurring as metallic particles enclosed in hydrothermal quartz; (3) carbonaceous ores, containing organic carbon and gold associated with sulphides; (4) pyritic ore, containing 3-10% pyrite with gold occurring as a coating on pyrite; (5) arsenical ore, containing realgar with some oripiment and minor sulpho-salts with gold occurring as films on framboidal and cubic pyrite.

Oxidised ore in the Carlin deposit consists of quartz, clay and dolomite. Limonite is common. The oxidized ore is believed to be formed by supergene oxidation of the primary sulphide-bearing ores (Bagby and Berger, 1985). However, Radtke et al. (1980) suggest that boiling of a geothermal system at and near the surface was responsible for oxidation and acid leaching, although the same authors state that boiling of hydrothermal waters usually occurs at depths typical of the porphyry environment.

Radtke et al. (1980) also established a paragenetic model for the Carlin deposit, consisting of four stages:

(1) anearly and main ore stage including quartz, pyrite, K-clays, hydrocarbon and an association of Au-Hg-As-Sb-Tl and early jasperoiod; (2) a late-main stage represented by jasperoid and quartz veins, As-Sb-Hg-Tl-S minerals and Pb-Zn-Cu-S minerals; (3) an acid-leaching and oxidation stage (post-main ore stage), with jasperoid and quartz veins, major barite, anhydrite and calcite veins; and (4) alate oxidation stage caused by weathering.

Radtke et al. (1980) also reported fluid-inclusion analyses on quartz, barite, realgar and sphalerite from the Carlin deposit. The inclusion data indicated that the temperatures of the hydrothermal fluids during the main stage of deposition were about 200^{o}C, pressures about 30 bars and salinities 3 wt.% NaCl eq. Temperatures increase to about 300^{o}C during the acid-leaching stage.

Cunningham et al. (1988) reported temperatures for several deposits in south-western China, ranging from 150° to $240^{\circ}C$, with salinities up to 5 wt.% NaCl eq., based on fluid-inclusion studies. The genetic aspect ofthe Carlin-type deposits will be presented in the following chapter of the dissertation.

5.0 GENETIC MODEL FOR EPITHERMAL GOLD DEPOSTIS CONNECTED WITH GEOTHERMAL SYSTEMS RELATED TO VOLCANIC CENTRES AND CALDERAS

It has long been recognized that epithermal precious-metal ore deposits are probably the fossil equivalents of high-temperature geothermal systems (Lindgren, 1933; White, 1981; Henley and Ellis, 1983). Recent studies of mineralization and alteration in drill core from active systems, like Broadlands, Maiotapu, Rotokawa (in the Taupo Volcanic Zone), New Zealand and from Steamboat Springs, Nevada, and of the geochemistry of geothermal fluids, have provided a basis for an understanding of the geochemistry of epithermal ore-forming systems (Browne, 1969; Hedenquist and Henley, 1985; Krupp and Steward, 1987; White, 1981). The studies indicated that epithermal ore deposits and geothermal systems have similar alteration patterns, mineralogy, temperatures, fluid compositions, and geochemical associations. The analogy between them is supported by the occurence of ore-grade concentrations of Au and Ag and other associated elements (As, Sb, Hg, Tl, W) in surface discharge material (White, 1981) from several geothermal systems.

It is the purpose of this chapter to summarize and discuss the principal physical and chemical characteristics of the active geothermal systems which are essential to an understanding of the genetic model of the epithermal gold deposits.

5.1. ACTIVE GEOTHERMAL SYSTEMS

According to White et al. (1971), geothermal systems include a source of heat within the earth's crust originating from magmatic intrusions, or regional heat flow and rocks and water affected by that heat. They include both the upwelling hot fluids and the marginal, convective down-flowing, cold recharge waters (Silberman and Berger, 1985). As with epithermal deposits, structural features control the major flows of hydrothermal fluids. The surface expressions of geothermal systems are hot springs, fumaroles, siliceous sinter and often hydrothermal explosion breccias, which are also found in some epithermal deposits (Browne, 1978). The high heat source, source of water and zone of permeability necessary for generation of geothermal systems are found in tectonically active regions, such as convergent plate boundaries related to relatively recent regimes of plate tectonic interaction (Henley, 1985). Fig. 5.1. shows the distribution of geothermal systems related to lithospheric plate boundaries.

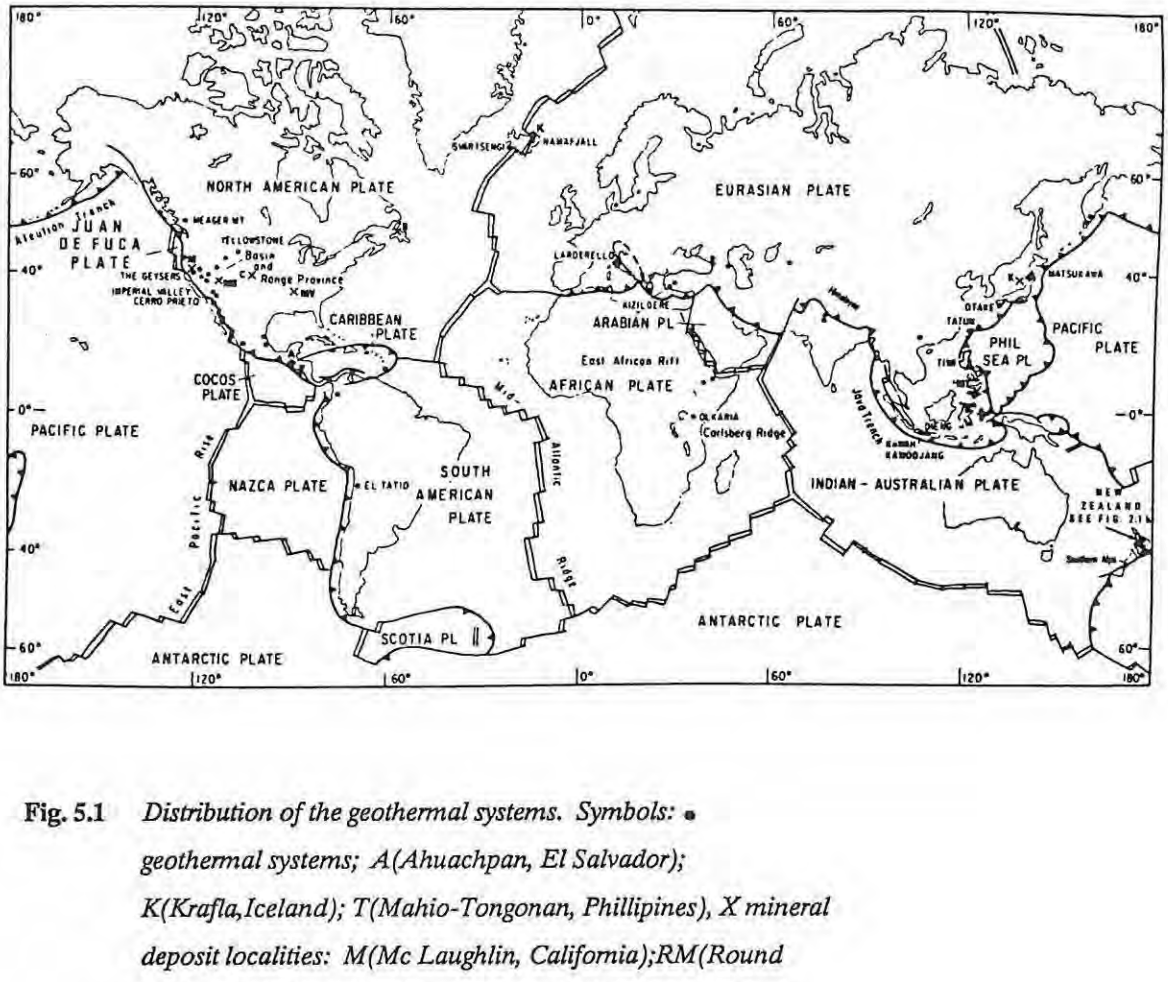


Fig. 5.1 *Distribution of the geothermal systems. Symbols:* • *geothermal systems; A(Ahuachpan, El Salvador); K(Krafla,Iceland); T(Mahio-Tongonan, Phillipines), X mineral deposit localities: M(Mc Laughlin, California);RM(Round Mountain, Nevada); C(Creede, Colorado); MV(Mississippi Valley); K(Kuroko,Japan) (after Henley et al., 1985).*

5.1.1 Physico-chemical characteristics of geothermal systems

The bulk of the fluids in typical geothermal systems have maximum temperatures of 350°C and waters which are predominantly meteoritic in origin, with near-neutral pH (Henley and Ellis, 1983). The salinity of the fluid is commonly low, about 1.6 wt% NaCl eq. The total gas content in the steam phase of geothermal discharges range from 0.01% to several tens of percent (Henley, 1985). The most abundant gas is CO2 which could originate from decarbonation reactions from carbonate rocks or from dispersed organic matter, from solution in meteoritic water or from magmatic source (Henley and Ellis, 1983). The next most abundant gas is H_2S, but CH_4, N_2, H_2 and NH_3 are also present, as well as traces of rare gases He, Ne, Ar, Kr and Xe. In liquid- dominated systems at deep levels, the gases are dissolved in the hot water, but when steam is formed by boiling, the gases pass into the steam phase. Concentrations of CO_2, H_2S and other major constituents in fluids discharged from geothermal wells in some representative geothermal fields are summarized in Table 5.1.

Geothermal fields	t^0C	Na	K	Ca	Mg	Cl	SO_4	Si_2	B	H_2S	CO_2
Wajrakei, New Zealand	250	926	154	11.8	0.04	1543	22	484	20	9	348
Broadlands, New Zealand	261	705	150	5	0.6	1238	7	556	34	72	4104
Rotokawa, New Zealand	320	265	87	0.5	0.01	520	4	710	15	245	4495
Matilo, Tongonan Philippines	324	5018	1379	122	1.4	9124	18	771	194	85	2945

Table 5.1 *Concentrations (mg/kg) of major constituents in geothermal systems (after Henley, 1991).*

Hot chloride waters (evolved by water-rock +/- magma interaction at depths of up to 8 km) convecting into the near-surface part of a large hydrothermal system may be dispersed by mixing with laterally flowing cold ground water or can be directly discharged to the surface (Henley, 1985). This is the consequence of boiling (phase separation) that occurs in the near-surface regime (in the upper 1/2 - 2 km of the systems) and results in the transfer of gases, (CO_2, H_2S etc.) into the vapour phase (Henley et al., 1985). If this phase (steam-phase) encounters cold ground waters, steam-heated waters will be formed from the condensation of the steam and oxidation of co-transported H_2S. The resultant oxidation of sulphide to sulphate leads to low-pH (0 - 3.0) conditions producing acid-sulphate waters which react rapidly with host rocks giving advanced argillic alteration. These features are incorporated in the general model of the structure of a geothermal system (Henley et al., 1985), similar of those in the Taupo Volcanic Zone, New Zealand, and are shown in Fig. 5.2.

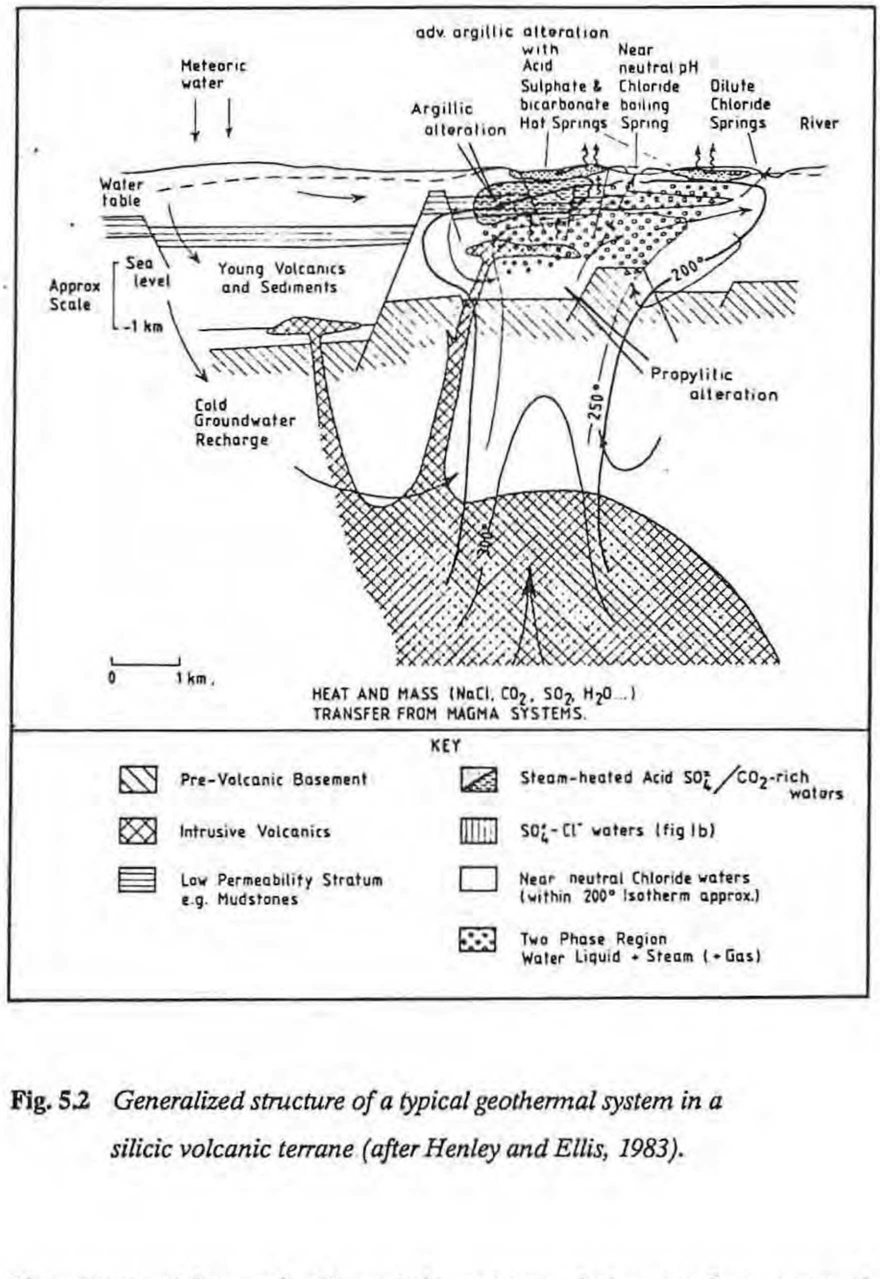


Fig. 5.2 *Generalized structure of a typical geothermal system in a silicic volcanic terrane (after Henley and Ellis, 1983).*

The resorption of CO_2-rich vapour into cooler ground water at deeper levels results in CO_2-rich waters of moderate acidity which are responsible for many of the illite-kaolinite "clay caps" often found in epithermal deposits (Henley, 1985) and the formation of adularia, calcite, quartz assemblage, indicative of near-neutral pH boiling conditions. Apart from steam-heated waters, alkaline chloride waters supersaturated with silica can be evolved in the near-surface discharge regime. These waters precipitate amorphous silica (forming sinter terraces) and silicified zones are formed, containing banded chalcedonic silica which is characteristic of hot spring epithermal environments (Henley, 1991).

Basically, the same processes described above occur in the formation of high relief andesitic-volcanic terranes, but in this case chloride waters can flow laterally for several kilometres from the hot upflow part of the system (Fig. 5.3).

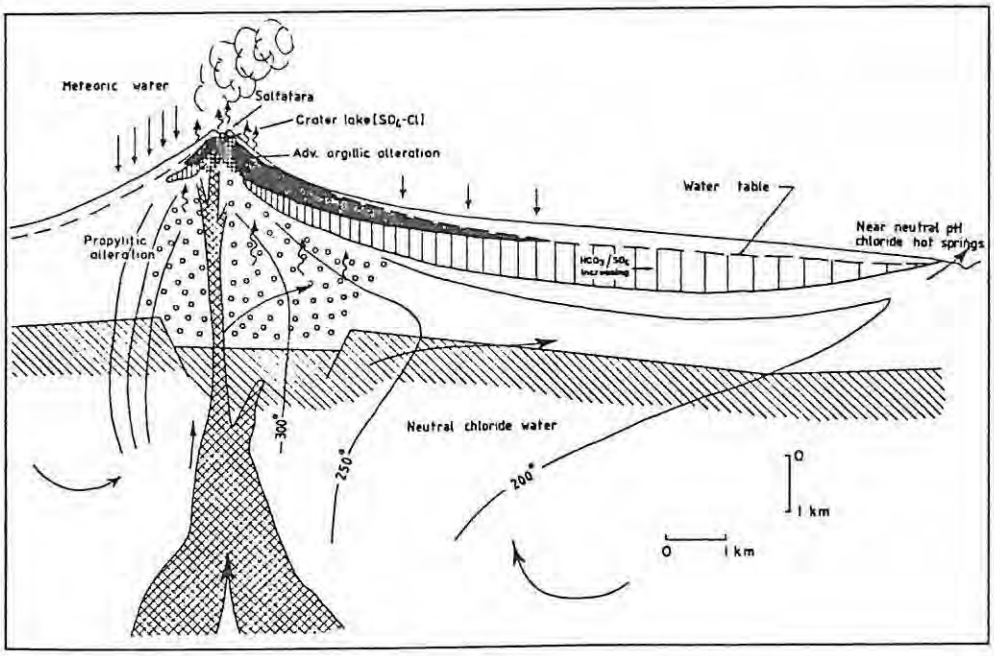


Fig. 5.3 *Structure of a typical geothermal system in andesitic-volcanic terranes. Symbols as for Fig. 5.2. (after Henley and Ellis, 1983).*

The occurrence of near-surface magmas exolving gases (HCl, SO_2, etc.) often produces high temperature fumaroles and/or acid-sulphate-chloride crater lakes (Fig. 5.3.) (Henley et al., 1985). Advanced-argillic alterations are formed by the activity of acid-sulphate waters in the upper portion of the hydrothermal systems such as Summitville (Colorado), Goldfield (Nevada). These features can also be related to the upper part of the porphyry copper systems (Sillitoe, 1983).

5.1.2 Hydrothermal alteration and mineralization

Hydrothermal alteration in geothermal systems varies with temperature, composition of the fluid and mineralogy of the host-rock. It is generally controlled by permeability, as impermeable rocks are usually less altered, even in higher temperatures, than more permeable rocks (Silberman and Berger, 1985). Table 5.2. shows the temperature ranges over which various minerals may be produced during hydrothermal alteration (Henley and Ellis, 1983).

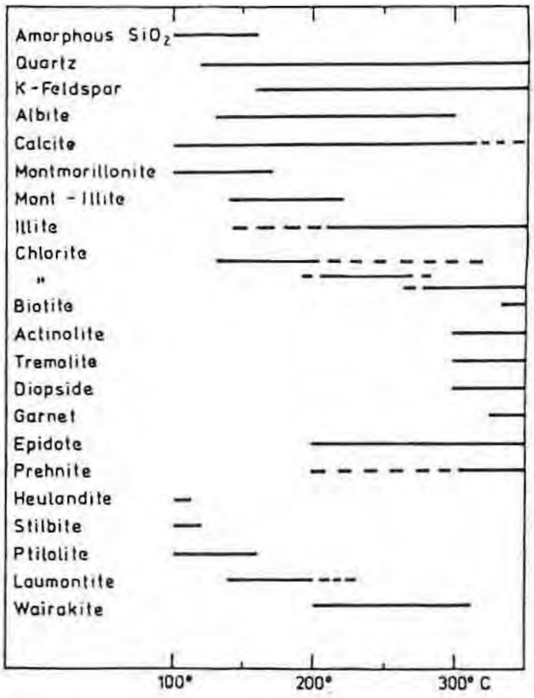


Table 5.2 *Mineral appearances according to different temperatures during hydrothermal alteration (after Henley and Ellis, 1983).*

It can be seen that at lower temperatures (at about 100°C) mineral assemblage in geothermal systems (as in the Taupo Volcanic Zone) is represented by amorphous silica or chalcedony, calcium zeolites and montmorillonite. The zeolitic assemblage shows a trend from ptilolite, at a temperature of about 100°C, through laumontite from 150° to 200°C, to wairakite at higher temperatures. With increasing temperature, montmorillonite-illite occurred until the appearance of illite at 230°C.

In the Reykjanes and Svartsengi, S.W. Iceland, geothermal system, a montmorillonite-zeolite-calcite alteration is followed by a chlorite-montmorillonite-phrenite intermediate assemblage and chlorite-epidote assemblage up to 300°C.

At Cerro Prieto, Mexico, kaolinite-montmorillonite-dolomite-Fe-hydroxides-pyrite assemblage at 140° - 160°C is followed by illite-chlorite with overgrowths of quartz and feldspar, together with calcite, pyrite and pyrrhotite. Above 250°C on epidote-wairakite assemblage is present; calcite is replaced by prehnite-actinolite and illite-chlorite above 300°C (Henley and Ellis, 1983).

The general sequence of hydrothermal alteration relates to the boiling point. In the two-phase region, the loss of steam with CO_2 and H_2S (see section 3.2 and 5.1.1) leads to increasing pH and formation of calcite and K-feldspar (adularia) in the near-surface environment. Generally, hydrothermal alteration at depth (250°-300°C) is characterized by quartz, adularia, illite, albite, chlorite, alcite, epidote and pyrite assemblages, passing into an intermediate argillic (clay) alteration assemblage with decreasing temperature,to advanced argillic alteration in the near-surface environment (Henley and Ellis, 1983). Advanced argillic alteration is represented by kaolinite, alunite, hydrated iron oxides and often opal.

High concentrations of Au and Ag and associated elements are found in surface or near-surface deposits in many geothermal systems. Fig. 5.4 shows environments of epithermal ore deposition in typical geothermal systems associated with a silicic volcanic terraine (A) and a calc-alkaline stratovolcano (B).

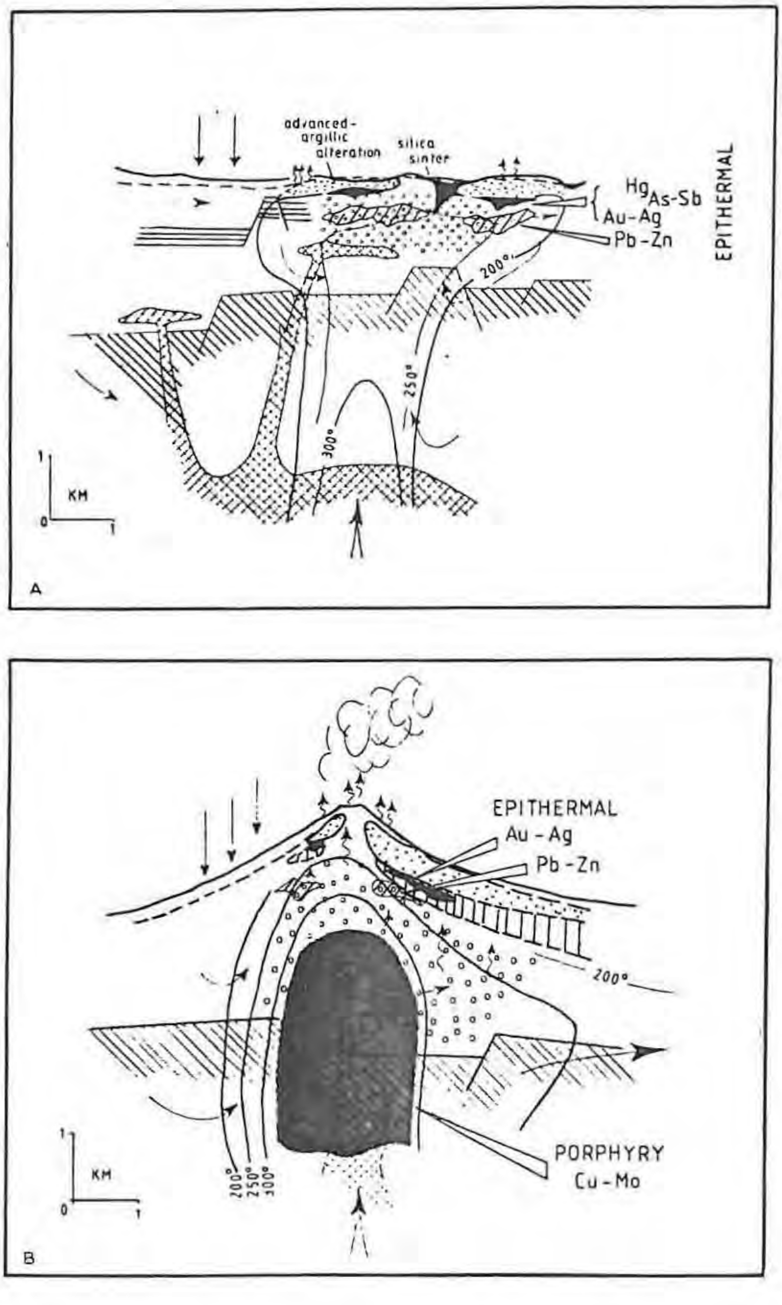


Fig. 5.4 *Models of the two types of epithermal ore deposits in a geothermal system. A - Adularia-sericite type, B - Acid-sulphate type (modified after Henley and Ellis, 1983).*

Usually groups of elements As, Sb, Hg, B, Tl and Au are concentrated in the upper part of geothermal systems and decrease with depth, whereas base metals Cu, Pb, Zn and other elements Bi, Se, Te, Co and to some extent Ag tend to be precipitated at greater depths in higher temperature zones (Fig. 5.4 A, B) (Silberman and Berger, 1985).

In Waiotapu and Broadlands, New Zealand, geothermal systems, Au and Ag reach 85 and 500 ppm in a hot spring environment, in the amorphous Au-Ag-Sb-As-S precipitates (Weissberg et al., 1979). Base metal sulphides (galena, sphalerite, chalcopyrite with pyrite and pyrrhotite) occur in veins and vugs often associated with adularia and calcite, at depths of 200-300m and temperatures between 120°-300°C (Browne, 1969). At Broadlands, metalliferous zoning is represented by Sb, Au and Tl enrichment near the surface and Pb, Zn, Ag, Cu, Bi, Co, Te and Se at depth. Gold in the dilute geothermal fluids in this field is transported as thio complexes and its deposition in hot springs and sub-hot spring environments reflects the loss of H_2S due to boiling. According to Seward (1991), colloidal As-Sb sulphides may scavenge gold and other metals and produce the ore grade precipitates.

5.2. GEOTHERMAL INTERPRETATION OF EPITHERMAL GOLD DEPOSITS

The conceptual base on which recent genetic models of epithermal gold deposits are established generally come from observation of active geothermal systems (White, 1981). The general structure of the volcanic-hosted epithermal gold deposits is shown in Fig. 5.4 (A,B) and should be compared with the general structure of the geothermal systems shown in Figs. 5.2 and 5.3 (see section 5.1).

The essential components required to form epithermal gold deposits, as with all hydrothermal deposits, are source of heat, fluids and metals and high-permeability structures which allow fluid convection and metal deposition.

5.2.1 Heat source

Geothermal systems occur in the tectonically active zones of the earth's crust. The most favourable regions for a source of high heat flows are convergent plate boundaries (Andean magmatic arcs, Pacific island arc; back-arc extensional cratonic terranes; as mentioned above (see sections 4.1.1; 4.2.1 a; 4.2.2 a). Hot spots, continental rifts and transform faults can also have high heat flows (Henley, 1985).

The high heat flows responsible for the formation of almost all epithermal gold deposits in these tectonic settings are related to magmatism (calcic, calc-alkaline, alkali-calcic). Doubt exists over the origin ofthe Carlin-type gold deposits, where the relationship between mineralization and magmatism is still not clear. However, the contemporaneity of magmatic activity and mineralization at Getchell, Prebble, Gold Acres and Cortez is compelling evidence for the involvement of magmatic heat in the formation of Carlin-type deposits (Berger and Bagby, 1991). For the formation of many Carlin-type deposits in extensional terranes, such as the Great Basin, western United States, probably the main source for the high heat flow was basaltic magma (Berger and Bagby, 1991). According to Henley and Hoffman (1987), the heat from about 100 km^3 of cooling basaltic magma or approximately 200 km^2 of granitic magma (large stock or batholith) is required for the formation of these deposits.

Generally, the high temperature upflow regime, having sufficient heat energy to support convective systems, can extend from depths of around 8 km and can have a horizontal cross-section of 3-5 km^2 (Henley and Hoffman, 1987). Some authors also reported that for the formation of a typical 30 x 10^6 epithermal gold deposit, approximately 1.3 x 10^{20}J of heat energy, at a temperature of 250°C is required, assuming that the fluid contains in the order of 10 µg/kg Au in solution. Additional heat of about 0.5 x 10^{20}J is required to bring the upflow region to the ambient temperature of the convecting fluids.

In other tectonic settings, such as the collision-related mountain belts in the Southern Alps (New Zealand), Tibet and Northern India, geothermal systems are related to rapid uplift, where hydrothermal fluids are of metamorphic dehydration origin in response to post-metamorphic uplift and/or overthrusting (Henley, 1985).

5.2.2 The origin and composition of hydrothermal fluids

Oxygen, hydrogen, carbon and sulphur stable-isotope studies (^{18}O, D, ^{13}C, ^{34}S) in many epithermal environments showed that epithermal isotope characteristics are intermediate between those of near-surface geothermal systems and deeper hydrothermal deposits (Field and Fifarek, 1985). Depletions of D and ^{18}O in the host rocks, minerals and inclusion fluids of the epithermal deposits and similar composition of these fluids to meteoritic waters, suggest that the hydrothermal fluids were predominantly ground water of meteoritic origin (Taylor, 1973; O'Neil and Silberman, 1974; Radtke et al., 1980; Field and Fifarek, 1985; Hayba et al., 1985), although some of them had varying amounts of magmatic components (Creede, Summitville, Sunnyside) (Field and Fifarek, 1985).

The ore fluids at adularia-sericite type deposits (Creede, Colorado) can have a magmatic component of up to ten percent. At Creede, episodic introduction of magmatic fluids into the circulating system has been recognised from isotopic studies of rodochrosite and siderite. The predominant meteoritic waters, were accumulated by processes of evaporation and diagenesis in the playa lake in the moat of the Creede caldera. Sulphur-isotope studies indicated many possible sources of sulphur. Most sulphate originated from playa lake sediments, whereas sulphide sulphur derived probably from large reservoirs or magmatic sulphur at depth (Bethke et al., 1973).

Carbon-isotopic data from Creede indicated a magmatic origin for carbon (Bethke and Rye, 1979), while in some other adularia-sericite type deposits, as in Sunnyside, it could have been derived from dissolution of marine sediments or from atmospheric CO_2 dissolved in circulating meteoritic water (Casadevall and Ohmoto, 1977).

There is no hydrogen-isotope data available for any acid-sulphate-type deposits and therefore the source of hydrothermal fluids is unclear (Hayba et al., 1985). However, oxygen-isotope data from Goldfield have been used to document the meteoritic origin of the fluid at Goldfield (Taylor, 1973). Bruha and Noble (1983) reported that a significant number of magmatic components was contributed to the hydrothermal fluids. Whitney (1984) calculated the sulphur speciation in quenched magmatic gases evolved from magmas of various oxidation states. Relatively oxidized magmas produce gases rich in SO_2, which is unstable at temperatures below 400°C disproportionating into H_2SO_4 and H_2S. According to Hayba et al. (1985), sulphuric acid produced by this process interacts with the wallrock resulting in the formation of advanced argillic alteration associated with ore in acid-sulphate type deposits. The formation of epithermal precious-metal deposits requires fluids of low-salinity and high gas content (H_2S, CO_2), whereas silver-rich base-metal deposits (relatively low in gold) require high-salinity, low-gas fluids in their hydrothermal systems. The salinities of the hydrothermal fluids reflect their volcanic-tectonic environment. For example, andesite-hosted hydrothermal systems are probably formedthrough high-level interaction with volcanic gases or the involvement of deep, connate waters, more saline than those of silicic-volcanic terranes, but less than those where evaporite sequences occur (Fig. 5.5).

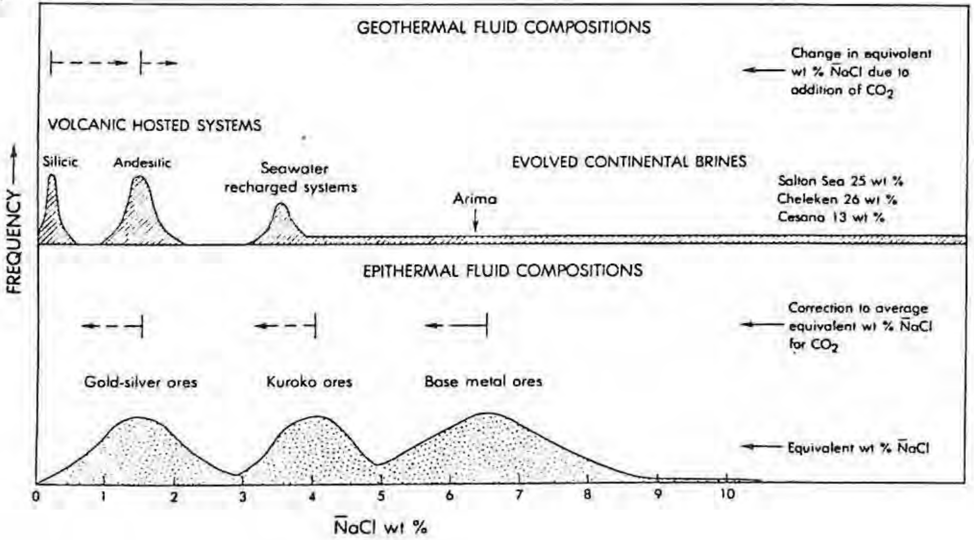


Fig. 5.5 *Distribution of fluid salinities in the earth's crust in relation to host-rock and crustal environments (after Henley, 1985).*

The ore-forming fluid in the Creede district has a salinity range from 4-12 wt % NaCl eq. (see section 4.2.1.d Table 4.3) which is higher than in any other fluids of adularia-sericite type gold deposits. Such high-salinity fluids occur in systems where evaporites are present, such as caldera moats and grabens and can also reflect climatic control (Henley, 1985).

The salinities of acid-sulphate type deposits are still not well documented. Fluid inclusion data indicated salinities of 4 and 6 wt % NaCl eq., at Summitville and between 1 and 6 wt % NaCl eq. at Red Mountain. (in Hayba et al., 1985). Ore deposition in both adularia-sericite and acid-sulphate type deposits occurred at temperatures between 190°-300°C and 40-50 bars pressure (Hayba et al., 1985).

Parameters which affect the chemistry of the hydrothermal fluids are temperature, pressure, activities of S_2 and O_2, pH, activity of Cl and the total sulphur concentration in the fluid (Heald et al., 1987). The log aS_2 - log aO_2 and log aO_2-pH diagrams (Fig. 5.6 and 5.7) show separate geochemical environments for the highly sulphidised and less sulphidised mineralogies of the acid-sulphate and adularia-sericite type deposits.

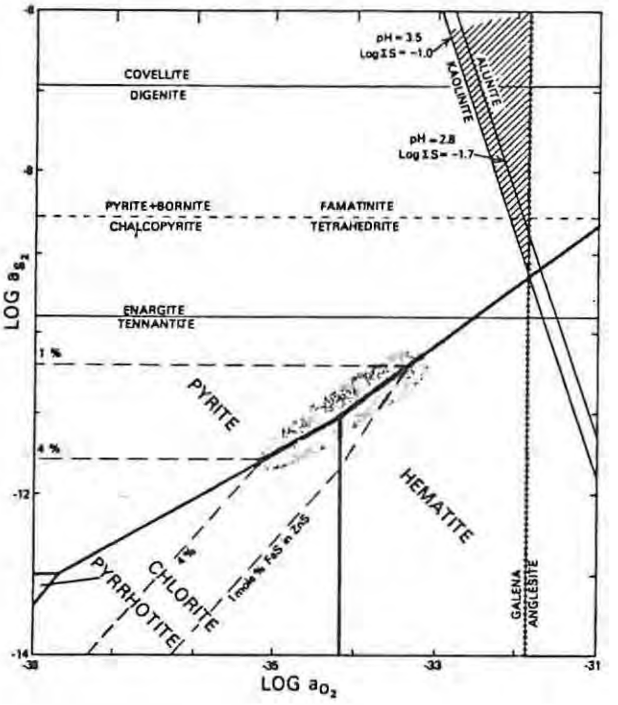


Fig. 5.6 *Log aS_2 - aO_2 diagram showing the mineral stability fields for the significant minerals in epithermal systems at 250 °C. The line pattern and the stippled pattern show the most likely environments for the acid-sulphate and adularia-sericite type deposits. Thick solid lines = boundaries between pyrite, pyrrhotite, chlorite and hematite; long light dashes = iron content of sphalerite; medium solid lines = boundaries between alunite and kaolinite at different total sulphur concentrations; light solid lines = reactions between sulphides and sulphosalts (after Heald et al., 1987).*

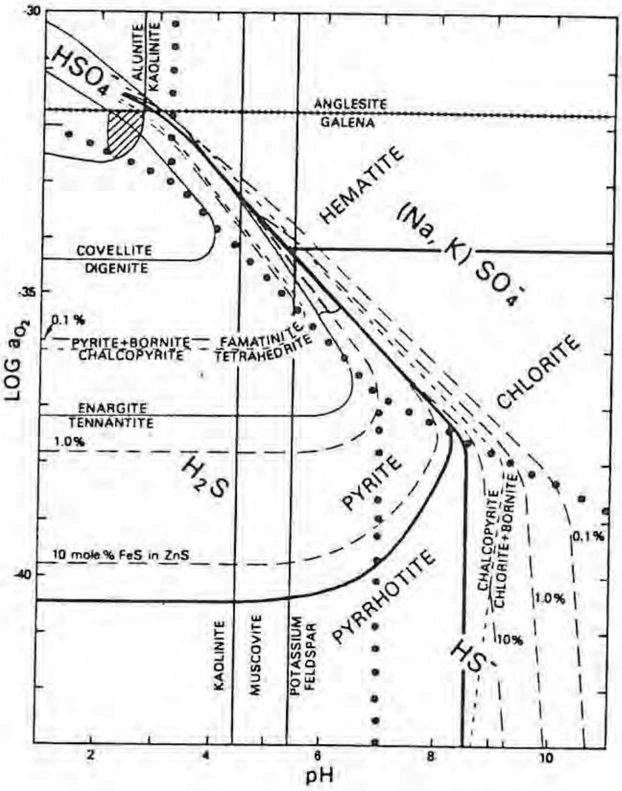


Fig. 5.7 *Log aO_2 - pH diagram at 250°C; ΣS = 0.02 molal (log ΣS = -1.7) and salinity = 1 molal with Na/K = 9. The line pattern, the stippled pattern, thick solid lines, long light dashes and light solid lines show the same features as Fig. 5.6. Heavy dotted lines = boundaries between aqueous sulphur species; short dashes = limit of stability of chalcopyrite; medium solid lines = boundaries between alunite, kaolinite, muscovite and K-feldspar (after Heald et al., 1987).*

Total sulphur concentration is likely to be larger in acid-sulphate deposits. The presence of alunite in association with ore and the presence of pyrite, galena, enargite and covellite indicate an environment of high sulphur fugacity (Fig. 5.6).

For adularia-sericite type deposits, an environment of lower sulphur and oxygen activity is indicated by the presence of pyrite, chlorite and hematite in these deposits (Fig. 5.6). Fig. 5.7 shows that adularia-sericite type deposits form at significantly higher pH than acid-sulphate type deposits (Heald et al., 1987).

The chemical nature of the ore-forming fluids in Carlin-type deposits is complex. Isotope and fluid inclusion gas chemistry data identified three chemically distinct fluids and two of these were recognised to be from deep upwelling brines (Hofstra et al., 1988). At Jerritt Canyon, the earliest stage (pre-ore jasperoid formation) was characterized by an isotopically heavy fluid, with 3 - 10 wt % NaCl eq. and CO_2 dominated the gas phase. This early fluid was at 250^o-300^oC and 1 to 1.5 kbars pressure. The fluid was responsible for the precipitation of silica, forming tabular, often stratiform jasperoid bodies in the ore host. This stage was followed by progressive mixing of the isotopically heavy fluid with a lighter overlying gold-and-CO_2-H_2S-bearing fluid (gold-depositing stage). The last stage was characterized by total swamping of the fluid system by meteoritic waters depositing late, cross-cutting calcite veins (in Berger and Bagby, 1991).

Radtke et al. (1980) suggested that the ore deposition at Carlin resulted from the mixing of an upwelling fluid of original meteoritic origin with a second dilute, meteoritic fluid predominating in the waning stages of mineralization.

The fluid associated with the gold depositional stage was low-salinity, high CO_2 and H_2S and 500-800 bars pressure and the latest stage fluid was low-salinity and low CO_2 (Kuehn and Rose, 1987).

Most of the sulphide-sulphur in pyrite-galena-realgar-sphalerite-stibnite ore at Carlin was derived from the hydrothermal remobilization of diagenetic pyrite in the sedimentary host rocks (Radtke et al., 1980). According to Radtke et al. (1980), most of the carbon in hydrothermal calcites was probably extracted by dissolution reactions from the carbonate host rocks at depth.

5.2.3 Sources of metals

Gold-depositing geothermal systems transport in the order of 5-10 μg/kg gold (Henley and Hoffman, 1987). If the gold is derived from leaching of metal from crustal rocks, then a typical rock containing an average of 3 mg/t gold, with a volume of 130 km^3 would be sufficient to supply the metal to the geothermal system, if extraction is 100% efficient. However, it is unlikely that the extraction process is 100% efficient, therefore a larger volume of rock is needed for each individual system (Berger and Bagby, 1991).

For many adularia-sericite-type deposits, Pb-isotopic studies indicated that the Pb may have been derived from Precambrian basement or Phanerozoic rocks and therefore suggested that presumably most of the other metals originated by leaching of wallrocks underlying the volcanic rocks (Doe et al., 1979).

For acid-sulphate-type deposits (Summitville and Red Mountain), Pb-isotopic studies showed that the galena is very similar isotopically to the enclosing volcanic rocks, suggesting that Pb originated either from adjacent rocks or related magmatic fluids (Doe et al., 1979).

The source of gold in Carlin-type ore deposits has not directly been determined. According to Henley and Hoffman (1987) and Berger and Bagby (1991), the source of gold is basaltic magmas emplaced into the lower or middle crust, rather than leaching from crustal rocks.

5.2.4 Gold transport in epithermal systems

Experimental studies and direct observation in geothermal systems confirm that gold in the crustal environment of epithermal systems is transported as gold bisulphide complexes, dominated at near-neutral pH by the $Au(HS)_2^-$ and temperatures up to 300°C. At slightly higher temperatures (350°C) the complex $AuHS^0$ is more concentrated than $Au(Hs)_2$ and therefore more important (see section 3.1). Analogous silver and copper bisulphide complexes control the transport of these metals under low-salinity conditions, but at higher salinities and temperatures (higher than 360°C) chloride complexes are responsible for metal transport (Henley and Brown, 1985). The solubility of gold and base metals (Ag, Pb, Zn) can be calculated for a wide range of salinities and H_2S contents. Since salinity effects a control on the equilibrium pH of the system fluid, the solubility of gold is itself a function of salinity and can be expressed as a function of salinity, temperature and CO_2 as shown in Fig. 5.8.

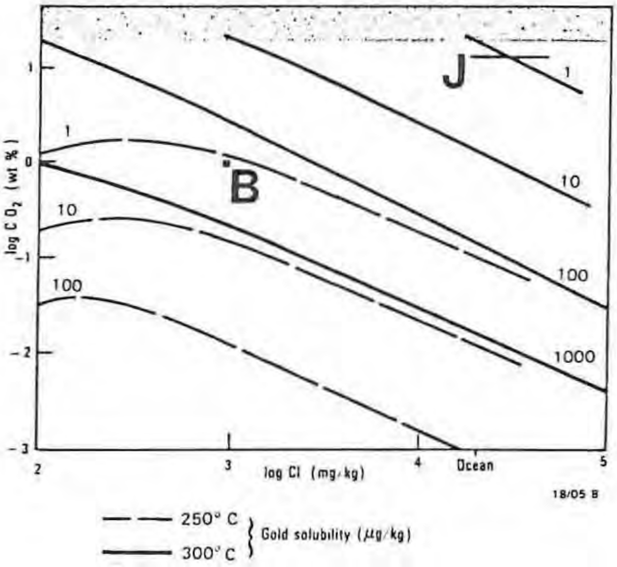


Fig. 5.8 *Equilibrium solubility of gold in hydrothermal fluids buffered by the assemblage K mica-K feldspar-calcite-quartz-pyrite-FeO silicate in chloride-sulphide-CO_2 solutions at 250^o and 300^oC. The stippled area provides a general indication of the two-phase region for CO_2 - H_2O (after Berger and Henley, 1989).*

It can be seen from the figure that gold solubility as a chloride complex is significantly higher in the temperature range 300^o to 350^oC than at lower temperatures. The solubility of gold decreases with increasing salinity (due to pH control) and increasing CO_2 content, but strongly increases with respect to temperature (Berger and Henley, 1989).

While in low-salinity systems high CO_2 contents and concomitant high H_2S contents are essential for significant gold transport, gold transport can occur in more saline environments at relatively low CO_2, provided that the source regime temperature is in excess of 250^oC (Berger and Henley, 1989).

5.2.5 Gold deposition in epithermal environments

Gold deposition in epithermal environments is caused by changes in pH, temperature, pressure, redox potential and activities of complexing ligands (Berger and Henley, 1989). The most common mechanisms for deposition are boiling (phase separation), mixing of fluids of different temperatures and salinities, reactions between fluids and wall rocks and dilution and cooling of fluids (Buchanan 1981; Hayba et al., 1985; Berger and Henley, 1989).

Boiling has attracted most attention in recent studies, as the process which is believed to be responsible for the deposition of gold, sulphide, carbonate and silicate minerals in most of the epithermal environments (Buchanan, 1981).

Fig. 5.9 illustrates the geologic and hydrologic framework of a boiling geothermal system and Fig. 5.10 (flow-diagram) shows the chemical components and processes (boiling and mixing reactions) in hydrothermal systems.

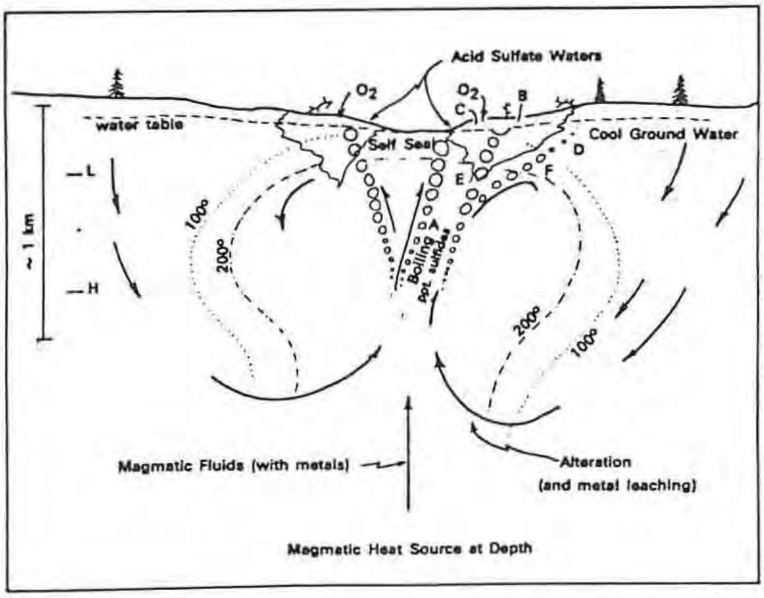


Fig. 5.9 *Schematic diagram of a boiling hydrothermal system. The letters (A - F) are described in the text (after Reed and Spycher, 1985).*

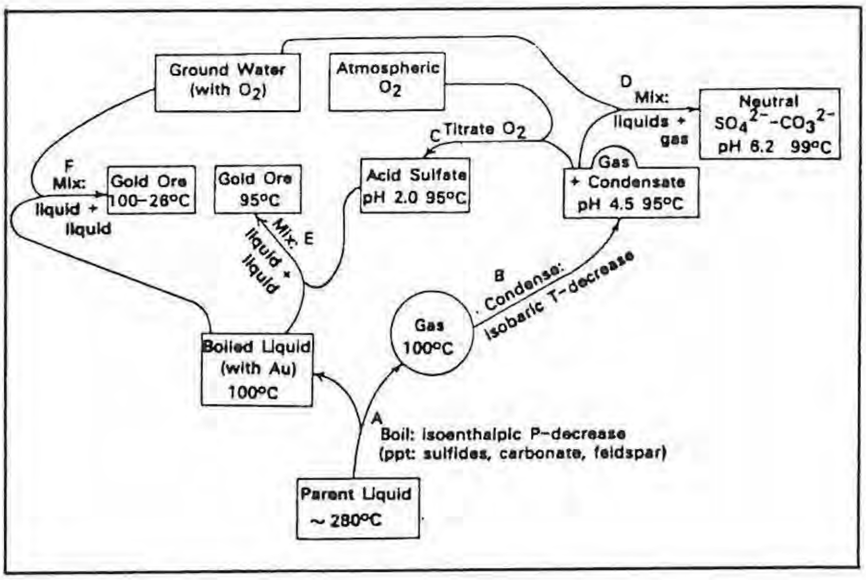


Fig. 5.10 *Flow-diagram showing boiling, mixing and reaction processes in geothermal systems. The letters (A - F) are described in the text (after Reed and Spycher, 1985).*

These include boiling (labelled by letter A), condensation of the boiled gas in rock (B), oxidation of the gas by the atmosphere (C), condensation followed by oxidation of the gas in cool, fractured ground (D), mixing of acid ground waters with the boiled liquid (E) and mixing of cold ground water with the boiled liquid (F) (Reed and Spycher, 1985).

The boiling starts at greater depth (H), where the hydrostatic pressure is equal to the vapour pressure (see section 3.2), or at shallower depth (between L and H), where pressure is higher than hydrostatic, but less than lithostatic (Fig. 5.9). The initially homogenous aqueous phase separates into gas, liquid and minerals (Fig. 5.10). Boiling induces a temperature decrease and CO_2 gas release which in turn causes increasing pH and precipitation of base-metal sulphides and carbonates (between 278°-245°C). Under these conditions sphalerite and galena precipitate before chalcopyrite because their chloride and carbonate complexes are unstable. Precipitation of chalcopyrite begins late in the boiling process because of its very stable bisulphide complex (Reed and Spycher, 1985). Additional sulphide continues to precipitate throughout boiling down to 100°C, but in smaller quantities.The temperature decrease causes precipitation of quartz, K-feldspar, chlorite, muscovite. Feldspar, muscovite and chlorite precipitated because of the destabilisation of the complexes of Al and Fe.

The critical reaction for gold precipitation in the acid to neutral pH range is the following:

$$\underset{(aq)}{8Au(HS)_2^-} + \underset{(aq)}{6H^+} + \underset{(aq)}{4H_2O} \longrightarrow \underset{(gold)}{8Au^0} + \underset{(aq)}{SO_4^{2-}} + \underset{(g\ or\ aq)}{15H_2S}$$

As the pH increases, gold solubility increases as H_2S dissociates to H_5. The escape of H_2S gas during boiling, combined with the precipitation of sulphide minerals, removes aqueous sulphide, tending to precipitate gold and destroying the gold bisulphide complex (Hedenquist and Henley, 1985; Reed and Spycher, 1985). Gold precipitates in the deeper vein environment, along with sulphides, if the system is sulphide-deficient. In these conditions, sulphide may be depleted by H_2S escape and sulphide precipitation. Base-metal-rich waters (large metal/sulphide ratio) are necessary to assure sulphide depletion with consequent gold precipitation and large salinities are necessary for large base-metal concentration. Gold and galena precipitate in accordance to the following reaction (Reed and Spycher, 1985):

$$\underset{(aq)}{8Au(HS)_2^-} + \underset{(aq)}{4H_2O} + \underset{(aq)}{15PbCl_2} \longrightarrow \underset{(gold)}{8Au^0} + \underset{(aq)}{SO_4^{2-}} + \underset{(aq)}{30Cl^-} + \underset{(galena)}{15PbS} + \underset{(aq)}{24H^+}$$

Thus, the deeper vein gold-sulphide ore is likely to have formed from more saline waters than near-surface (hot spring) deposits.

Condensation of the boiled gases (Fig. 5.10) at 95°C produces carbonic acid, causing a decrease of the pH to 4.5. This pH is acidic enough to effect alteration of volcanic wall rocks to illite or kaolinite, but not enough to produce kaolinite-alunite alteration at shallow level. For this, acid-sulphate waters of very low pH (2 - 4) are responsible. These are produced by oxidation of the condensate and remaining H_2S by atmospheric O_2.

Boiled gases can oxidize by mixing with oxygenated ground water producing sulphate-carbonate water of near-neutral pH (Fig. 5.10). Ground waters carry significant concentrations of iron and react with H_2S (from boiling hydrothermal water) producing a "pyrite halo" around epithermal gold deposits as well as around porphyry copper deposits (Reed and Spycher, 1985).

Acid sulphate waters mix with gold-bearing liquid remaining from boiling and cause the precipitation of covellite and gold by acidification which converts aqueous bisulphide into H_2S destroying the $Au(HS)_2^-$ complex (at 95°C). According to Henley (1985), As, Hg and Sb sulphides also precipitate by this process.

When cold ground water mixes with the residual boiled gold-and-base-metal-bearing liquid, it causes precipitation of sphalerite, galena, acanthite and chalcopyrite (above 65°C). Pyrite precipitates when iron from ground water reacts with sulphide in the hot water. Gold precipitates above 70°C because the aqueous sulphide concentration has been sufficiently decreased by oxidation to destroy the gold-bisulphide complex (Reed and Spycher, 1985).

For most of the adularia-sericite- type deposits, boiling is the major mechanism of deposition. Gold deposition in these environments is accompanied by banded silica, calcite, adularia and pyrite. Deposition generally occurs in the fractures or major vein systems (Berger and Henley, 1989). In the Creede district, two processes, boiling and mixing, were responsible for mineral deposition (Hayba et al., 1985). Intense sericite alteration overlying a zone of vein adularia (+ carbonate in some districts) provides evidence for boiling in the system. However, mixing is believed to be a more important process causing most, if not all, ore deposition at Creede (Hayba et al., 1985). Base- and precious-metals are deposited by the mixing of the deep, saline ore fluid with overlying fresh ground water.

Reaction between sulphuric acid and wall rocks produces quartz + kaolinite + alunite alteration assemblages in acid-sulphate-type deposits. According to Hayba et al., (1985) there are three different origins of acid-sulphate alteration due to the formation of sulphuric acid in three different environments in accordance with the following reactions:

(1) $4SO_4 + 4H_2O \rightarrow 3H_2SO_4 + H_2S$
(2) $H_2S + 2O_2 \rightarrow H_2SO_4$
(3) $2FeS_2 + 7H_2O + 15/2O_2 \rightarrow 2Fe_2O_3.3H_2O + 4H_2SO_4$

Reaction (1) is the most important process by which sulphuric acid is formed by disproportionation of magmatically exsolved SO_2 gas deep in hydrothermal systems. Many active geothermal systems associated with andesitic volcanism in island arc settings contain deep, high-temperature, acid waters (Stoffregen, 1987; Hayba et al., 1985).

Reaction (2) is the oxidation of hydrogen sulphide occurring when the vapour phase, generated by boiling of the deep waters, contacts the atmosphere just above the water table (Berger and Eimon, 1983; Hayba et al., 1985). The sulphuric acid generated percolates back into, and acidifies, the steam-heated ground waters which overly the deeply circulating hydrothermal cell. These surficial, acid, steam-heated waters are essential elements of almost all high-temperature geothermal systems, from which both acid-sulphate and adularia-sericite types of epithermal deposits form (Henley and Ellis, 1983; Hayba et al., 1985).

Reaction (3) involves the production of sulphuric acid during supergene alteration of sulphide ore. Such oxidation produced alunite at Creede and Goldfield. Reactions (2) and (3) can also occur in adularia-sericite systems, as surficial zones of acid-sulphate alteration due to the oxidation of H_2S in the steam-heated water zone and are common in active geothermal systems (analogous to fossil systems with adularia-sericite characteristics).

The gold deposition in Carlin-type deposits was due to the mixing process (Berger and Bagby, 1991). Magmatic, metalliferous and gas-bearing fluid interacts with sediments producing an isotopically heavy, CO_2-enriched hydrothermal solution. These early brines which carry base metals as chloride complexes react with carbonate rocks, replacing carbonate with silica and forming pre-ore jasperiods, without precipitating base- and precious-metals. When a steady state is reached, the magmatic input continues and circulating meteoritic waters begin to replace the basin brines. The system reaches thermal equilibrium with the country rocks. Within the upwelling plume, aqueous sulphide dominates as metal complexes and gold transport takes place.

Fig. 5.11 shows the schematic form of Carlin-type deposits.

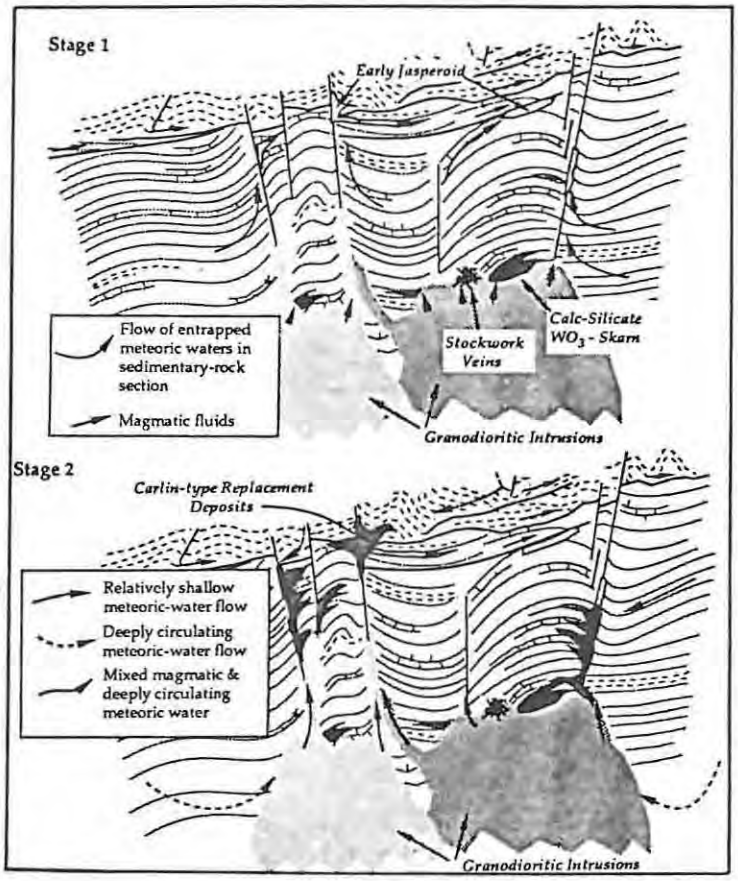


Fig. 5.11 *Schematic model for the formation of Carlin-type deposits. Stage 1: Initial mineralization consists of probable skarn and possible stockwork metallization and formation of early jasperoids, distal to the complex. Stage 2: Gold mineralization (close or distal to the intrusion) coincides with early, pre-ore jasperoids. The early jasperoids may be above, within or beneath the gold mineralization (after Berger and Bagby, 1991).*

Gold deposition is due to chemical changes associated with wall rock reactions (Berger and Bagby, 1991). Hydrothermal conditions were quite acid at the site of gold deposition at the Getchell and Mercur deposits. However, at Getchell, replacement of marcasite by pyrite and association of pyrite with gold in most deposits indicate changes in chemical conditions during gold deposition (Berger and Bagby, 1991).

With time, magmatic input decreases and dilute meteoritic waters dominate the systems. Lesser amounts of gold were deposited at this stage. Marcasite and barite of late paragenesis imply a lower pH in the waning stages of the hydrothermal system.

6.0 APPLICATIONS TO EXPLORATION

Epithermal gold deposits are attractive exploration targets currently dominating the mineral exploration scene in the western United States and countries in South America's Cordillera and the circum-Pacific Rim. Their widespread distribution and varied character in these regions encourages the view that there are opportunities for further discoveries. Regions which are relatively well known for hosting gold mineralization allow better and easier discrimination of favourable areas than regions where little exploration has been done. The latter areas provide opportunities to approach exploration from the regional to the prospect scale exploration. Only the quantitative understanding of genetic models for epithermal gold deposits, including tectonic, igneous and structural settings and the physico-chemical characteristics of their hydrothermal systems provide the basis for their systematic and successful exploration.

This chapter attempts to apply the models developed in chapter 5.0 in order to select all favourable areas and approach exploration from a regional to a prospect scale. An area which is not known, to date, to host significant epithermal gold deposits is for example Serbia (eastern Yugoslavia), in south-east Europe, and will be used as an example in section 6.4.

6.1 SELECTION OF AREAS FOR REGIONAL EXPLORATION

In previous chapters (4.0 and 5.0), the environments of epithermal mineralization were discussed in terms of tectonic, igneous and structural settings and the physico-chemical processes of gold deposition. Of these, the first two represent the basis for the selection of favourable areas for regional exploration. Ideal tectonic settings for porphyry-related and volcanic-hosted epithermal deposits related to geothermal systems are subduction-related plate margins.

The critical aspect controlling whether epithermal mineralization occurs or not, is not the distribution of volcanic rocks or host rock lithologies, but rather the distribution of intrusions below the surface that provide heat for hydrothermal circulation and magmatic components for the system. Therefore the extent of the igneous province in favourable tectonic settings should first be determined. This includes all the igneous rocks (volcanic and plutonic) related to the igneous phase. Calc-alkaline to alkaline provinces have been recognized to be the most prospective. According to Ishihara (1981), there is a clear-cut association of the hydrothermal Cu and Au deposits with I-type, magnetite-series, meta-aluminous-peralkaline felsic magmas, generated from hornblende melting in island arcs and continental lip arcs, such as the circum-Pacific Rim. This association has been linked further and specifically to subduction of hydrated oceanic lithosphere at compressional convergent margins (Sillitoe, 1987).

On the other hand, there is a notable absence of significant Cu and Au hydrothermal deposits related to ilmenite-series per-aluminous magmas generated from muscovite melting in continental collisional tectonic environments (Mason, 1991a). Epithermal deposits are formed at shallow crustal levels, so regions which have been deeply eroded are in general less prospective.

Having selected regions likely to have had favourable heat flow characteristics and which have not been too deeply eroded, the next step is the structural consideration. The major structural patterns, such as concentric and radial volcanic-related structures, calderas, intersections of major faults, deep-seated extensional fault zones and rifts, may be recognized on a variety of regional data sets, from geological maps to aeromagnetic surveys and satellite images.

In order to select favourable regions for Carlin-type gold exploration, the following factors should be present:

- The area should have undergone extensional tectonic processes. Most of the known Carlin-type deposits occur within or adjacent to overthrust terranes related to continental margin tectonics.

- Occurrence of mid- to upper-crustal plutonism.

- The most favourable rocks hosting the Carlin-type deposits are silty to argillaceous, thinly laminated, marine, carbonate rocks or carbonate-bearing siliciclastic rocks.

- The presence of high-angle normal and strike-slip faults, their intersections, antiformal structures and thrust faults.

Once the favourable area has been selected, regional exploration for epithermal gold deposits can commence.

6.2 REGIONAL EXPLORATION

The discovery of many epithermal gold deposits in the circum-Pacific region and western United States has resulted from the application of basic geology and geochemistry, with geophysics playing a minor role. More recently, increased use of geophysical methods has resulted in rewarding advances in the geological mapping of gold prospects and in the definition of drill targets (Irvine and Smith, 1990).

Geophysical surveys, such as aeromagnetic, radiometric and resistivity are very useful in regional- and semi-regional-scale exploration to locate epithermal systems and to help guide initial drilling.

The relatively large areas of argillic-propylitic alteration that overlie or are marginal to epithermal systems are characterized by low resistivity and low magnetisation. Very low resistivity zones may be caused by the reaction of the rocks with acid, steam-heated waters. Most hydrothermal fluids, including relatively cool CO_2-bearing waters, destroy magnetite and cause demagnetization anomalies. Airborne electrical and magnetic survey methods are therefore ideal for detecting epithermal systems in large exploration areas (Allis, 1990).

The discovery of the Hishikari (Japan) epithermal deposit was the result of an airborne electro-magnetic survey combined with ground resistivity syrveys. Airborne magnetic surveys have been very successful in epithermal exploration throughout eastern Australia (Irvine and Smith, 1990). They also provide basic geological information and can locate major structures which may control the location of epithermal gold deposit.

An airborne gamma-ray spectrometry survey may also be very useful in detecting high potassium (adularia, illite) anomalies and may assist in surface hydrothermal alteration mapping on a regional scale. Strong variations in electrical properties within hydrothermally altered rocks allow conductive clay-pyrite zones and resistive silicified zones to be mapped in detail by standard resistivity methods (Irvine and Smith, 1990). IP/resistivity and TEM, are effective in detecting buried conductive argillic alteration systems (e.g. at Hishikari, Japan, Ohui, New Zealand, and Mt Aubrey, New South Wales, Australia), which have very low resistivity. However, the precipitation of adularia and quartz causes an increase in apparent resistivity. Their subsurface extension may be difficult to delineate by electrical methods because of current channelling in adjacent low resistivity zones. The main IP anomalies can be associated with adularia-sericite zones, if they contain significant sulphide concentration (Allis, 1990).

Because of the extremely variable nature of the epithermal environment, geophysical surveys should be carefully undertaken. However, the nature of the epithermal environment is intimately related to the geological and geochemical signatures which have traditionally been the basis of epithermal gold exploration.

Geological prospecting and geochemical surveys are the most effective methods for establishing targets for project-scale epithermal gold exploration. The objective of these methods is to identify structurally complex zones where repeated faulting has prepared the ground for mineralizing hydrothermal fluids and where there is evidence of intense alteration.

Extensive areas of intense alteration are commonly associated with the upper levels of hydrothermal systems and can be up to 10 Km^2 wide (White and Hedenquist, 1990). For example, propylitic alteration (an assemblage of chlorite, pyrite, carbonate, montmorillonite and illite) can form halos hundreds of metres wide around the mineralized veins. Silicification can extend hundreds of metres above the ore horizon and often extends well below the bottom of the precious-metal horizon (Buchanan, 1981). In many districts, silicified and adularized vein walls form a thick selvage around the mineralized veins. They can be tens of meters wide, but usually are in the order of one meter or less.

Neither propylitic alteration nor silification/adularization are optimum ore guides, the former being too widespread, the latter being too narrow.

The alteration assemblage that is small enough to pin-point individual targets, extends well above the ore level and is genetically related to the ore formation process is represented by: alunite, sericite, illite, kaolinite, montmorillonite, or any of the kaolin clay minerals (Buchanan, 1981). This alteration forms a halo around and a cap above individual ore shoots and is absent below the lowest ore horizon.

At the hot spring level, zones of siliceous sinter and opal are mixed with or form a cap over alunite and kaolinite. Although these zones are often barren of gold mineralization, they are important because, beneath these layers, alteration assemblages of adularia and illite formed primarily by the loss of CO_2 caused by boiling. In the exploration for the Carlin-type deposits, jasperoids have been used as an important tool and the paragenesis outlined in Table 4.5 serves as a visual guide to jasperoid evaluation and sampling for geochemical analysis (Holland et al., 1988).

A search for alteration is an indirect approach to exploring for epithermal deposits. Only geochemistry offers a direct approach to locating mineralization. A good exploration guide for epithermal gold deposits is the geochemical association of gold, silver, arsenic, mercury, antimony, barium, tungsten, molybdenum and thallium.

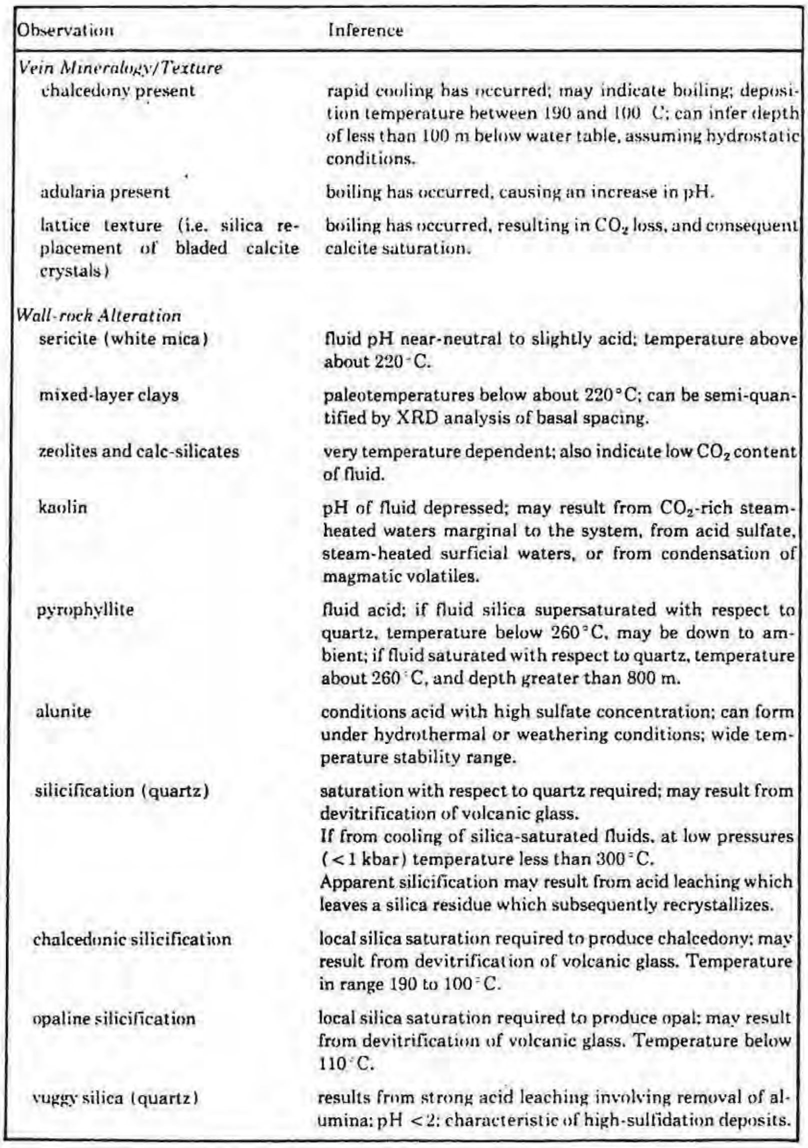
Favourable mineral association is represented by the following minerals (Mason, 1991b):

- Cinnabar; occurs at the top of the epithermal deposit; the broadest anomalies of Hg are due to its high degree of mobility in a supergene environment.
- Native sulphur; also occurs in the top level of the epithermal deposit, in association with sinters.
- Oripiment/Realgar; occurs in the top level, found in Carlin-type deposits.
- Stibnite; occurs in the top level; favourable particularly in Carlin-type deposits.
- Alunite; occurs in high to very high levels; it is evidence of K, AL metasomatism.
- Adularia; occurs lower in the epithermal deposit; it is one of the most important indicators of gold; it is associated with the boiling zone.
- Calcite; occurs at high levels.
- Barite; usually occurs at moderately high levels.

6.3 PROSPECT-SCALE EXPLORATION

In prospect-scale exploration, geological and geochemical studies are focussed on zones of enhanced permeability and mechanisms of ore deposition. Permeable zones can be recognized directly by an increased density of mineralized veins and the presence of hydrothermal eruption breccias. They can also be recognized from vein mineralogy and zoning of hydrothermal alteration assemblages (White and Hedenquist, 1990). An important aspect of structural preparation in epithermal gold systems is that the grade distribution and dimensions of any deposit will be intimately related to this structural preparation, because that is the key factor in opening the depositional site to mineralizing fluids (Mason, 1991b).

Deposition of the gold in an epithermal environment is caused by several mechanisms of which boiling, mixing of fluids and wall-rock reactions are the most common (see section 5.2.5). Boiling is the dominant process, controlling the temperature of hydrothermal fluids near the surface. Characteristic textural and mineralogic evidence that suggests boiling can be observed in the field (Table 6.1).



Observation	Inference
Vein Mineralogy/Texture	
chalcedony present	rapid cooling has occurred; may indicate boiling; deposition temperature between 190 and 100 °C; can infer depth of less than 100 m below water table, assuming hydrostatic conditions.
adularia present	boiling has occurred, causing an increase in pH.
lattice texture (i.e. silica replacement of bladed calcite crystals)	boiling has occurred, resulting in CO_2 loss, and consequent calcite saturation.
Wall-rock Alteration	
sericite (white mica)	fluid pH near-neutral to slightly acid; temperature above about 220 °C.
mixed-layer clays	paleotemperatures below about 220°C; can be semi-quantified by XRD analysis of basal spacing.
zeolites and calc-silicates	very temperature dependent; also indicate low CO_2 content of fluid.
kaolin	pH of fluid depressed; may result from CO_2-rich steam-heated waters marginal to the system, from acid sulfate, steam-heated surficial waters, or from condensation of magmatic volatiles.
pyrophyllite	fluid acid; if fluid silica supersaturated with respect to quartz, temperature below 260°C, may be down to ambient; if fluid saturated with respect to quartz, temperature about 260 °C, and depth greater than 800 m.
alunite	conditions acid with high sulfate concentration; can form under hydrothermal or weathering conditions; wide temperature stability range.
silicification (quartz)	saturation with respect to quartz required; may result from devitrification of volcanic glass. If from cooling of silica-saturated fluids, at low pressures (< 1 kbar) temperature less than 300 °C. Apparent silicification may result from acid leaching which leaves a silica residue which subsequently recrystallizes.
chalcedonic silicification	local silica saturation required to produce chalcedony; may result from devitrification of volcanic glass. Temperature in range 190 to 100 °C.
opaline silicification	local silica saturation required to produce opal; may result from devitrification of volcanic glass. Temperature below 110 °C.
vuggy silica (quartz)	results from strong acid leaching involving removal of alumina; pH < 2; characteristic of high-sulfidation deposits.

Table 6.1 *Mineralogy and texture as evidence for mineralizing conditions (after White and Hedenquist, 1990).*

Fluid mixing can also be an important mechanism of gold deposition. The main evidence for fluid mixing is alteration assemblages, represented by mixed layers of clays around the margins of the hydrothermal system, or by an acid overprint near its top (White and Hedenquist, 1990).

Fluid inclusion studies are of great importance in characterizing deposits. For example, if boiling and the mineralization are genetically related, then fluid-inclusion studies can provide evidence of boiling and the pressure and depth, calculated from inclusions trapped in a boiling system, define the level of the earth's crust which might host these deposits.

However, temporal and spatial relationships of boiling and ore deposition are highly variable in some of the epithermal deposits. For example, at Sunnyside, boiling did not occur during gold deposition, but was observed at the later,quartz-rhodochrosite-fluorite, stage (Casadevall and Ohmoto, 1977). At Carlin, boiling was also associated with the late acid leaching stage and was not responsible for quartz-pyrite-gold deposition (Radtke et al., 1980).

Thus, the presence of boiling may be useful for identifying epithermal systems, but cannot be used to predict with certainty whether or not precious-metal deposition has taken place (Bodnar et al., 1985).

Efficient prospect-scale exploration requires the careful integration of all available data, coupled with a good understanding of the geological and physico-chemical processes that occur in an epithermal environment.

6.4 SELECTION OF AREAS FOR REGIONAL EXPLORATION IN SERBIA (EASTERN YUGOSLAVIA), SOUTH-EAST EUROPE

Serbia is located in the eastern part of Yugoslavia (Fig. 6.1), which in turn forms part of South-east Europe (Fig. 6.2).

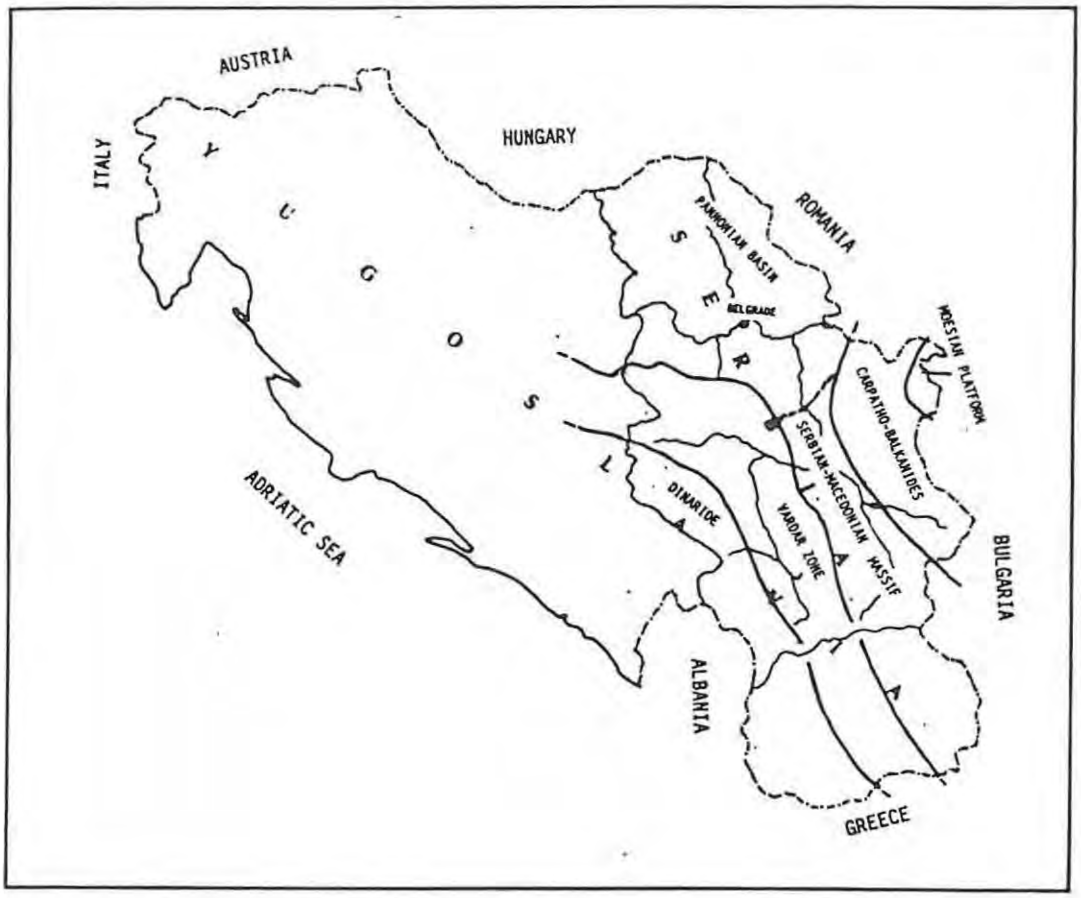


Fig. 6.1 *Location of Serbia and a scheme of its major tectonic units (modified and translated from Janković, 1990).*

The geologic evolution of Serbia in the post-Paleozoic period was closely related to the evolution of the Tethys ocean (its opening, stretching, closing) and tectonic movements between the Euroasian and Afroarabian plates (Dimitrijević, 1982; Janković, 1990). The terranes of Serbia belonged to the northern margins of Gondwanaland. Its fragmentation resulted in the formation of different microcontinents and islands in the Tethys ocean (e.g. Dinaride-Hellenide-Taurus, Rhodopian, Anatolian, etc.).

In the Late Jurassic period, an early closing phase of the Tethys commenced by subduction under the Eurasian plate (Dixon and Pereira, 1974; Channell and Horvath, 1976; Burchfiel, 1980). According to Jankovic (1990), subduction of Afroarabian and Indian oceanic crust under the Euroasian plate, following the closure of the Tethys ocean during Late Jurassic - Early Cretaceous, resulted in the formation of

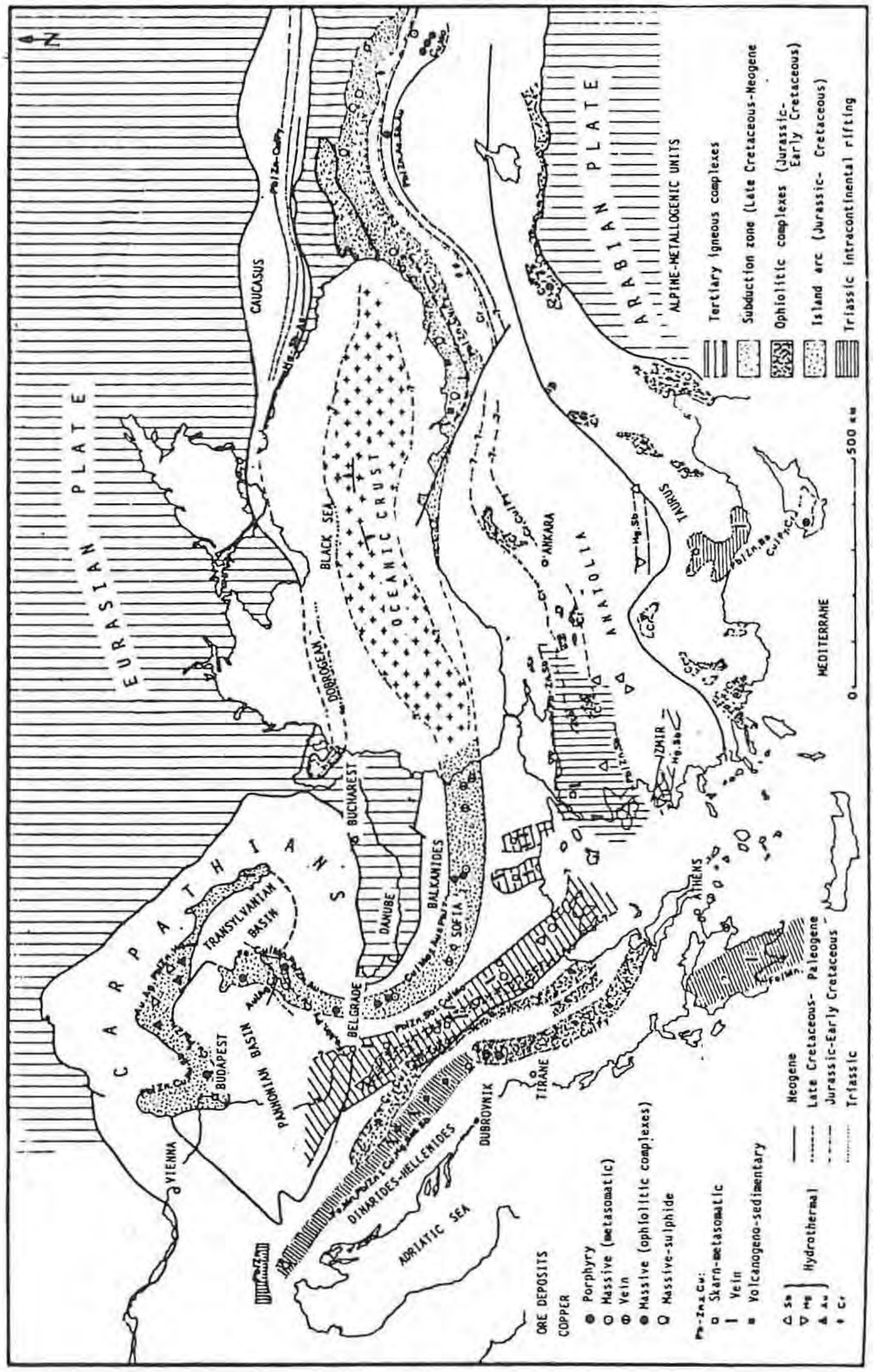


Fig. 6.2 *Main regional metallogenic units of the north-eastern Mediterranean area (modified and translated from Janković, 1990).*

a volcano-plutonic belt starting from the Carpatho-Ballkanides, through East Pontide (Turkey), Little Caucasus (Fig. 6.2) and Central Iran to Tibet and Sumatra. Therefore, Serbia is only a very small sector of a major global tectonic setting where subduction took place.

Several geotectonic units were developed in Serbia, such as the Moesian and the Pannonian platforms, the Dinarides, the Vardar Zone, the Serbian-Macedonian Massif and the Carpatho-Balkanides (Petković, 1981; Janković, 1990; Fig. 6.1) Based on the plate tectonic concept, ore deposits in Serbia are related to intra-continental rifting (within the Dinaride Zone), ophiolite complexes (in the Vardar Zone), subduction (in the Carpatho-Ballianides) and continent-continent collision (in part of the Vardar Zone and within the Serbian-Macedonian province) (Janković, 1990; Fig. 6.2).

The most favourable tectonic setting for exploration of porphyry-related and geothermal-related epithermal deposits is the Carpatho-Balkanide metallogenic province. The province includes a few metallogenic zones of which the most important are the Ridanj-Krepoljin in the west (bordering the Serbian-Macedonian province) and the Bor zone in the central part (Figs. 6.3 and 6.4). These zones host numerous ore deposits, the most important of which are porphyry and massive-sulphide copper, with significant amounts of gold and silver locally, and polymetallic deposits (Pb-Zn, Cu) associated with host-graben structures formed above the subducted oceanic floor (Janković, 1990). Majdanpek and Bor are the most significant porphyry and massive-sulphide copper deposits within the Bor metallogenic zone which have been exploited for more than 80 years for their copper, with gold as a by-product. They were also exploited during Roman times.

Bor is a massive-sulphide enargite-bearing deposit hosted in andesite (hornblende-biotite) and its pyroclastite, with minor amounts of dacite (Janković, 1990) and is similar to the Lepanto deposit (see section 4.1), in the western Pacific, and Recsk, in Hungary (Sillitoe, 1983).

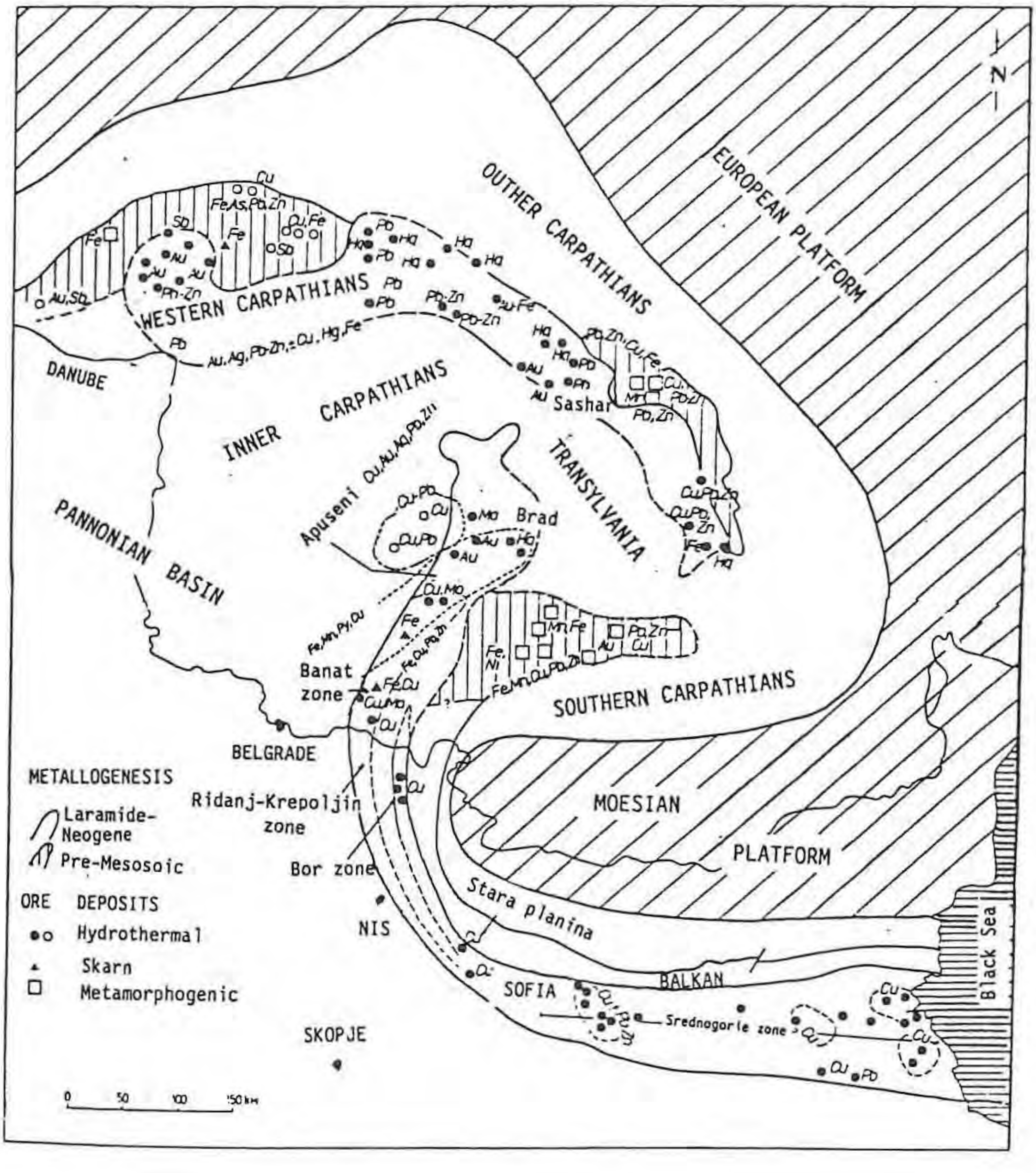


Fig. 6.3 *The Carpatho-Balkanide province : The major regional units (translated from Janković, 1990).*

Magmatic activity in the Bor zone (80km in length and 20km in width) began in Turonian time and lasted intermittently until the Paleogene (mainly between 90 and 60 my) following the subduction of the Tethys ocean (Grubić, 1974; Boccaletti et al., 1974; Janković, 1990).

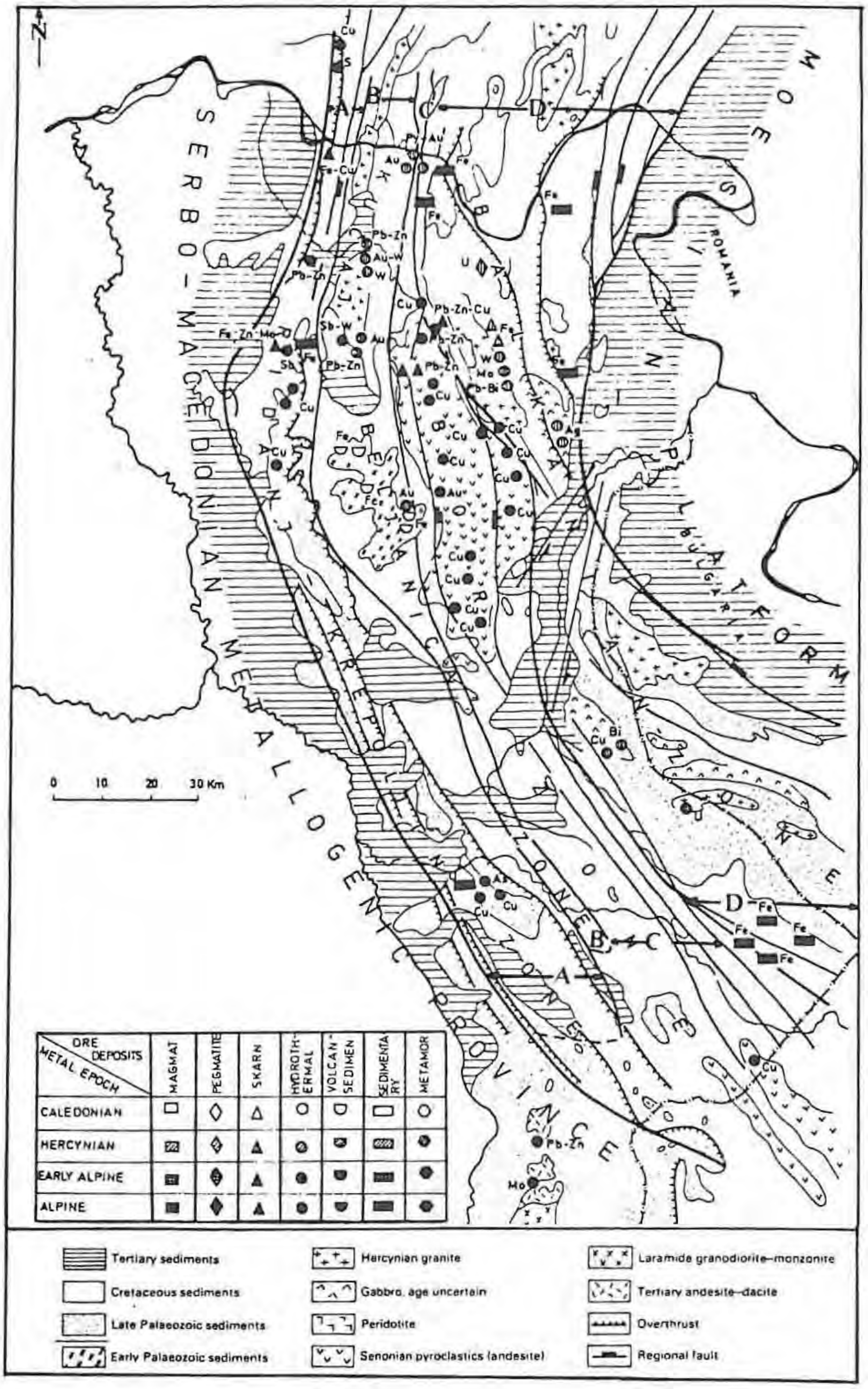


Fig. 6.4 *Structural-metallogenic zones in Eastern Serbia. A : Ridanj-Krepoljin zone; B : Neresnica-Beljanica zone; C : Bor zone; D : Balkan zone (modified after Janković, 1982).*

Hypabyssal rocks occur as multistage composite complexes ranging in composition from syenite, monzonite and granodiorite to quartz diorite, diorite and gabbro. The petrochemical composition of intrusions is variable, ranging from saturated to oversaturated with respect to silica and from calc-alkaline to alkaline (Janković, 1990). Porphyry Cu(-Mo-Au) deposits in the Bor zone (Majdanpek, Veliki Krivelj, Borska Reka) belong to the regional metallogenic belt hosting numerous Cu(-Mo-Au) porphyry deposits, including the Recsk Hungary; Rosia Poieni and Moldova Noua in Romania; Medet, Assarel and Elatsite in Bulgaria; Kadzaran in Little Caucasus; Sarh Chesmeh in Iran; Seindak in Pakistan and Yulong in Tibet.

Most of these deposits formed in the Upper Cretaceous/Lower Eocene period (including those in Serbia) and some of them are of Eocene, Oligocene and Miocene age. They are related to quartz diorite, granodiorite, monzonite and subvolcanic andesite (Janković, 1990).

At present gold is being exploited in Serbia only as a by-product of porphyry and massive-sulphide Cu deposits in the Bor zone. Economically significant epithermal gold deposits have not yet been discovered in Serbia. However, there are many hydrothermal occurrences and a few deposits of gold in the Carpatho-Balkanides and the Sebian-Macedonian metallogenic province, some of which may be of epithermal origin. Detailed geochemical studies are not sufficient at this stage, to classify these deposits further (Kondžulović, pers. com.).

The most important gold deposits are located in the Carpatho-Balkanide province, namelyBlagojev Kamen in the Neresnica-Beljanica zone (Fig. 6.4). Native gold occurs with scheelite in quartz veins cutting or parallel to the foliation of early Paleozoic green schists (Petković, 1989). The grade of gold ranges in average between 8 - 15 g/t, but can be up to 100 g/t locally (Janković, 1990).

High contents of gold also occur in hydrothermal replacement Pb-Zn-Au deposits in Kucajna, within the Ridanj-Krepoljin zone in the Carpatho-Balkanide province. Deposits are hosted in the Jurassic and Cretaceous limestones in contact with dacito-andesite.

Along the Bor zone hydrothermal, quartz veins are present with irregularly dispersed gold and silver mineralization in association with antimony, barium, tungsten, molybdenum, sometimes with arsenic and mercury and thallium (Janković, 1990). However, these occurrences have not yet been found to be of economic importance.

It is important to mention that volcanic-hosted gold deposits occur in some other segments of the Tethyan Euroasian metallogenic belt in analogous geotectonic settings sauch as the Carpatho-Balkanides in Eastern Serbia. In Rumania, gold and gold-based metal ores in the Sashar-Valea-Rosie district are related to Neogene volcanism, predominantly with andesite characteristics (Ivanovici and Borocos, 1982). The alteration patterns and mineralization in this district show the characteristics of adularia-sericite type deposits.

Gold-silver telluride deposits in the Brad-Sacarimb district of Rumania, which include base-metal and porphyry copper deposits, are related to Neogene volcanism, which is hornblende-andesitic, quartz andesitic and rhyodacitic in character (Ivanovici and Borocos, 1982). These deposits show mainly the characteristics of acid-sulphate-type deposits. They were exploited during Daco-Roman times.

In the East Pontides, Turkey, the Akarsen porphyry copper deposit contains significant grades of gold. The deposit is associated with Upper Cretaceous dacitic volcanism. Genetically, the acid-sulphate type of gold mineralization is probably related to this porphyry system (Özgür, 1991).

Sediment-hosted disseminated Carlin-type gold deposits can also be expected to occur in the Carpatho-Balkanide province in Serbia. A particularly favourable area is the western Ridanj-Krepoljin zone and the contact with the Serbian-Macedonian province where a thrust belt developed (Fig. 6.4) during the Upper Cretaceous/Lower Eocene period (Boccaletti et al., 1974; Janković, 1990). The Ridanj-Krepoljin zone, which is only about 20km wide, is characterized by deep faulting and Miocene-Oligocene dacito-andesite rocks extending southward for several dozen kilometres, intruding Paleozoic phylitic basement and overlying Jurassic and Cretaceous limestones. Plutonic rocks of granodiorite magma are known in Rumania, but in the Ridanj-Krepoljin zone they have not been reported so far. However, there are indications that they exist at greater depth (Janković, 1990).

Several skarn lead-zinc occurrences and deposits, small stibnite occurrences in limestones, hydrothermal Cu deposits and Sb-W veins occur in the zone. The occurrence of tungsten deposits can be a good indicator of Carlin-type deposits.

Of importance is the Kucajna Pb-Zn (Ag,Au) deposit in limestone which can be considered to be one of the criteria which can be used in exploration for the Carlin-type gold deposits. It is a hydrothermal replacement disseminated-to-massive its Pb-Zn deposit which, apart from Pb and Zn mineralization, was exploited for its extremely high Au and Ag content. Mining took place during Roman times and from 1862 up to 1894, when it was closed because of technical problems. However, some recent exploration programmes have failed to discover further significant mineralization in the area.

Mineralization in the Kucajna deposit is represented by pyrite, sphalerite, galena, tetrahedrite, chalcopyrite, arsenopyrite, native gold and silver and is located in Jurassic and Cretaceous limestones, in close contact with andesitic rocks. The limestones overly the Paleozoic phyllitic basement and are themselves overlain by thrusted Permian Red sandstones, and locally by a flysch sequence. The geotectonic setting and lithology of the Kucajna deposit is very similar to that of southern Tuscany, Italy, where Carlin-type gold deposits have recently been discovered (Tanelli et al., 1991). The mineralization in Tuscany consists of jasperoid masses replacing carbonate rocks typically at the contact with overlying flysch of the allochtonous Ligurian sequence. The mineralogy is represented by pyrite, galena, sphalerite, tetrahedrite, arsenopyrite, cinnabar, native Au and Ag. It is related to Pliocene-Quaternary magmatism (Tanelli et al., 1991).

Consideration of the most important guidelines (tectonic, igneous and structural characteristics, known mineralization, etc.) discussed above, suggests that the Carpatho-Balkanide province and its contact with the Serbian- Macedonian province represent the most favourable areas for epithermal gold deposits in Serbia. It also indicates the most promising terranes within the province for exploration from regional to prospect scale (sections 6.2 and 6.3).

7.0 CONCLUSIONS

Epithermal gold deposits form in the upper few kilometres of large-state hydrothermal convection systems in the earth's crust and at temperatures between 50°C and 350°C. They can be connected with copper-molybdenum porphyry systems centred on felsic intrusive centres or with geothermal systems related to volcanic centres and calderas (adularia-sericite type, acid-sulphate type and disseminated, replacement Carlin-type gold deposits).

The essential components which allow a common genetic framework to be established for all epithermal gold deposits include the following:

- A heat source to drive a hydrothermal system in the upper crust. The most favourable tectonic setting for porphyry and geothermal systems are convergent plate boundaries (Andean magmatic arcs, Pacific island arcs, back-arc extensional cratonic terranes). In most cases, the high heat flows are related to magmatism which represents I-type, magnetite-series magmas generated from hornblende melting.

- Source and composition of hydrothermal fluids. A meteoritic water component is predominant throughout the ore-forming stages in volcanic-hosted (adularia-sericite and acid-sulphate types) gold deposits, in the later stage of porphyry related epithermal deposits and in the intermediate to later stages of the Carlin-type systems. The ore fluids can have a varying number of magmatic components (porphyry-related epithermal systems and some of the adularia- sericite type deposits).
  The formation of epithermal gold deposits requires near-neutral (adularia-sericite and some of the Carlin-type deposits) to acid (acid-sulphate type and most of theporphyry-related epithermal systems) fluids of low salinity and high gas content (H_2S; CO_2).

- Source of metals. In most epithermal gold deposits, gold and other metals are derived by leaching from crustal rocks. Basaltic magmas emplaced into the lower or middle crust are likely to be the source of gold in the Carlin-type deposits, rather than leaching from crustal rocks.

- High permeability structures (usually deep faults and fractures) which allow fluid convection and metal deposition.

Gold and copper in the porphyry system were probably introduced as chloride complexes and precipitated with the K-silicate assemblage when high temperature chloride- and sulphur-bearing fluids escaped into cooler wallrocks.

In epithermal environments, including the upper part of porphyry Cu (-Mo)-Au systems at temperatures up to 300°C, gold is transported as a gold bisulphide complex. At slightly higher temperatures (between 300°-350°C) the complex $Au\ HS^{o}$ is more concentrated, and therefore more important than $Au\ (HS)_2^-$.

Gold deposition in epithermal environments is caused by changes in pH, temperature, pressure, redox potential and the activities of complexing ligands. Boiling is the mechanism responsible for most of the adularia-sericite type deposits. Gold deposition in these environments is accompanied by banded silica, calcite, adularia and pyrite. In acid-sulphate type deposits, gold, covellite and enargite precipitate throught the mixing of gold-bearing liquid remaining after boiling and acid sulphate waters produced by the oxidation of the boiled gas condensates. Reaction between sulphuric acid and wall rocks produces a quartz + kaolinite + alunite alteration assemblage in these deposits. Most of the gold and pyrite in the Carlin-type deposit precipitates by mixing of meteoritic waters and basin brines and reaction with the wall rocks. Dilute meteoritic waters in the late stage of deposit evolution are responsible for the rest of gold deposition.

Consideration of genetic models of epithermal gold deposits, including tectonic, igneous and structural settings and the physico-chemical characteristics of hydrothermal fluid systems, indicates the favourable areas for regional to prospect-scale exploration.

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