EPITHERMAL PRECIOUS METAL DEPOSITS : PHYSICOCHEMICAL CONSTRAINTS, CLASSIFICATION CHARACTERISTICS, AND EXPLORATION GUIDELINES

by

Donald A McIver

Dissertation submitted in partial
fulfilment of the requirements for
the degree of Master of Science
(Economic and Exploration Geology)
at Rhodes University, Grahamstown.

This dissertation was prepared in
accordance with specifications
laid down by the univeristy and
was completed within a period
of ten weeks full-time study.

January, 1997

EPITHERMAL PRECIOUS METAL DEPOSITS : PHYSICOCHEMICAL CONSTRAINTS, CLASSIFICATION CHARACTERISTICS, AND EXPLORATION GUIDELINES

by

Donald A McIver

Rhodes University
1997

ABSTRACT

Epithermal deposits include a broad range of precious metal, base metal, mercury, and stibnite deposits. These deposits exhibit a low temperature of formation (180-280°C) at pressures of less than a few hundred bars (equivalent to depths of 1.5 - 2.0km). Epithermal gold deposits are the product of large-scale hydrothermal systems which mostly occur in convergent plate margin settings. Associated volcanism is largely of andesitic arc (calcalkaline to alkaline), or rhyolitic back-arc type. Porphyry Cu-Mo-Au deposits form deeper in the same systems. Genetic processes within individual deposits take place in an extremely complex manner. The resultant mineral associations, alteration styles and metal deposition patterns are even more complicated.

Many attempts have been made to classify epithermal deposits based on mineralogy and alteration, host rocks, deposit form, genetic models, and standard deposits. For the explorationist, the most useful classification schemes should be brief, simple, descriptive, observationally based, and informative. Ultimately, two distinct styles of epithermal gold deposits are readily recognised: high-sulphidation, acid sulphate and low-sulphidation, adularia-sericite types. The terms high-sulphidation (HS) and low-sulphidation (LS) are based on the sulphidation state of associated sulphide minerals, which, along with characteristic hydrothermal alteration, reflect fundamental chemical differences in the epithermal environment. High-sulphidation-type deposits form in the root zones of volcanic domes from acid waters that contain residual magmatic volatiles. The low-sulphidation-type deposits form in geothermal systems where surficial waters mix with deeper, heated saline waters in a lateral flow regime, where neutral to weakly acidic, alkali chloride waters are dominant. The HS/LS classification, combined with a simple description of the form of the deposit, conveys a large amount of information on mineralogy, alteration, and spatial characteristics of the mineralisation, and allows inferences to be drawn regarding likely regional controls, and the characteristics of the ore-forming fluids.

The modern understanding of these environments allows us to quite effectively identify the most probable foci of mineral deposition in any given district. Current knowledge of these deposits has been derived from studies of active geothermal systems. Through comparison with alteration zones within these systems, the exploration geologist may determine the potential distribution and types of ore in a fossil geothermal system. Alteration zoning specifically can be used as a guide towards the most prospective part of the system. Epithermal gold deposits

of both HS- and LS-styles are nevertheless profoundly difficult exploration targets. Successful exploration must rely on the integration of a variety of exploration techniques, guided by an understanding of the characteristics of the deposits and the processes that form them. There are no simple formulae for success in epithermal exploration: what works best must be determined for each terrain and each prospect.

On a regional scale tectonic, igneous and structural settings can be used, together with assessment of the depth of erosion, to select areas for project area scale exploration. Integrated geological-geophysical interpretation derived from airborne geophysics provides a basis of targeting potential ore environments for follow-up. Geology, geochemistry and surface geophysics localise mineral concentrations within these target areas.

TABLE OF CONTENTS

LIST OF FIGURES

LIST OF TABLES

EPITHERMAL PRECIOUS METAL DEPOSITS: PHYSICOCHEMICAL CONSTRAINTS, CLASSIFICATION CHARACTERISTICS, AND EXPLORATION GUIDELINES

INTRODUCTION

The term "epithermal" was defined by Lindgren (1922; 1933; in Heald *et al.*, 1987) to include a broad range of precious metal (± tellurides or selenides), base metal, mercury, and stibnite deposits which he believed formed from aqueous fluids charged with igneous emanations at low temperature (less than 200°C) and moderate pressure. As more data have become available on the temperature of mineralising fluids, and also on the depth of mineralisation based on both boiling-depth curves and stratigraphic evidence for the thickness of the cover during mineralisation, it has become evident that most epithermal deposits form within a temperature range of 180 - 280°C or 300°C, and at depths rarely exceeding 1.5 or at most 2 km (White & Hedenquist, 1990; in Mitchell & Leach, 1991). These deposits are thus considered to be those which exhibit a low temperature of formation and pressures of less than a few hundred bars, with characteristic hydrothermal alteration (Bonham, 1986; Hayba *et al.*, 1985; Heald *et al.*, 1987; Hedenquist, 1987; in White *et al.*, 1995), and textures (Berger & Eimon, 1983; in White *et al.*, 1995).

Epithermal gold deposits are the product of large-scale hydrothermal systems in volcanic terrains (Figure 1). Their essential ingredients are a magmatic heat source in the upper few kilometres of the crust, a source of groundwater, metal and reduced sulphur, and zones of brittle fracture. These ingredients have been available throughout crustal history so there is no restriction on age, only on preservation (Henley, 1991). They mostly occur in convergent plate margin settings, the same environment where high temperature, near-surface geothermal activity is concentrated today (Figure 2). The associated volcanism is largely either of andesitic arc (calcalkaline to alkaline), or rhyolitic back-arc type (White & Hedenquist, 1990; Henley, 1991). Other types of hydrothermal mineralisation such as porphyry Cu-Mo-Au deposits form deeper in the same or similar systems (Henley & Ellis, 1983; Sillitoe, 1988; in Clarke, 1991).

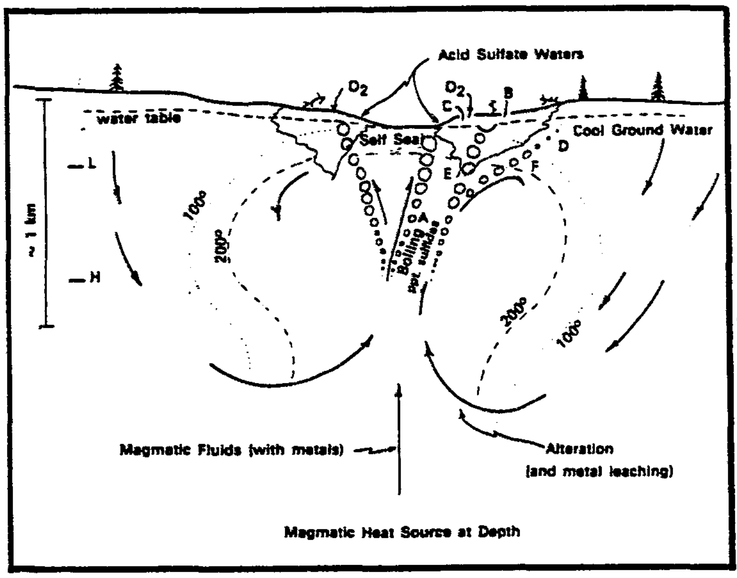


Figure 1. Schematic diagram of a boiling hydrothermal system based in part on White *et al.*, (1971), Henley & Ellis (1983), Berger & Eimon (1983), and Steven & Eaton (1975). Ascending hot waters begin to boil (indicated by circles) at the depth labelled H if pressure is hydrostatic, or at some shallower depth between H and L if pressure exceeds hydrostatic but is less than lithostatic. Boiling and various mixtures of boiled hot water and gases with near-surface ground water and atmospheric oxygen produce hotwater compositions and ore-forming environments labeled by letters A through F (after Reed & Spycher, 1985).

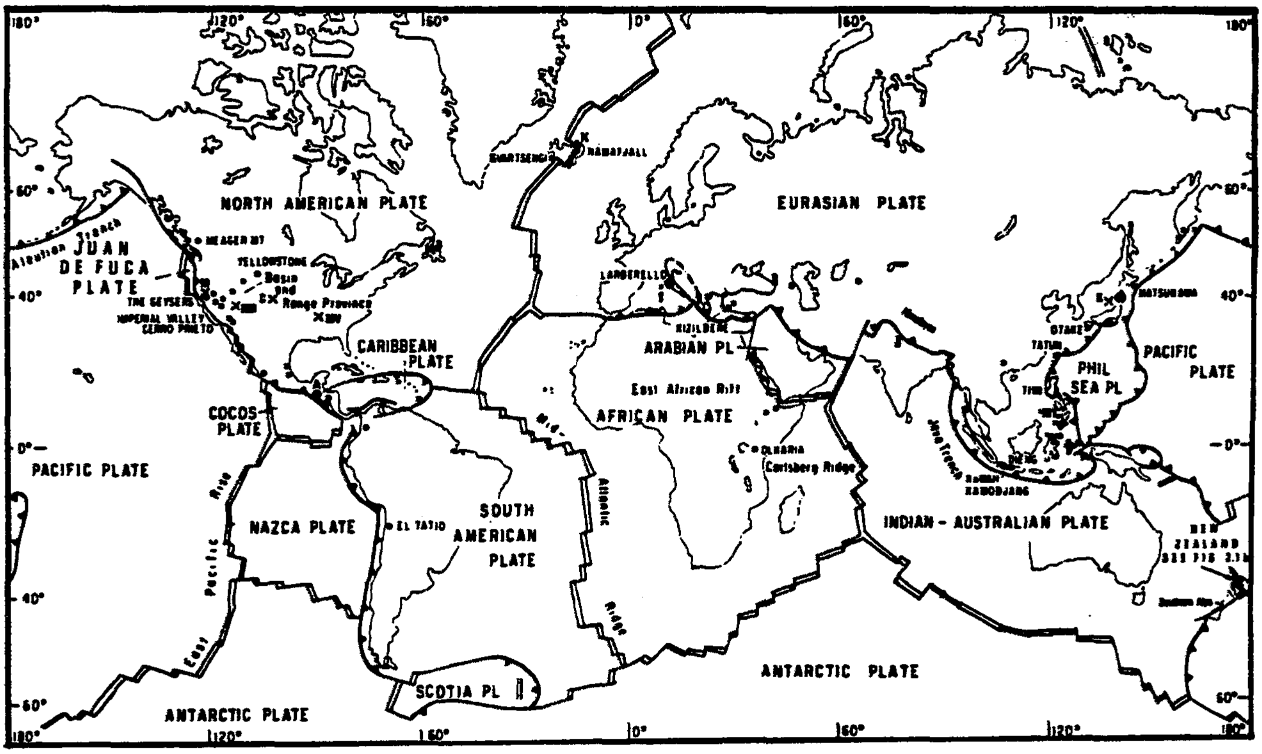


Figure 2. Lithospheric plate boundaries provide the framework for the global distribution of major geothermal systems. Symbols: ● geothermal systems; A (Ahuachapan/El Salvador), K (Krafla/Iceland), MT (Mahio-Tongonan, Phillipines), ✖ mineral deposit localities; M (McLaughlin, California), RM (Round Mountain, Nevada), C (Creede, Colorado,) MV (Mississippi Valley), K (Kuroko District/Japan). (Adapted with permission from Rybach & Muffler, 1981; in Henley *et al.*, 1984).

Epithermal precious-metal deposits are usually of vein type, are associated with either propylitic (+ adularia) or advanced argillic alteration, and cut predominantly mid- to Cenozoic volcanic sequences (Sillitoe, 1977). Deposit-type models are useful for focusing attention on changes of mineral assemblages and physical characteristics of the deposits with depth, and form a framework within which to discuss geochemical variations. It must be kept in mind however, that in individual deposits, (as in present-day geothermal systems), genetic processes take place in an extremely complex manner, in which the effects of vertical and lateral variations in temperature due to such effects as positions of the water table, boiling zones, ratio of meteoric to magmatic water, and channels of fluid movement occur. The resultant mineral associations, alteration styles and metal deposition patterns are made even more complex by changes in these effects with time (Silberman & Berger, 1985). For example, a deposit that starts out with dominantly magmatic hypo- to mesothermal fluids (below the zone of boiling) is likely to be inundated by meteoric waters later in its history and to end its life as epithermal (above the zone of boiling). It should also be remembered that individual deposits are ordinarily formed during several stages of mineralisation (Elston, 1994).

Virtually all economic concentrations of epithermal gold occur in magmatic arcs, particularly Tertiary-aged island arcs, and are epigenetic with respect to the enclosing host rocks (Mitchell & Leach, 1991). They are recognised in for example the Aleutians, Kuriles, Japan, Taiwan, the Philippines, and Eastern Indonesia, and the Islands of Papua New Guinea, Palau and Fiji in the Western Pacific (Figure 3); in the Western Burma arc bordering the Indian Ocean; and in the Caribbean. Epithermal deposits are also found in continental margin arcs in mainland New Guinea, Thailand, Borneo, Sumatra, Vietnam, Eastern China, New Zealand, and the Andes. In Eastern Australia, Eastern Europe and the Trans-Himalayas epithermal gold occurs in former magmatic arcs which are now part of a continent (Mitchell & Leach, 1991).

The youthfulness of the principal gold deposits, irrespective of their depths of emplacement, is attributed to the rapid uplift and erosion rates that characterises most island arcs. In these active volcanic terrains, erosion rates as high as 0.75 mm yr^{-1} have been estimated (Ruxton & McDougall, 1967; in Sillitoe, 1988), which imply stripping of 1.5 km of volcanic cover in 2 Ma.

The most basic characteristics of any ore deposit are the form of the orebody and its mineralogy, the textures of ore and gangue minerals, and alteration zoning (White &

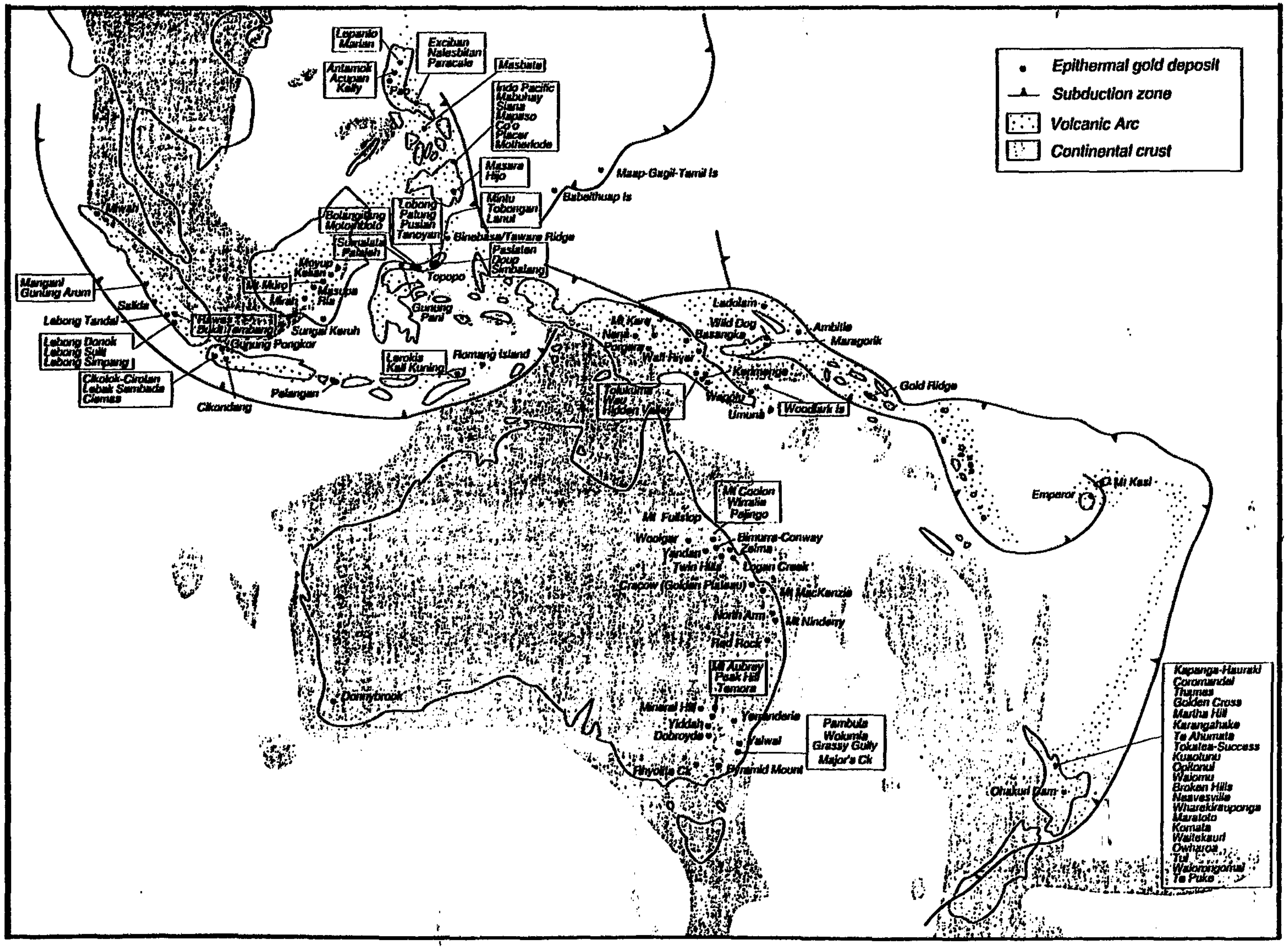


Figure 3. Map of the Southwest Pacific, showing the location of epithermal deposits in relation to present-day subduction and volcanic arcs, and continental crust (after White *et al.*, 1995).

Hedenquist, 1995). Following Bonham, (1986; in Sillitoe, 1988) and Heald *et al.,* (1987; in Sillitoe, 1988), the epithermal gold deposits may be subdivided into: high-sulphidation, acid sulphate and low-sulphidation, adularia-sericite types. Comparison of observations (Sillitoe, 1977; 1993; Buchanan, 1981; Heald *et al.,* 1987; and White *et al.,* 1995; in White & Hedenquist, 1995) of the low- and high-sulphidation deposits highlights the considerable overlap in characteristics (to be discussed in 2.2).

The two types of deposits appear to form under similar pressure-temperature conditions but in different geological and geochemical environments in ancient geothermal systems. The acid-sulphate-type deposit forms in the root zones of volcanic domes from acid waters that contain residual magmatic volatiles. The adularia-sericite-type deposit forms in geothermal systems where surficial waters mix with deeper, heated saline waters in a lateral flow regime, high above and probably offset from a heat source at depth; neutral to weakly acidic, alkali chloride waters are dominant (Heald *et al.*, 1987).

Mineral deposits need a metal source, a mechanism of metal transportation to a place of deposition, a mechanism of precipitation, and sufficient time for the transportation and deposition mechanisms to operate so that economic concentrations of metals can accumulate (Silberman & Berger, 1985). These issues and associated physicochemical topics are considered in Part A of the dissertation. Part B investigates the various classification systems of epithermal precious metal deposits. In reviewing the characteristics of these deposits, the discussion is limited to two dominant styles of epithermal precious metal occurrence *ie.* the high-sulphidation and low-sulphidation types. This regrettably results in the omission of several major epithermal precious metal districts, like for example the sediment-hosted (typically carbonate replacement) disseminated fine-gold lode deposits that occur at Carlin, Jerritt Canyon, Pinson, *etc.* These are known as Carlin-type deposits and occur in marine sedimentary sequences where magmatic activity, in response to regional extension tectonics, affects the middle to upper crust, but where there is no geologic evidence that large volcanic centres were formed in the immediate vicinity of the deposits. Mineralisation in general occurs at greater depth than for epithermal deposits in volcanic terrains (Berger & Henley, 1988). These deposits differ also in host lithologies and regional setting. Understanding of their origin suggests that they are genetically dissimilar to most epithermal deposits, and form outside the normally recognised epithermal environment (Berger & Henley, 1989; Berger & Bagby, 1990; in White & Hedenquist, 1990).

Another unusual epithermal style that will be omitted is that which forms in Eastern Mindanao. Here gold occurs in pyritic clays, in fragments and blocks of vein quartz and opaline to jasperoidal silica and as free detrital gold within Upper Miocene sediments that include debris flows with blocks of andesite porphyry, basalt, mudstone and coal that commonly underlie or overlie mineralised andesitic rocks. The deposits, which are along strike from both low-sulphidation vein gold in Mindanao and the "exhalative" base-metal-gold deposits on Samar, appear to have resulted from the reworking of pre-existing epithermal deposits within a protected shallow marine or near-shore environment (Mitchell, 1992).

Detailed discussion on gold-telluride deposits too will be omitted. These are characterised by the association of gold-telluride mineral with fluorite, carbonate mineral, and adularia, and are commonly associated with alkaline volcanic rocks and breccia pipes. Examples of gold-telluride deposits in the United States include Cripple Creek, Colorado (Thompson *et al.*, 1985; in Ericksen, 1988), Golden Sunlight, Montana (Porter and Ripley, 1985; in Ericksen, 1988), and Jamestown, Colorado (Nash and Cunningham, 1973; in Ericksen, 1988).

Part C considers guidelines towards developing an effective exploration strategy for implementation when actively prospecting for epithermal precious metal deposits. Issues are investigated from the regional reconnaissance scale at which prospective regions are identified, to the reconnaissance and follow-up stages.

PART A: PHYSICOCHEMICAL CONSTRAINTS ON EPITHERMAL MINERALISING SYSTEMS

1.1 GEOTHERMAL SYSTEMS

White *et al.*, (1971; in Silberman & Berger, 1985) define a geothermal system as a source of heat within the earth's crust, be that from magmatic intrusion, or regional heat flow, and the rocks and water affected by that heat. The geothermal system includes upwelling hot fluids and marginal convective downflowing, cold recharge waters. Henley (1985) shows typical schematic cross sections of geothermal systems hosted in silicic volcanic terrains (Figure 4), common on continental margin areas, such as Yellowstone and Broadlands and andesitic stratovolcanic terrains (Figure 5), common in island arc areas, such as Matsao, Taiwan. The surface expressions of geothermal systems are hot springs, fumaroles and other indications of hydrothermal activity such as altered rocks (Silberman & Berger, 1985).

It has long been recognised that active geothermal systems in volcanic terrains are the present-day equivalents of those ancient systems responsible for gold- and silver-bearing mineralisation in epithermal mining districts (Lindgren, 1933; White, 1955; 1981; Henley & Ellis, 1983; in Henley, 1991). The analogy between epithermal ore deposits and geothermal systems is supported by the occurrence of ore-grade concentrations of Au and Ag and other associated elements (*e.g.* As, Sb, Hg, Tl, W) in surface discharge material from several active geothermal systems, such as Steamboat Springs, Nevada, Ohaaki-Broadlands, and Waiotapu, New Zealand. Features characteristic of geothermal systems, such as siliceous sinter and hydrothermal explosion breccias are found in some epithermal deposits, and can be part of the ore (Barrett, 1985; Silberman *et al.*, 1979; Wallace, 1980; 1984; in Silberman & Berger, 1985).

The advantage in studying geothermal systems hosted within active volcanic-arcs is that discrete fluids can be sampled, allowing an examination of 'time-slices' through the history of hydrothermal activity (Hedenquist & Lowenstern, 1994), thus providing some evidence for the processes that occurred in ancient hydrothermal systems (Mitchell, 1992). By the same token the study of fossil hydrothermal systems, which have been deeply dissected by erosion, offers information on the deep characteristics of active geothermal systems which is otherwise quite unattainable (Henley & Ellis, 1983).

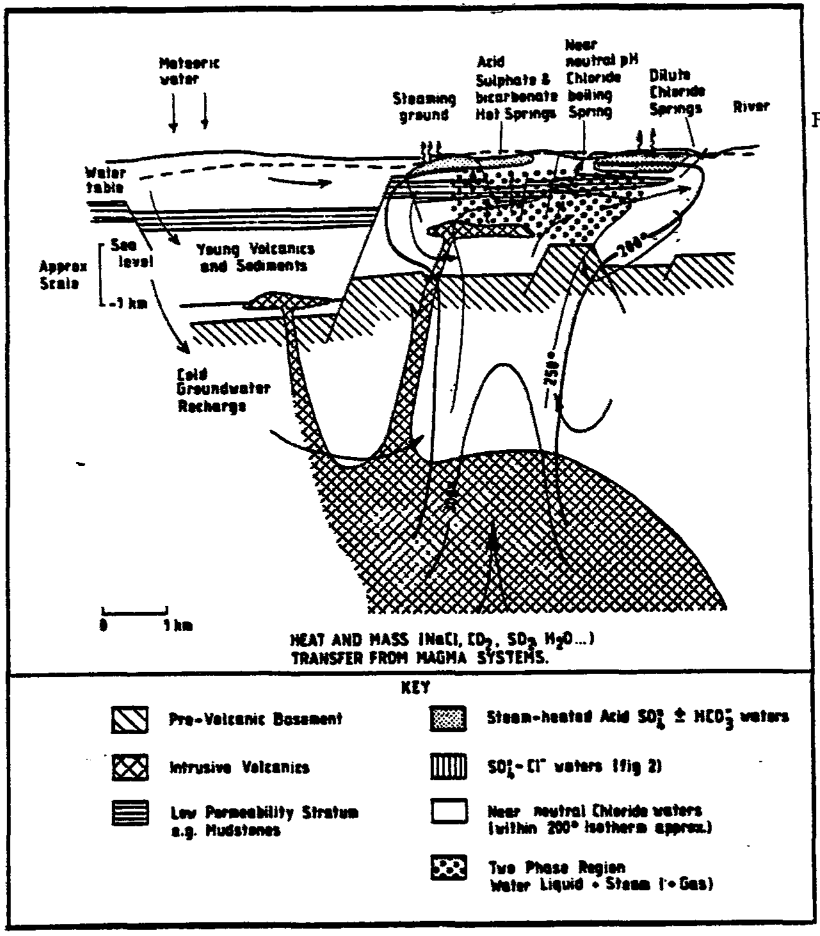


Figure 4. Schema of the main features of a geothermal system typical of those in silicic volcanic terranes. The system is supplied by ground-water in this case derived from meteoric water. Heat, together with some gases, chloride, water and some other solutes, is assumed to be supplied by a deeply buried magmatic system and results in a convecting column of near-neutral pH chloride water with two phase conditions in the upper part of the system. Steam-separation processes give rise to fumaroles and steam absorption by groundwater, with oxidation of H_2S at the water table, which gives rise to isotopically enriched steam-heated acid sulphate and bicarbonate waters. Mixing may occur between the deeper chloride waters, steam-heated waters and fresh groundwater to give a range of hybrid waters. Outflows from the deep chloride system occur either as boiling alkaline springs often associated with silica terraces, or after mixing with cold groundwaters, as near-neutral pH, relatively dilute, chloride springs (after Henley *et al.*, 1984).

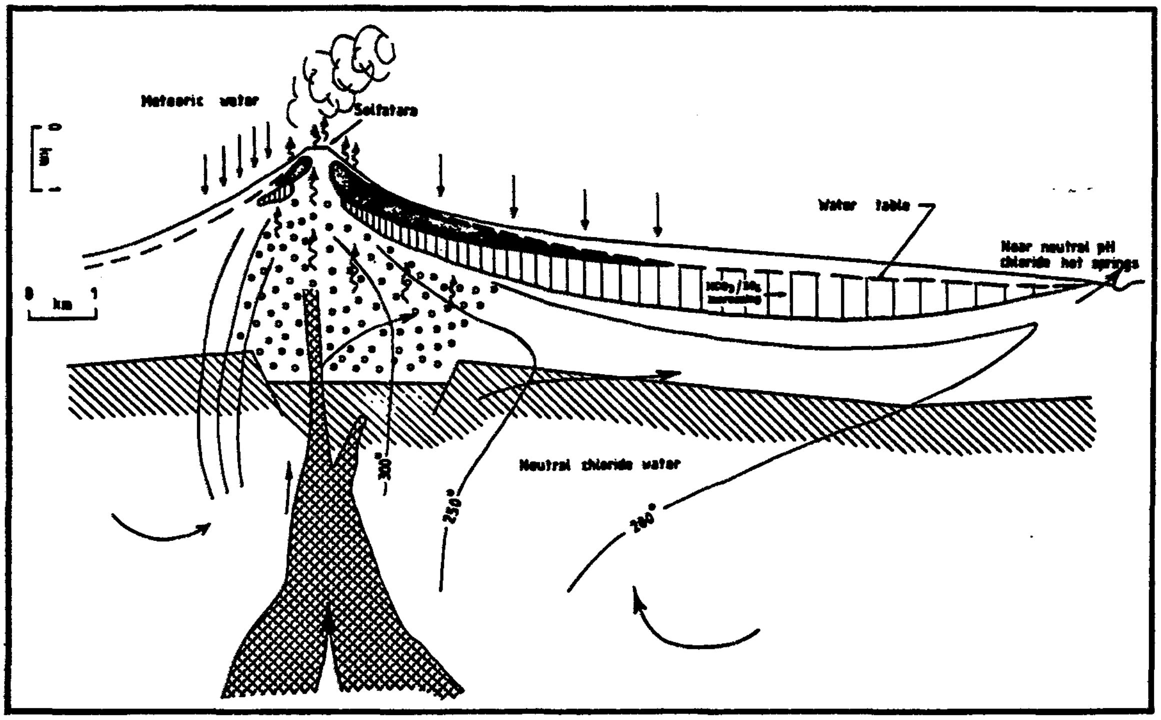


Figure 5. Schema of a geothermal system typical of active island-arc andesite volcanoes. Lower permanent water tables in tropical regions and the high relief of the volcanic structures result in a scarcity of chloride water discharges except at some distance from the upflow center. The latter may be revealed by fumaroles, intense rock alteration and steam-heated, often perched, aquifers. Near-surface condensation of volcanic gases and oxidation result in acid sulphate waters in the core of the volcano and an acid crater lake may also form (after Henley *et al.*, 1984).

1.2 HYDROLOGIC ENVIRONMENT WITHIN THE UPPER CRUST

In epithermal deposits the hydrology of the parent hydrothermal system plays a crucial role in determining the sites of focused fluid flow essential to produce economic concentrations of metals. The hydrology also determines the patterns of hydrothermal alteration associated with the ore deposits (Hanaoka, 1980; Taguchi *et al.,* 1986; Allis, 1990; Reyes, 1990; Simmons, 1991; in White *et al.*, 1995).

Hydrothermal systems powered by magmatic intrusions dominate fluid movement in the upper crust, and are responsible for convecting a large proportion of Earth's heat to the surface. At the same time, the fluids transport metals, forming the single most important class of ore deposits, those of the hydrothermal category. For an epithermal ore deposit to form, the flow of metal-bearing fluid must be focused and coupled with a precipitation mechanism operating in a restricted space, details of which may vary from deposit to deposit (Hedenquist & Lowenstern, 1994). As hot water rises toward the surface, it will interact with other magmatic volatile-hosting waters, forming a hydrothermal plume. The pressure imposed on the rising plume by overlying fluid decreases, and it eventually reaches a level at which a vapour phase separates and migrates to the surface independently - *i.e.* boiling occurs (Henley *et al.*, 1984).

The movement of fluids in such a system has been modelled by Henley & McNabb (1978; in Henley *et al.*, 1984). As hot waters rise convectively, they react with their host rocks, dissolving some constituents such as silica and altering primary minerals to develop a new mineral suite, the constituents of which reflect the reaction temperature and chemistry.

Chemical processes resulting from hydrologic activity may contribute to permeability and fluid flux enhancement. Two main types of chemical processes are important. The first involves selective dissolution of minerals during alteration and the consequent production of increased porosity and permeability. The second occurs where alteration changes the mechanical properties of the rock mass, especially where it leads to a greater propensity for fracture (*e.g.* by silicification) (Henley & Etheridge, 1995).

Henley & Ellis (1983) and Bogie & Lawless (1986; 1987; in Clarke, 1991) have summarised the hydrologic environment which gives rise to geothermal activity beneath stratovolcanoes (Figure 6). The proximal vent zone is at a higher elevation than its surrounds, thus geothermal circulation is recharged from relatively low elevations; and the piezometric surface to the deep geothermal reservoir will lie at significant depth (Leach *et al.*, 1985; in Clarke, 1991). Above this surface, vapour evolved from the hydrothermal system may condense with downward percolating meteoric water. The zone beneath the piezometric surface will be a locus for fluid mixing. These geothermal systems may have long lateral outflows which tend to dissipate the energy of the system and, combined with a deep piezometric surface reduce the likelihood of hydrothermal eruption (Clarke, 1991).

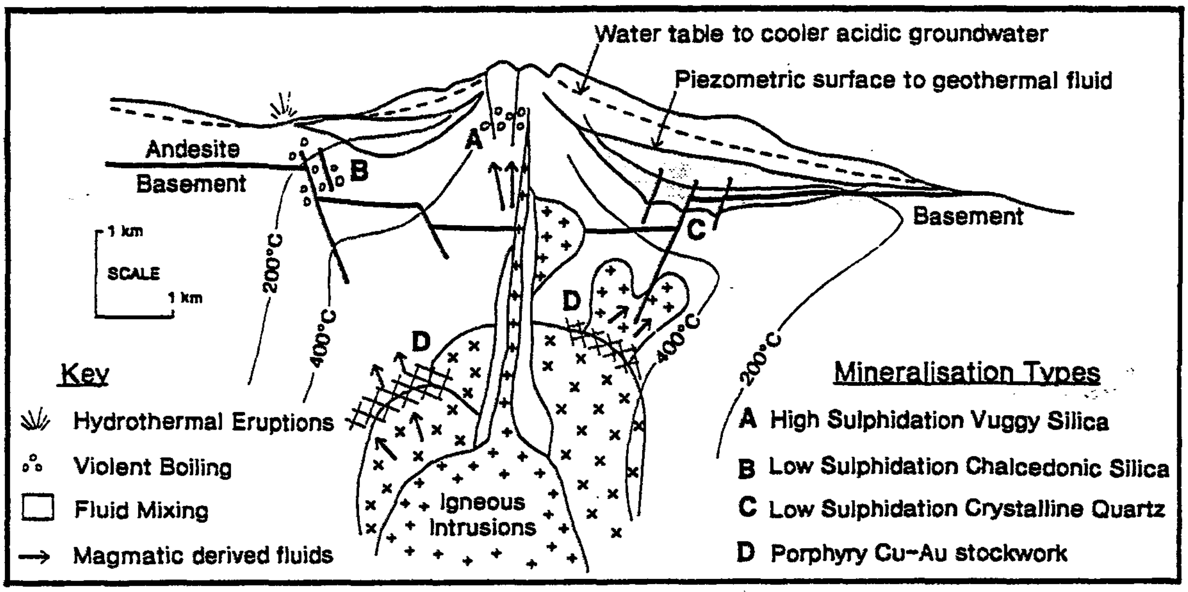


Figure 6. Schematic section through an andesitic stratovolvano, showing a stylized hydrologic regime and resultant hydrothermal mineralisation (derived from Henley & Ellis, 1983; Bogie & Lawless, 1987; Hedenquist, 1987; and Sillitoe, 1988, 1989; after Clarke, 1991).

In contrast, rhyolitic terrains are characterised by geothermal systems in geomorphologic lows, with a piezometric surface for the deep geothermal reservoir near or at the ground surface, and groundwater recharge derived from similar elevations (Figure 4). In these systems the same basic processes occur as those beneath stratovolcanoes (Henley *et al.*, 1984), but there is only limited development of condensed acid waters, where the piezometric surface lies below ground level, and hot spring environments with sinters occur where the deep geothermal fluids vent to surface. Notice that chloride water springs occur several kilometres from the hot upflow part of the system (Henley *et al.*, 1984). Hydrothermal eruptions are common (Hedenquist & Henley, 1985; in Clarke, 1991).

In general, temperatures in the upper 0,5 - 2 km of the deep systems follow a boiling point to depth relationship. The boiling that occurs in this region results in the transfer of gases (CO_2, H_2S, CH_4, *etc*) into a vapour phase. This phase then migrates independently to the surface to form fumaroles. It may encounter near surface cold groundwater into which it will condense to form steam-heated waters; oxidation of H_2S in this environment produces acid-sulphate waters (with low chloride contents and pH in the range of 0 - 3.0) which react rapidly with host rocks to give advanced argillic alteration assemblages (kaolinite, alunite, *etc*). Alternatively, deep chloride waters may flow directly to the surface to form boiling, near neutral to alkaline pH, high-chloride springs, or may become diluted to give relatively dilute chloride waters. Depending on the hydrology of the surface system each of these water types may interact with each other to give hybrid water types (Henley *et al.*, 1984).

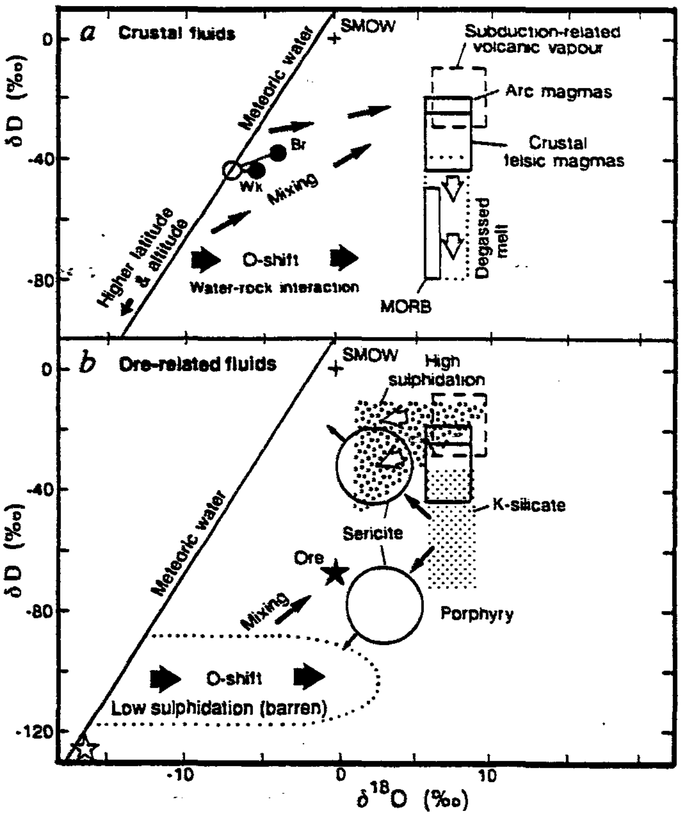
1.3 ORIGIN AND NATURE OF MINERALISING HYDROTHERMAL FLUIDS

Isotope systematics provide information on the sources and nature of hydrothermal fluids, and also on the mechanisms responsible for ore deposition (Golding & Wilson, 1984). Fluid inclusions, trapped in many minerals formed in the epithermal environment, yield additionally useful thermometric data. Quartz usually provides the most fertile opportunities for collection of interpretable fluid-inclusion data (Bodnar *et al.*, 1985).

These studies, particularly those of isotopic ratios of oxygen and hydrogen in geothermal systems and epithermal ore deposits (Figure 7), have shown that the hydrothermal fluids in such environments are predominantly meteoric water in origin (Taylor, 1973; 1974; White, 1974; Bethke & Rye, 1979; O'Neil & Silberman, 1974; in Silberman & Berger, 1985). The fluids in most of the systems are dilute, with NaCl eq. of 0.5 to 5 wt. percent. There is, however, also isotopic (Giggenbach, 1992a; in Hedenquist, 1995) and chemical (Giggenbach, 1992b; in Hedenquist, 1995) evidence that magmatic fluid (vapour and in some cases hypersaline liquid) mixes with deeply circulating meteoric water during heating near an igneous intrusion (Figure 8).

Since many ascending magmas are saturated with a fluid phase, some shallow (<5 km depth) magma bodies may have significant amounts of aqueous fluid of magmatic origin stored within (and possibly adjacent to) the magmatic system. This fluid will contribute to

Figure 7. *(a)* Variation in the O- and H-isotope composition of crustal magmas and waters $\delta^{18}O$ and δD, respectively, relative to Standard Mean Ocean Water - (SMOW). Felsic magmas underlain by continental crust have slightly heavier H-isotope ratios than island arc magmas, presumably due to incorporation of crustal material. These magmatic waters are both isotopically distinct from that associated with mantle-derived mid-ocean-ridge basalts (MORB). The isotopic composition of 'subduction-related magmatic vapours' outgassing from arc volcanoes has a narrow range, enriched from their parent arc magmas due to fractionation during degassing, with the degassed melt (open arrows pointing down) becoming lighter in H-isotope composition. There are two trends for neutral-pH geothermal waters, one caused by meteoric water line; (for example, that of Wairakei, Wk), and the other from mixing between meteoric water of variable (local) isotopic composition and low-salinity magmatic vapour (small solid arrows with trends pointing towards volcanic vapour; for example, that of Broadlands, Br). Connate waters trapped in sediments and waters related to metamorphism have wide ranges of isotopic composition (not shown here). *(b)* The isotopic composition of hydrothermal fluids associated with porphyry ore deposits is deduced from the composition of high-temperature K-silicate alteration minerals (stippled field). Lower-temperature sericitic alteration indicates various mixtures (typically plotting in the two large circles) between late-stage magmatic water (derived from a degassed melt) and meteoric water with a range of isotopic compositions depending on the palaeolatitude and altitude; this meteoric water is sometimes O-shifted. 'High-sulphidation' waters are similar in isotopic composition to volcanic vapour, with trends projecting towards meteoric water of variable isotopic composition (open arrows) depending on location of the deposit. The waters that precipitate barren quartz in 'low-sulphidation' deposits commonly show an O-shift from local meteoric water values, whereas high-grade ore samples (Comstock lode; solid star) have both an O- and H-isotopic shift from local meteoric water (open star) caused by a component of magmatic water; such a huge shift cannot be caused by boiling of meteoric water (after Hedenquist & Lowenstern, 1994).



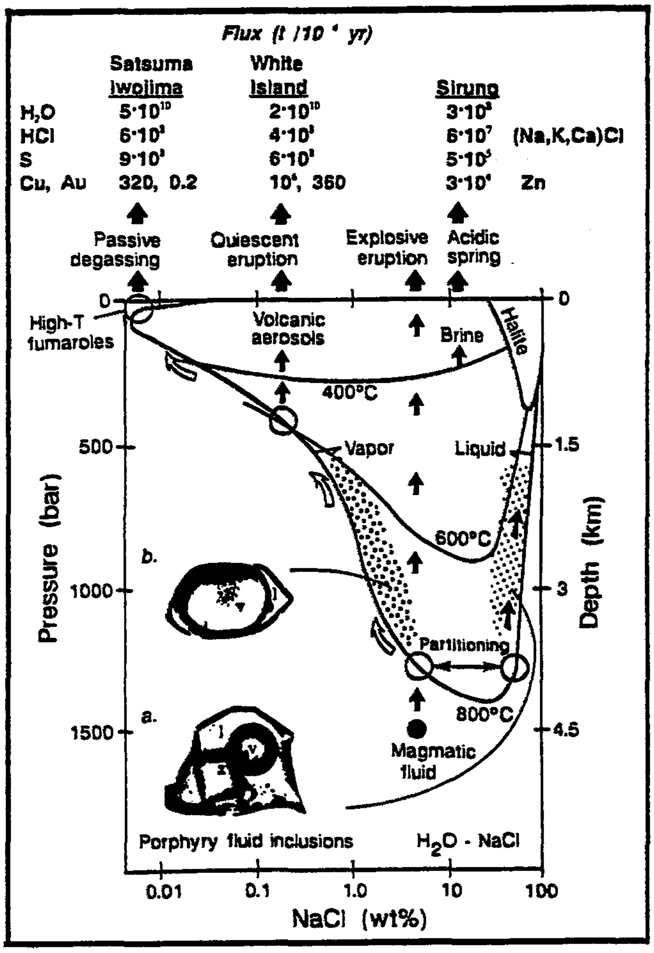


Figure 8. Immiscibility in the system H_2O-NaCl at 800, 600 and 400°C as a function of pressure (Pitzer & Pabalan, 1986); the approximate depth is for a lithostatic pressure gradient. Exsolution of an aqueous fluid from a typical magma may have a bulk NaCl concentration of several wt.%, though at depths <5 km hypersaline liquid and volatile-rich vapour will form (Shinohara, 1994); chloride-complexed metals will tend to partition to the hypersaline phase. There is fluid-inclusion evidence for hypersaline liquid (stippled field) and low-density vapour (open stipple pattern) in most porphyry ore deposits. The photographs show representative inclusions from Bingham porphyry Cu deposit (courtesy of E. Roedder); (a) hypersaline fluid inclusion with liquid (l), vapour (v) and daughter salt crystals including halite and sylvite (x); (b) vapour-rich fluid inclusion. Explosive eruption of the magma should result in the bulk aqueous fluid being discharged to the atmosphere. By contrast, quiescent eruption from the top of a shallow-emplaced magma body may not include a representative sampling of the hypersaline phase; aerosols accompanying quiescent eruption indicate that NaCl concentrations are <1 wt.% in the erupting fluid, though with appreciable metal concentrations (*e.g.*, White Island). Vapour that passively degasses from high-temperature fumaroles (*e.g.*, Satsuma Iwojima) has a very low content of NaCl (and metals), consistent with the low NaCl solubility in vapour at near-atmospheric pressure. Thus, despite the H_2O, quiescent eruption and passive degassing, the metal fluxes associated with low-pressure vapour are very low. The Zn-rich saline brine discharging at Sirung may have formed by separation of medium-temperature (400°C) vapour (after Hedenquist, 1995).

the formation of hydrothermal systems adjacent to the intruding magmatic body (Hedenquist, 1995).

As magmatic fluid systems cool, the brittle-ductile transition eventually enters the area within the crust of deep crystalline rocks with their saline liquid phase fluids (Figure 9). As far as stable isotope geochemistry is concerned, the deep crystalline rocks between the brittle-ductile transition and the water rich carapace of the magma can be viewed as a major reservoir of evolved magmatic fluids for epithermal ore deposits. A lithostatic to hydrostatic pressure transition is probably the trigger that episodically is tripped by stresses in the system to literally fire these dense, magmatic, saline fluids to higher-level permeability zones. Such fluids as those at Providencia, Mexico, for example, ascend so rapidly that they retain their deep-seated characteristics and are out of isotopic and chemical equilibrium with wall rocks at the site of ore deposition. Substantial mixing occurs only in the shallow parts of the system (Rye, 1993).

Stable isotope studies of epithermal systems, when supported by detailed geologic framework studies and fluid inclusion and other geochemical data, not only identify magmatic fluid components but also place important constraints on their evolution in the system. In such systems the magmatic contribution to the epithermal environment will be largely limited to heat and gases from the magma (Rye, 1993).

A predominance of meteoric water has been documented in several epithermal deposits (Figure 10). It should be noted, however, that a magmatic component of up to ten percent could be hidden in uncertainties and cannot be ruled out (Hayba *et al.*, 1985). Recent studies, coupled with those of active hydrothermal systems, show that magmatic fluids are commonly present, but that their signatures may be masked or erased owing to later overprinting by large volumes of meteoric water (Hedenquist & Lowenstern, 1994).

The evidence for meteoric-water dominance in epithermal systems is most obvious in regions where magmatism was most voluminous, because the large magmatic heat source creates large and long-lived cells of convecting meteoric water; convection that will erase most of the evidence for any early component of magmatic water. These meteoric waters continue their deep circulation long after an intrusion has solidified (Hedenquist & Lowenstern, 1994).

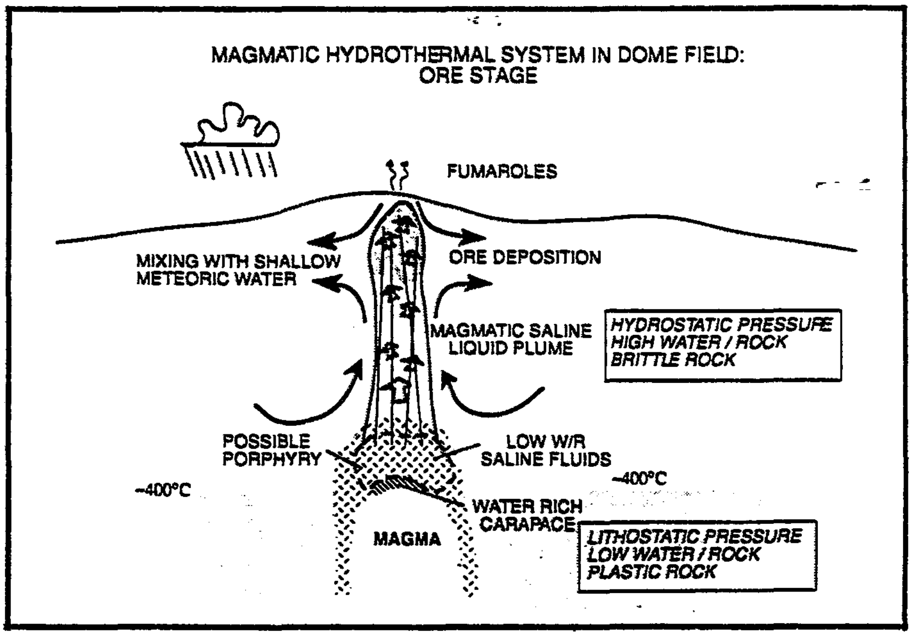


Figure 9. Model of high-level magmatic hydrothermal system in dome field showing relationship of ore-stage saline magmatic fluids to the water-saturated carapace of the magma and the brittle-ductile transition and mixing with shallow meteoric water (after Rye, 1993).

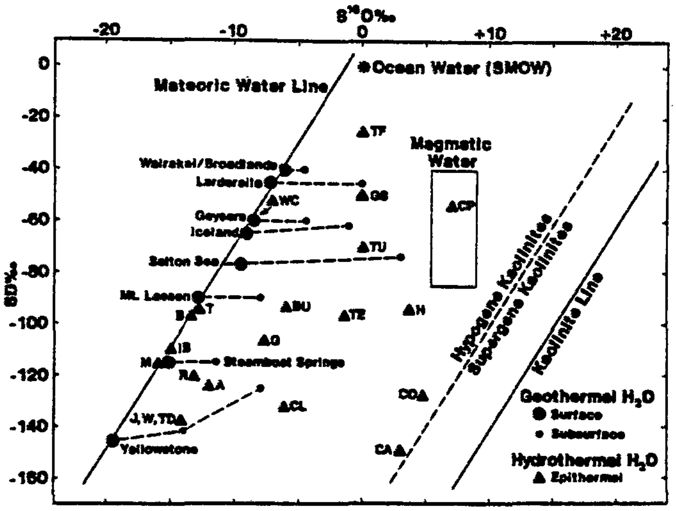


Figure 10. Distributions of δD and $\delta^{18}O$ in various waters, minerals, and hydrothermal fluids of epithermal deposits (after Field & Fifarek, 1985).

In conjunction with fluid and laboratory studies it may be shown that the formation of gold-silver (base-metal poor) epithermal deposits requires fluids of low salinity and high gas content (*i.e.* H_2S in association with CO_2). Such studies also highlight the important role of gas composition in controlling the physical processes within systems and their metal transporting capability, as well as demonstrating that two processes (adiabatic boiling and mixing) provide the principal controls on fluid chemistry in the epithermal environment (Section 1.6). Access to some gas source (CO_2 + H_2S) becomes paramount in determining whether a given low-salinity system may or may not develop a gold deposit (Henley, 1985).

The flux of metals measured from some erupting volcanoes indicate that, given sufficient time and a concentration mechanism, degassing magmas can exsolve sufficient metals to create an ore deposit. The best current estimate of the initial isotopic composition of H_2O in arc and crustal felsic magmas (Figure 7a; solid-line boxes) is intermediate between that of the degassed magma (Figure 7a; dotted-line box) and the isotopically heavy outgassed H_2O, the latter sampled as volcanic vapour (Figure 7a; dashed-line box) (Hedenquist & Lowenstern, 1994).

Evidence clearly indicates that magmas contribute components such as water, metals and ligands to hydrothermal fluids. Not surprisingly, this evidence wanes as the distance from the intrusion increases, meteoric water becomes dominant, and the fluid salinity and acidity decrease (Hedenquist & Lowenstern, 1994).

When considering the predominantly meteoric nature of mineralising epithermal fluids, the origin of the precious metal content in the fluids would seem to be by extraction from the source rocks at depth as the circulating fluids are heated. Meteoric fluids could theoretically descend to the base of the relatively permeable brittle layer of the crust, where they encounter the brittle-ductile transition zone corresponding to a temperature of around 400°C (Kuznir & Park, 1983; in Mitchell & Leach, 1991). Nesbitt (1988; in Mitchell & Leach, 1991) has argued that in the volcanic environments where epithermal gold systems develop, the geothermal gradient approached 65°C/km, implying a depth to the 400°C isotherm of 5-6 km. The source of gold in a predominantly meteoric epithermal field therefore lies within 6 km of the surface (Figure 11).

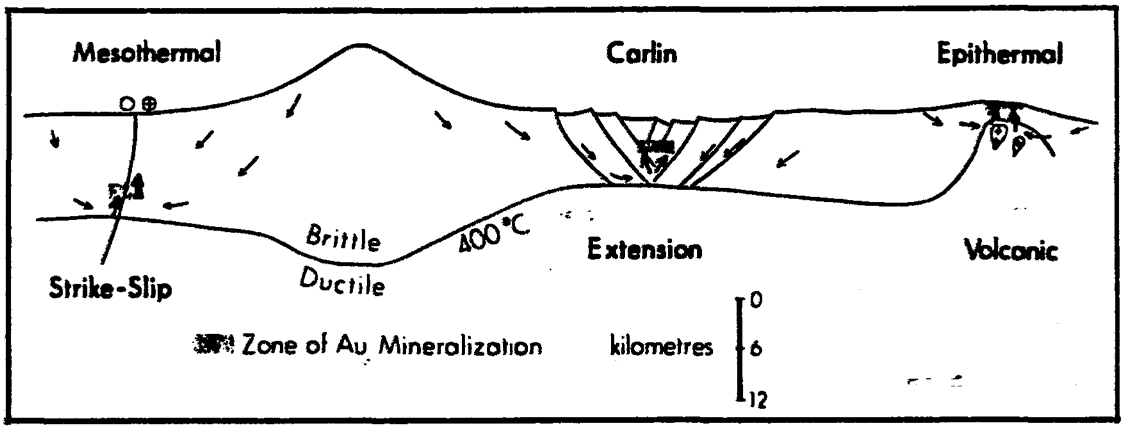


Figure 11. Schematic diagram illustrating convection of meteoric water in various tectonic settings and resulting gold mineralisation. Thickness of arrows is roughly proportional to value for water/rock ratios (from Nesbitt, 1988; in Mitchell & Leach, 1991).

Alternatively, if a passively degassing magma chamber exsolves a high-pressure vapour with a metal content similar to that deduced for vapour associated with quiescent eruption, and this metal-rich vapour is absorbed at depth by meteoric water, a potential ore-forming liquid may be generated. Evidence for the formation of a metal-rich hypersaline liquid is commonly seen in fluid inclusions associated with igneous and volcanic rocks. Assuming that the metal fluxes of quiescently erupting volcanoes (Table 1) can be extrapolated over a period of 10 000 years, then it is possible that, to account for the size of typical epithermal ore deposits, sufficient amounts of metals can be degassed from a magma. The fluid thus generated will have a low salinity (<1wt.% NaCl), and will be relatively acidic and oxidised (due to the HCl and SO_2 content) (Hedenquist, 1995).

Table 1: Component flux from quiescently erupting volcanoes (after Hedenquist, 1995)

			COSPEC SO_2	H_2O	S	CO_2	Cl	F
Volcano	**Refs**	**Date**	**t/d**		**(x10^6 t/yr)**			**(t/yr)**
White Island, NZ #	10, 11, 12	1983, 85	350-1200	1.9	0.078	0.49	0.03	420
Arenal, Costa Rica	13	1982	210	0.7	0.038	0.07	nd	nd
Colima, Costa Rica	13	1982	320	nd	0.058	nd	nd	nd
Redoubt	3	1989-91	3400	nd	0.25	>0.5	>0.2	nd
Etna**	14	1987	190-3200 (av. 1120)	nd	0.2	nd	0.047	58400
Stromboli, Italy	15	1990-93	320-1200 (av. 800)	2.3	0.15	2.1	0.045	1390

\# plus Cu 110 t/yr, Au>36kg/yr; Na/S=$55x10^3$, Cu/S=$2400x10^{-6}$

** plus (t/yr) Cu 580, Zn 4700, Mo 84, and As 31, and Au 84 kg/yr, Na/S=$64x10^{-3}$, Cu/S=$2860x10^{-6}$

1.4 MINERALOGY AND GEOCHEMISTRY OF MINERALISING HYDROTHERMAL FLUIDS

There are at least seven parameters affecting the chemistry of hydrothermal fluids that need to be considered whilst investigating fluid mineralogy and geochemistry. These 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).

Figure 12 presents the distribution of fluids in a hydrothermal system thought to be responsible for the formation of a volcano-hosted, epithermal, now LS-type mineral deposit. The system is assumed to be situated on the flanks of a volcanic structure and therefore to be supplied by "magmatic" vapours. Oxidation of H_2S carried by the vapours gives rise to the production of dilute, acid SO_4 waters, percolating downward to considerable depth, up to 500m, and the establishment of extensive zones of argillic alteration (Reyes, 1991; in Giggenbach, 1992). The multiple source of components in LS epithermal fluids is caused largely by the extensive interaction between thermal waters and crustal rocks that occurs in this environment (Hedenquist & Lowenstern, 1994). Isotopic and chemical evidence suggest that the major fluid components in hydrothermal systems associated with volcanic activity along convergent plate boundaries (H_2O, CO_2, and probably Cl^-), may to a certain extent be derived from subducted marine sediments and seawater (Figure 13).

Since many hydrothermal fluids are close to calcite saturation, indicating that calcite plays a major role as a mineral-gas buffer, the presence or absence of calcite in a hydrothermal mineral assemblage directly reflects the concentration of aqueous carbon dioxide of the coexisting fluid (Ellis, 1969; 1970; Ellis & Mahon, 1977; Arnórsson, 1978; Giggenbach, 1981; 1988; in Simmons & Christenson, 1994).

Therefore the gangue mineralogies of epithermal deposits record such chemical processes in the depositional regime. Loss of CO_2 due to boiling leads to supersaturation with respect to calcite according to,

$$\boxed{CaCO_3 + CO_{2(g)} + H_2O \quad = \quad Ca^{++} + 2HCO_3^-}$$

(Henley, 1991).

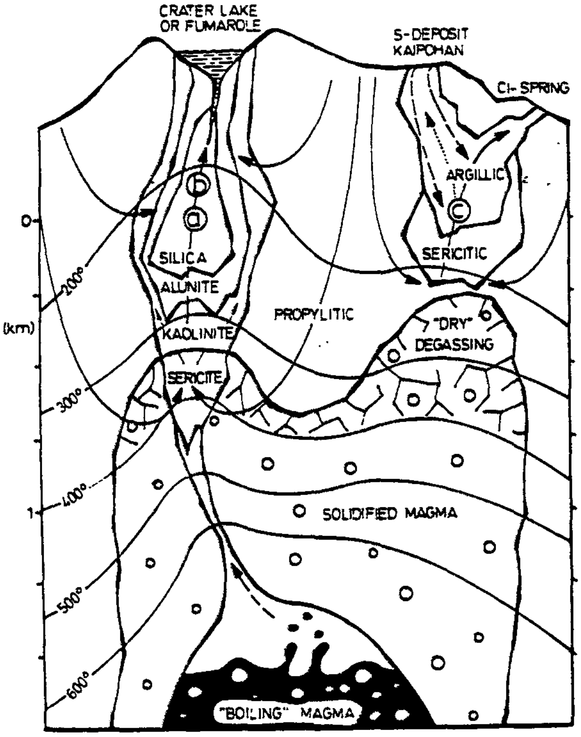


Figure 12. Schematic models of hydrothermal systems responsible for the formation of "high sulphidation" (paths a and b) and "low sulphidation" (path c) epithermal mineral deposits. The low sulphidation system, on the right, is assumed to be situated on the flanks of a volcanic structure. The dotted line represents the path of gas-charged vapours, the dashed lines that of downward-percolating air-oxidised, acid sulphate waters (after Giggenbach, 1992).

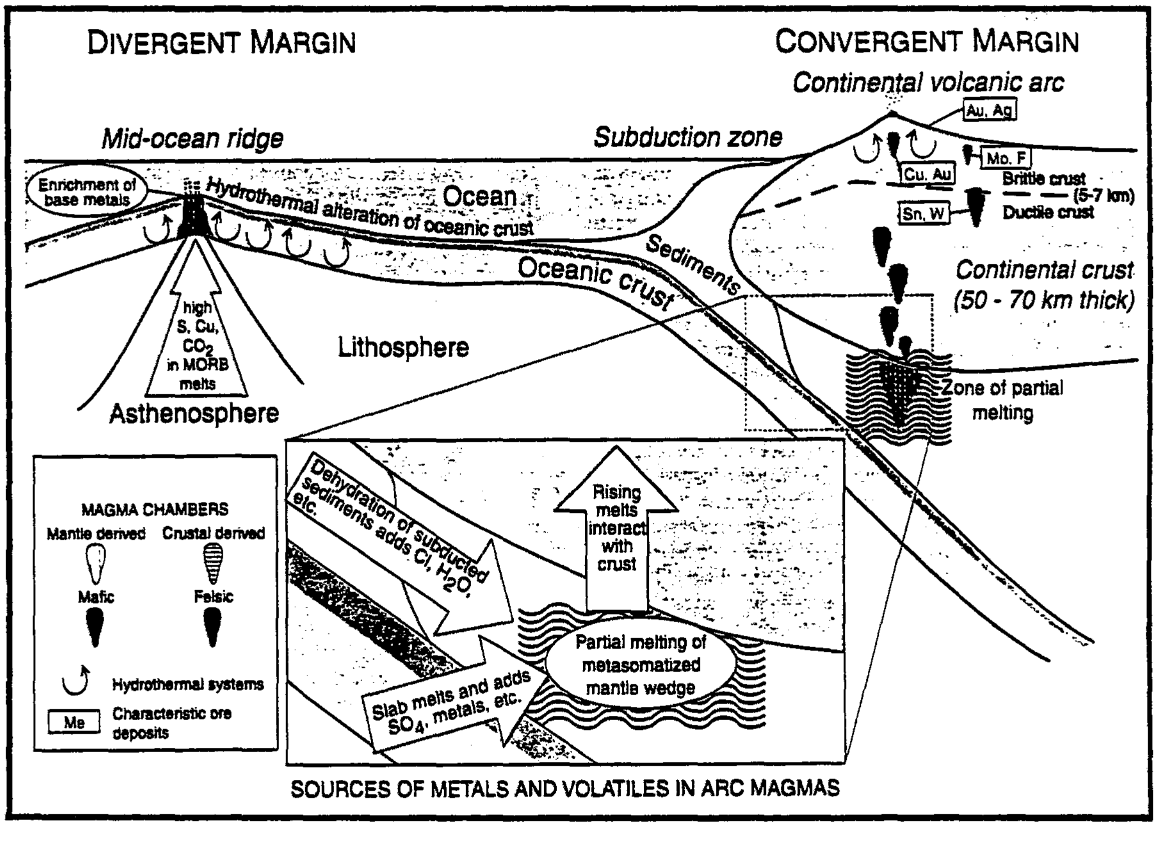


Figure 13. Schematic section showing the principal components of magma genesis, fluid flow and metallogenesis in divergent- and convergent-margin settings (after Hedenquist & Lowenstern, 1994).

Also, the change in pH attendant on boiling commonly leads to the precipitation of adularia in vugs (*e.g.* Kelian and Ladolam) or interbanded with silica (*e.g.* Hishikari, Golden Cross, and McLaughlin). In exploration, it must be remembered that the occurrence of adularia and calcite are indicative of phase separation and are not themselves indicative of the gold potential of a deposit or prospect (Henley, 1991). Voluminous deposits of siliceous sinter generally indicate deposition from neutral to slightly alkaline (by loss of CO_2), chloride-rich waters that flowed quickly to the surface from a reservoir with a temperature in the range 200°C to 270°C (Fournier, 1985).

In studies of mineral deposits, alteration assemblages are frequently used in conjunction with salinity estimates from fluid inclusion data to indicate the pH of ore-forming fluids (Henley *et al.*, 1984). Fluid trapped in inclusions during hydrothermal mineral growth is the only sample available of the fluid actually present during formation of these ore deposits (Roedder, 1967; 1976; in Hedenquist & Henley, 1985). Fluid inclusion, stable isotope, and mineral alteration studies of high-level hydrothermal ore deposits in volcanic settings have led to the recognition of physical and chemical environments analogous to those encountered in many present-day geothermal systems, such as those in the circum-Pacific region (Ellis & Mahon, 1977; Henley & Ellis, 1983; in Henley *et al.*, 1984).

At magmatic temperatures many species are associated as neutral complexes rather than ionic compounds. So, for example HCl, NaCl, and KCl are not ionic, but neutral molecules at temperatures above about 500°C or so (Franck, 1956; Quist & Marshall, 1968; in Henley *et al.*, 1984). Similarly, the sulphurous gases are present as H_2S and SO_2 rather than ionised species. As a consequence, reactions at magmatic temperatures are largely between neutral species.

Redox reactions are important in the oxidation of H_2S, the precipitation of native metals, pyrite and other sulphides, the destruction of sulphides by oxidation, and the disproportionation of SO_2 into H_2S and $SO^{=}_4$ on cooling from high temperature (Henley *et al.*, 1984). Given the appropriate fluid composition, most ore metals can be partitioned strongly into a magmatic-hydrothermal fluid as chloride, bisulphide and hydroxyacid complexes. At magmatic temperatures and pressures, insufficient data exist to define the ideal conditions for partitioning of any given metal from a magma into an exsolved hydrothermal fluid (Hedenquist & Lowenstern, 1994). However, exploration drilling has identified two end-member redox environments in active geothermal systems: a) the relatively reducing environment of the deep, chloride-water systems which are associated with pyrite-rich propylitic alteration and with H_2S as the dominant aqueous sulphur species, and b) the relatively oxidising environment in the upper part of systems where H_2S oxidation results in acidic, sulphate-dominated waters associated with advanced argillic alteration. A similar, relatively oxidising environment occurs in the upper part of a volcano where volcanic H_2S and SO_2 are oxidised to H_2SO_4 (Henley *et al.*, 1984).

The concentration of a metal, such as iron, is controlled by the relative strength of its complexes with the available anions. Sulphur, for example, is not free to react to form sulphides, unless the reaction involving H_2S or SO_2 is favourable. As temperatures decrease, however, the situation changes drastically as species begin to dissociate. H^+, for example, is formed from the dissociation of HCl, and the pH of a chloride solution tends to decrease as temperature falls. Other species such as SO_2 or CO_2 become unstable and will disproportionate. Any of the species may be out of equilibrium with their host rocks and may react, changing both the rock and the fluid (Henley *et al.*, 1984). The addition of SO_2 and other gases to magmatic systems decreases the solubility of H_2O and CO_2 in the melt by decreasing their activity in the fluid phase(s), resulting in fluid saturation with even lower concentrations of gases dissolved in the melt (Hedenquist & Lowenstern, 1994).

1.5 GOLD METAL TRANSPORT IN HYDROTHERMAL FLUIDS

Why and where gold precipitates in the Earth's crust to form an ore deposit is determined to a considerable extent by the nature of the gold complexes in the ore-transporting fluid (Seward, 1988). Gold occurs predominantly in the +1 oxidation state in geothermal fluids in the Earth's crust, and hydrothermal ore solutions contain a number of components which can form stable complexes with gold (Seward, 1982).

From fluid inclusion studies and the analysis of fluids discharged from active geothermal systems, we know that ore fluids responsible for gold deposition are multicomponent electrolyte solutions having sodium chloride as the predominant salt. At temperatures above 300°C, the $AuHS^\circ$ complex will become increasingly important, but there are at present no reliable high-temperature thermodynamic data available. The $Au(HS)_2$ species will remain extremely stable from 300° to 400°C, but its field of maximum stability shifts to more alkaline pH and hence to conditions less relevant to natural hydrothermal fluids (Seward, 1988). Seward (1973; 1982; in Henley *et al.*, 1984) has shown that thio-complexes of gold are stable to at least 300°C and dominate transport of the metal in geothermal fluids (Figure 14). Seward (1984; in Wake & Taylor, 1988) also demonstrates that gold in solutions of high chloride content, low sulphur content, or that are strongly oxidising, will be carried as chloride complexes.

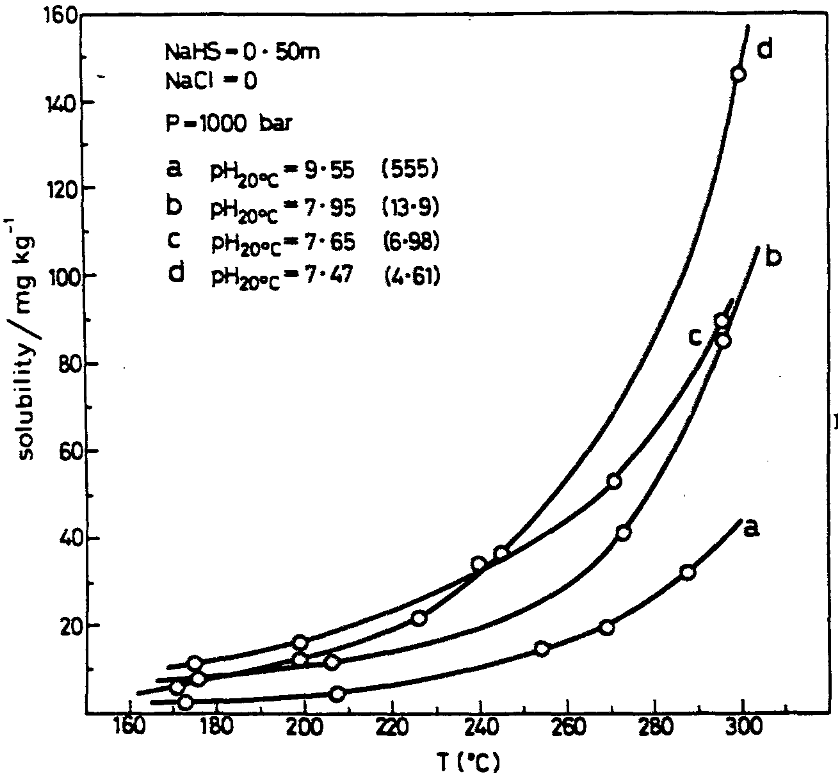


Figure 14. Solubility of gold in sulphide solutions up to 300°C at 1 000 b pressure; the numbers in brackets are the ratio, HS^-/H_2S; from Seward (1973; in Seward, 1982).

The frequent association of gold with CO_2-rich fluids and carbonate gangue led Boyle (1979) and Boyle *et al.*, (1985; in Wake & Taylor, 1988) to suggest gold solubility in carbonate solutions. Fyfe & Kerrich (1984; in Wake & Taylor, 1988) have documented experimental data indicating solubility as carbonate complexes. Saunders & Romberger (pers.comm., 1985; in Wake & Taylor, 1988) describe gold-telluride deposits from the Colorado mineral belt, also deposited by CO_2-rich fluids, where gangue mineralogy suggests an oxidising environment of transportation and deposition of gold transported as the ditelluride (Te_2^{2-}) complex.

While gold may be carried in solution as a sulphide or chloride, in dilute brines at epithermal temperatures, gold occurs largely as a bisulphide complex, $Au(HS)_2$ (Seward, 1973; Henley, 1986; Cole & Drummond, 1986; in Mitchell & Leach, 1991). Gold deposition requires destruction of this complex, the most effective mechanisms for which are loss of H_2S gas as the fluid rises and decompresses, cooling of the fluid, or mixing (Reed & Spycher, 1985).

Confirmation that the principal gold-transporting species is $Au(HS)_2$ comes from the observation by Brown (1986; in Henley, 1991) of bonanza-grade (6wt% gold) precipitates resulting from boiling of Broadlands geothermal fluid near the control plates of discharging geothermal wells. These data confirm that the Broadlands fluid is, at depths of a few hundred to a thousand metres, close to saturation with gold as a bisulphide complex. Equivalent gold arsenide species may contribute in a small way to the solubility but, as shown in Figure 15, the systematics of gold transport and deposition may be discussed satisfactorily in terms of the dominant bisulphide complex (Henley, 1991).

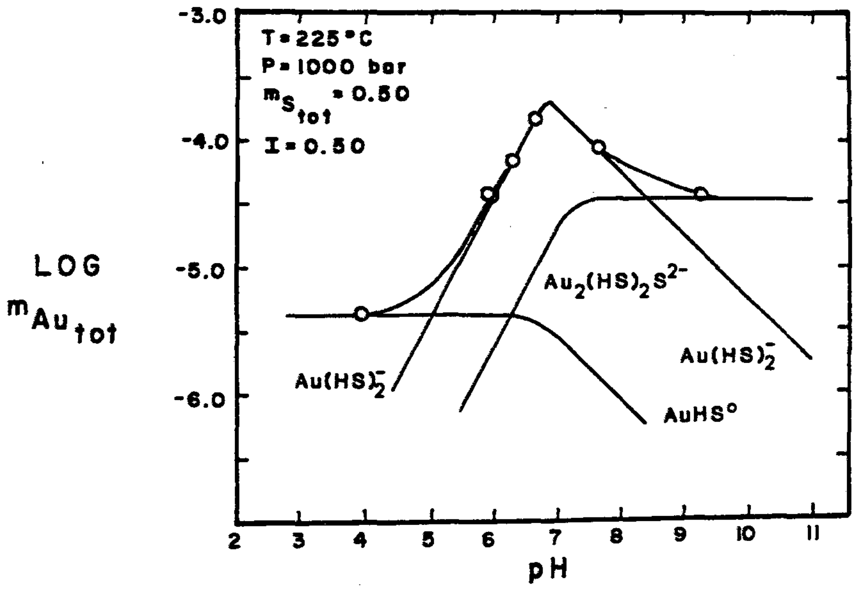


Figure 15. Calculated solubility curves for three thio-gold complexes compared with experimental data at 225 °C and 1 000 bars pressure (from Seward, 1973; in Henley *et al.*, 1984).

Based on solubility reactions of the form;

$$\begin{aligned} Au + 2H_2S &= Au(HS)_2 + H^+ + 0.5H_2 \\ ZnS + 2H^+ + 2Cl &= ZnCl_2 + H_2S \end{aligned}$$

for Au, Ag, Cu, Zn, Pb, Henley (1985; 1990; in Henley, 1991) has shown that the geologic solubility of gold and base metals (Ag, Pb, Zn) may be calculated for a wide range of salinity's and H_2S contents.

These findings indicate that the solubility of gold in epithermal environments is inversely proportional to salinity and directly proportional to m_{H_2S} (Figure 16). The solubility of chloride-complexed metals is the inverse of this behaviour and this provides the essential chemical distinction between epithermal precious metal and epithermal base metal-silver deposits. Gold transport is maximised in solutions containing relatively high concentrations of H_2S which is inversely related, through fluid-mineral equilibria, to the concentration of CO_2. The implication of these data is that the solubility of gold in crustal fluids in volcanic terrains is strongly linked to CO_2 concentration and to temperature in the alteration-buffered roots of hydrothermal systems. In addition to its role as a source of metals (Au, Cu, *etc*), this highlights the importance of magma degassing and magma evolution in developing epithermal ore-forming systems (Henley, 1991)

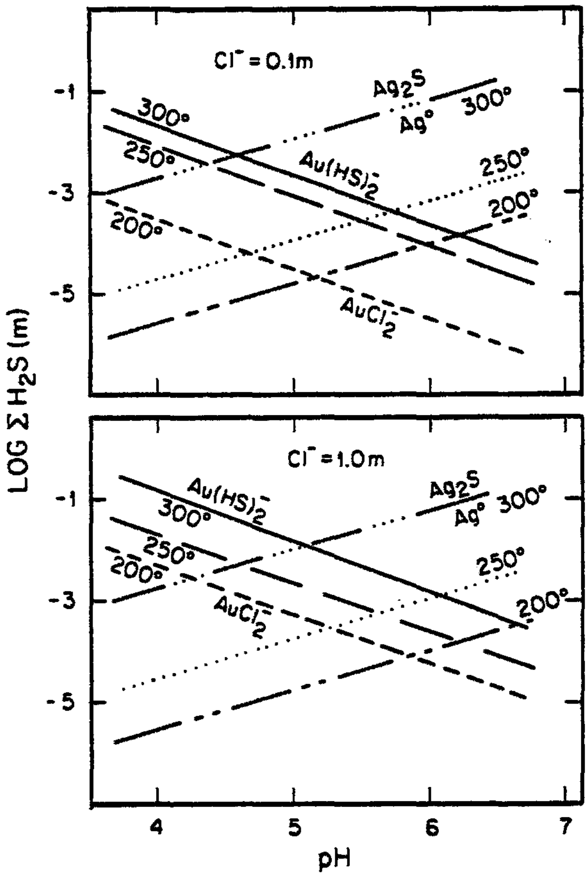


Figure 16. The stability boundaries (equal concentration) between the two dominant gold complexes, $Au(HS)_2^-$ and $AuCl_2^-$ represented in ΣH_2S versus pH space for two chloride concentrations (0.1 and 1.0 m). The stability boundaries for the solid silver phases, argentite (Ag_2S) and native silver (Ag°), are also shown. The two plots were constructed with a $\Sigma H_2S/\Sigma SO_4$ ratio of 10 (after Cole & Drummond, 1986).

Knowledge of variables such as pH and redox state, along with temperature, fluid salinity and gas content, are fundamental to understanding the controls on metal transport and deposition. For example, in low-salinity reduced fluids (characteristic of LS systems), gold is likely to be transported in solution as a bisulphide complex. Boiling is a common if not ubiquitous process operating in conduits at epithermal depths; it causes acid gases (dominantly CO_2 and H_2S) to be lost from the liquid, resulting in an increase in pH. This initially increases the solubility of gold, but eventually the loss of H_2S from the liquid causes the solubility of gold to decrease, thus leading to its precipitation (Henley *et al.,* 1984; in White & Hedenquist, 1995). By contrast, in oxidised and acidic solutions of moderate salinity (characteristic of HS systems), gold may be transported as a chloride complex, with different controls on its precipitation, such as dilution and/or cooling (Giggenbach, 1992; Hedenquist *et al.,* 1994).

Another possibility involves the meteoric water absorption of high-pressure vapour with a metal content similar to that indicated by the bulk composition of fluid erupting from volcanoes (Table 1). This suggestion is consistent with the evidence for low-salinity liquid (2-4 wt.% NaCl equivalent) having formed enargite in the Lepanto HS gold deposit (concluded from infrared fluid-inclusion study of enargite; Mancano & Campbell, 1995; in Hedenquist, 1995). There is, however, also evidence in some deposits for liquids of higher salinity (>15 wt.% NaCl) having been present at the depth of the ore zone (Arribas, 1995; in Hedenquist, 1995). If this saline liquid is responsible for mineralisation in some cases, it may be related to the hypersaline counterpart of the gas-rich vapour.

The variable involvement of these principal fluids (saline liquid, vapour, deep meteoric water, and groundwater; Arribas, 1995; in Hedenquist, 1995) will be largely governed by the hydrology of each individual system; such hydrological variation may thus account for the range of characteristics noted by Arribas for these deposits. Further infrared fluid-inclusion work and detailed stable-isotope studies on ore-mineral inclusion fluids (Deen 1990; in Hedenquist, 1995), coupled with geologic and hydrologic reconstructions, are necessary to help in resolving this as yet unanswered question.

Other key unresolved questions relating to metal transport ask what exactly controls fluid compositions, which fluids are most likely to scavenge metals from silicate melts, and how much fluid can be channelled to a site of metal deposition (Hedenquist & Lowenstern, 1994).

1.6 PRESSURE AND TEMPERATURE OF MINERALISING HYDROTHERMAL FLUIDS

A hydrothermal system undergoes abrupt physical and chemical change at the shallow depth that characterises most epithermal deposits. This occurs because of the change from lithostatic to hydrodynamic pressure (resulting in boiling, Figure 17), interaction of fluids derived at depth with near-surface water (Figure 18), permeability changes, and reaction between fluid and host rocks. These changes near the surface are the reason that an "epithermal" ore environment exists, as they affect the capacity of the hydrothermal fluid to transport metals in solution. Focusing of solubility of metals in the fluid, will then result in metal deposition within a restricted space (White & Hedenquist, 1990).

Gold is transported by hydrothermal fluids in the Earth's crust over a wide range of temperatures. The temperatures of formation of most gold deposits are in the range 175 - 450°C (Seward, 1982). Fluid-inclusion data for HS-type deposits indicate that ore deposition occurs primarily at temperatures similar to those of LS-type deposits (200°C to 300°C) (Hayba *et al.*, 1985).

Pressures associated with many hydrothermal environments to depths of 4 - 5 km are approximately those for a column of water at boiling point throughout its length (Seward, 1982). Palaeodepth estimates from both geologic reconstructions and pressures estimated from fluid-inclusion studies show that most of the LS-type deposits appear to have formed at palaeodepths of 300-600 m. Palaeodepth estimates for HS-type deposits also appear to be similar to LS-type deposits (Hayba *et al.*, 1985).

1.7 THE ORE FORMING PROCESS AND MECHANISMS OF GOLD DEPOSITION

Through a synthesis of detailed investigations of actual deposits, experimental studies, fluid-inclusion research, and studies of active geothermal systems, a generalised model for the formation of epithermal ore deposits has evolved.

1.7.1 ORE-FORMING PROCESSES

These include:

a) BOILING

In numerous documented deposits, the boiling horizon is the region commonly associated with the zone of most intense mineralisation. This type of spatial association between

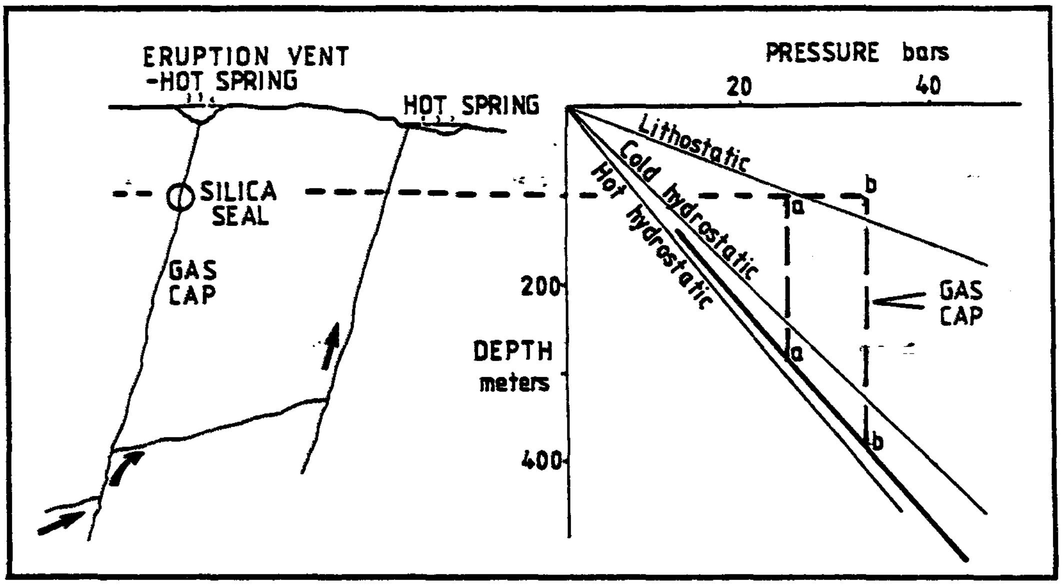


Figure 17. Pressure vs Depth relation for the evolution of hydrothermal eruptions in hot-spring systems and geothermal wells. The formation of a local seal in the system (*e.g.*, through silica deposition, closure of well head valve) allows gas, exsolved during flow through the temperature gradient of the underlying fissure system, to accumulate as a "gas-cap". Prior to sealing, pressures in the flowing system are just above hot hydro-static but as the gas accumulates, progressively greater aquifer pressures are transmitted to the seal by the gas-cap. Eventually the transmitted pressure exceeds lithostatic and an eruption may occur when triggered by seismic shock or opening of the well-head valve. Much of the succeeding eruption and crater formation results from erosion of the fissure walls by the expanding gas + steam + water mixture (after Henley *et al.*, 1984).

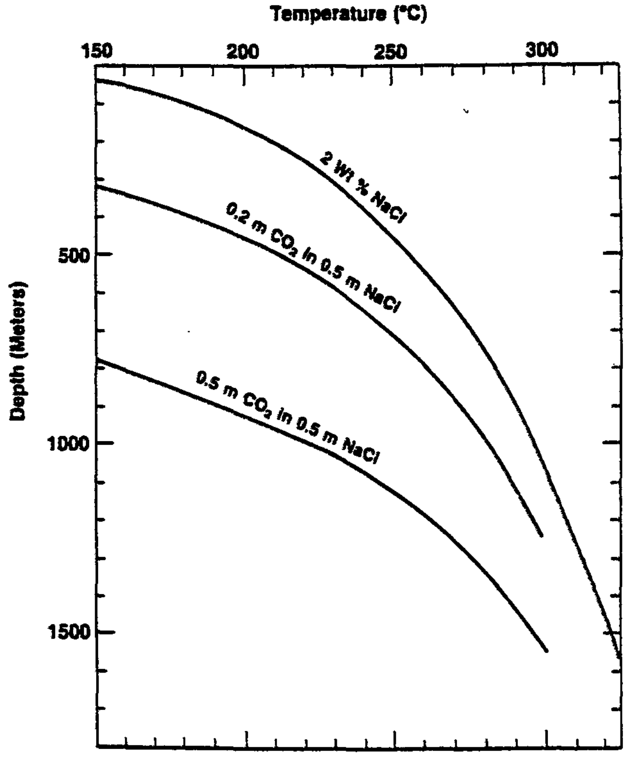


Figure 18. Relationship between temperature and the depth at which boiling will commence for an H_2O-NaCl solution (2wt.%-NaCl) and H_2O-NaCl-CO_2 solutions containing 0.2 and 0.5 molal CO_2 and 0.5 molal NaCl (from Loucks, 1984; in Bodnar *et al.*, 1985).

liquid-vapour conditions and ore deposition prompted many early investigators to suggest that there was a genetic relationship between boiling and mineralisation (Drummond & Ohmoto, 1985). Theoretical and experimental studies (Drummond & Ohmoto, 1985; in Bodnar *et al.*, 1985) have subsequently shown that gold complexed as a thio-complex may be precipitated by any process which causes a decrease in the activity of reduced sulphur (Seward, 1982). It has now generally been accepted (Figure 19), that the most effective mechanism for loss of H_2S gas and gold deposition is that of boiling (White & Hedenquist, 1990; in Mitchell & Leach, 1991). Seward (1982), indicates that if gold is present as $AuCl_2^-$, then deposition may occur in response to decreasing temperature as well as by a decrease in chloride activity caused by dilution. An increase in pH arising from a loss of CO_2 because of boiling also leads to a decrease in $AuCl_2^-$ concentration (Figure 20).

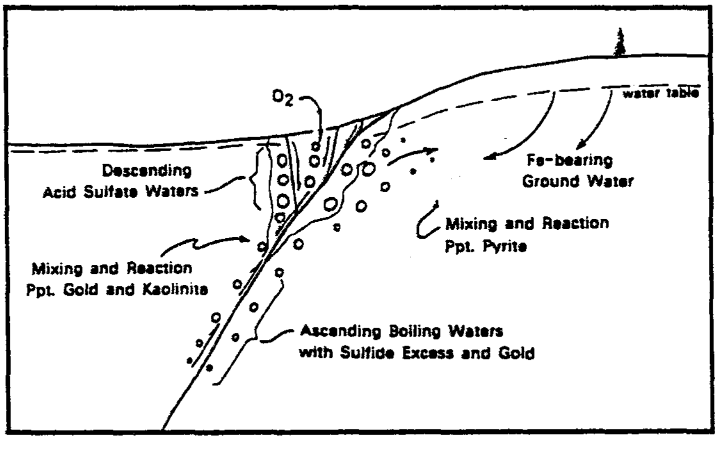


Figure 19. Schematic diagram of fluid-mixing zones in the upper part of a boiling hydrothermal system. In the center of the diagram, ascending, boiled gold-bearing waters encounter descending acid-sulphate waters produced above the water table. The waters mix and react precipitating gold. The descending waters also alter the wall rock to kaolinite, particularly in the hanging wall of the vein. The right-hand side of the diagram shows mixing and reaction of boiled gases (with cold, recharge groundwater), producing a neutral carbonate-sulphate water. If the recharge water is iron rich, reaction of H_2S gas with iron could produce pyrite, which would appear as a "pyritic halo" on the ore deposit (after Reed & Spycher, 1985).

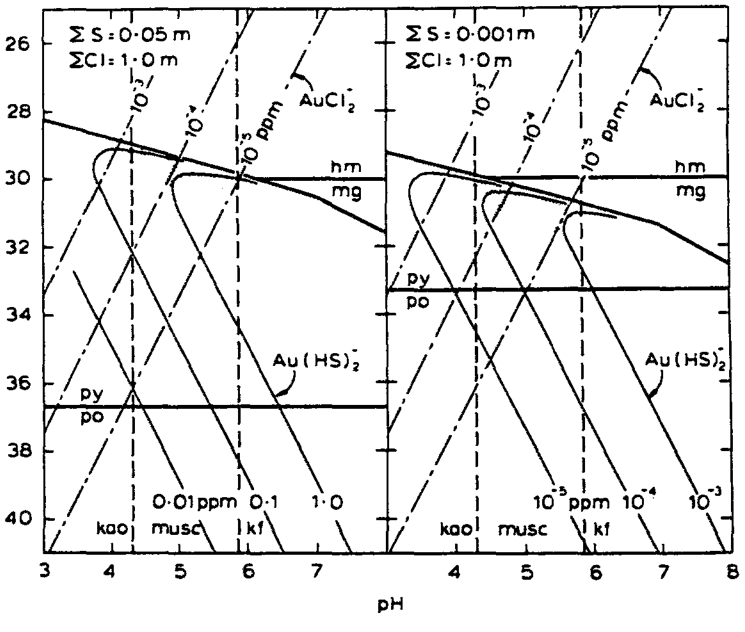


Figure 20. Gold solubility contours for $Au(HS)_2^-$ and $AuCl_2^-$ (in ppm) at 300°C with respect to the stability fields of pyrite (py), pyrrhotite (po), hematite (hm) and magnetite (mg); the vertical dashed line gives the equilibrium pH for the potassium feldspar-muscovite and muscovite-kaolinite hydrolysis reactions; the stippled area indicates the frequently encountered region of ore deposition (after Henley *et al.*, 1984).

b) DECREASE IN TEMPERATURE

Although boiling is an effective mechanism for depositing ores (Henley *et al.*, 1984; Henley, 1985; Drummond & Ohmoto, 1985; Reed & Spycher, 1985; in Hayba *et al.*, 1985), it is not the only mechanism. If gold is transported in hydrothermal ore solutions as a simple hydrosulphide complex like $Au(HS)_2^-$ (Figure 14), a decrease in temperature attending an ascending fluid or arising from interaction with cooler near-surface waters will also cause precipitation of gold (Seward, 1982). Additionally, cooling through mixing with groundwater (without boiling), has recently been invoked to explain mineralisation in permeable tuffs at Round Mountain, Nevada (Sander, 1990; in Mitchell & Leach, 1991).

c) MIXING OF FLUIDS

The mixing of fluids from two or more sources does play a role in the thermal history of some LS-type deposits and has been documented by fluid-inclusion and stable-isotope studies at Creede (Hayba *et al.*, 1985). In HS systems (Bonham, 1986; in Seward, 1988), the mixing of near-neutral, gold-transporting fluids with cooler, acid sulphate waters will similarly cause gold (and probably enargite) deposition in response to decreasing pH and temperature.

d) DILUTION

Deposition of gold through dilution of hypogene fluid with groundwater in the absence of boiling is a possible cause of gold deposition in some deposits of the Pacific Rim too (White & Hedenquist, 1990). Dilution can also be an important process leading to gold deposition in more saline systems. In such systems, dilution of the chloride ion and accompanying temperature drop lead to precipitation of co-transported base metals as sulphides. The consequent removal of H_2S lowers the solubility of gold and leads to the deposition of discontinuous gold-rich shoots in some polymetallic veins (*e.g.* El Bronces, Chile - Skewes & Camus, 1988; in Henley, 1991).

e) WALLROCK REACTION

Wallrock reaction, too, is important as a mechanism of concentrating precious metals in some epithermal systems. Vuggy silica precious metal mineralisation forms primarily by rock reaction, either as magmatically-derived acidic fluids react with country rock, or as subsequently entrained deeply circulating meteoric fluids encounter this altered rock (Berger & Henley, 1989; in Clarke, 1991).

1.7.2 MECHANISM OF GOLD DEPOSITION

The key process in ore deposition mechanisms is fluid focusing. This is most readily achieved by contemporaneous deformation, igneous intrusion and hydrothermal activity since the formation of a hydrothermal ore deposit is simply a function of the coincidence in time and space of hot fluids, active structures and reactive environments (Henley & Hoffman, 1987; in Henley & Etheridge, 1995). This is most effectively portrayed where episodic, violent boiling, restricted to local structural dilation sites supplied by accompanying hydraulic fracturing and hydrothermal eruptions (characteristic of geothermal systems in rhyolitic terrains), occurs. Such boiling of deeply circulating meteoric groundwater, with some magmatic fluid contribution, provides chalcedonic veining, episodic banding and brecciation textures, adularia stability, bladed calcite, and extreme loss of H_2S from the fluid to cause major gold deposition, potentially producing high grade "bonanza" gold mineralisation (Clarke, 1991).

The depth at which boiling commences becomes shallower with either decreasing temperatures and/or CO_2 concentration for a constant salinity (Figure 21). Solutions with the highest initial gas concentrations will tend to deposit gold and silver over the greatest vertical distance, and generally produce deposits with lower ore grades (Cole & Drummond, 1986). The depth of boiling also shifts dramatically depending on pressure. Thus, as pressure fluctuates owing to self-sealing and re-breaking, the depth interval of boiling changes, resulting in spatial overlapping of radically different chemical regimes. These effects could produce "mineralising" and "barren" stages in ore formation, as indicated by crosscutting vein relationships (Reed & Spycher, 1985).

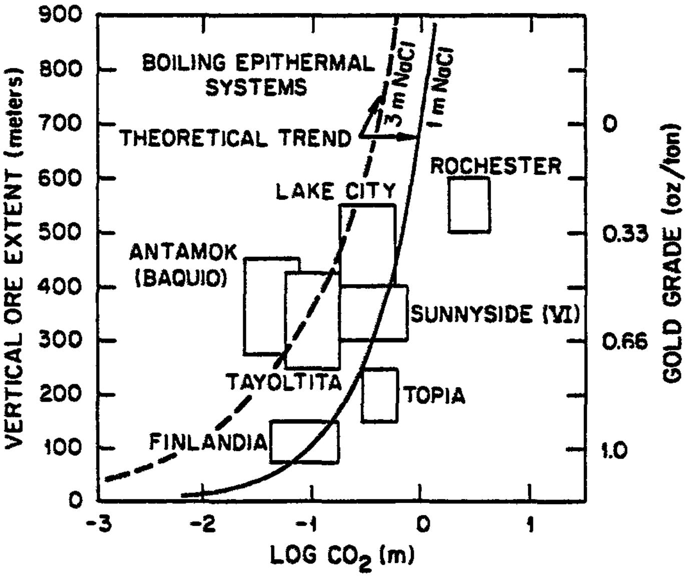


Figure 21. Vertical extent of ore deposition plotted against CO_2 concentration for five boiling epithermal systems. A secondary axis for the gold ore grade in these deposits is also shown. The sources of data are: Rochester - Vickre (1981); Lake City - Slach (1980), Sunnyside - Casadevall & Ohmoto (1977), Antamok (Baguro) - Sawkins *et al.* (1979), Toyoltita - Smith *et al.* (1982), Topia - Loucks & Sommer (1984), Finlandia - Kamilli & Ohmoto (1977). The theoretical trends of vertical extent of deposition versus CO_2 content are given for open, isothermal boiling conditions at 250°C, $\Sigma H_2S=0.001$ m, $\Sigma SO_4=10^{-5}$ m, and $\Sigma Cl^- \approx 1.0$ and 3.0 m (after Cole & Drummond, 1986).

Buchanan (1981), recognised four profound changes in the physical and chemical state of fluids due to boiling:

i) Significant amounts of CO_2 and usually lessor amounts of H_2S are partitioned into the vapour phase, according to the simple reactions:

$HCO_3^- + H^+$	=	CO_2 (vap.) + H_2O
$HS^- + H^+$	=	H_2S (vap.)

This release of volatiles results in a pH rise in the remaining solutions;

ii) The salinity of the remaining solutions will rise, a result of simple concentration of salts by the loss of H_2O steam;

iii) Oxygen fugacity in the remaining liquid increases (Drummond & Ohmoto, 1979; in Buchanan, 1981); and

iv) Major loss of CO_2 and lessor loss of H_2S results in a rise in the activity of $S^=$ and HS^-, thus leading to the formation of strong thio- complexes with Au, As, Sb, and Hg (Weissberg, 1969; in Buchanan, 1981).

These results of boiling cause first the base metals, then silver sulphides, and later gold to deposit in a well-recognised temporal and vertical sequence (Buchanan, 1981). Gold mineralisation at Summitville, for example, occurs in association with the shallower covellite-bearing assemblage but not with the deeper chalcopyrite assemblage. Gold also occurs in a near-surface barite + jarosite + goethite assemblage. This association of gold with assemblages indicative of relatively oxidising and acidic conditions, confirms that gold deposition occurred in response to destruction of gold-bisulphide complexes, due to an increase in acidity and oxygen fugacity with increasing elevation in the system (Stoffregen, 1987).

1.8 LOCALISING CONTROLS ON EPITHERMAL ORE-BODY FORMATION

Modern active geothermal systems are analogues of the ancient systems that formed those epithermal deposits discovered to date. The model for these systems requires that fluids be focused along structures *e.g.* faults, and be brought to the near surface environment where a 2-phase (liquid-gas) boiling zone occurs. It is the complexities of this zone in epithermal deposits that results in the observed mineralisation and alteration patterns (Swindell, 1986). Selected examples of structural, hydrothermal and lithological controls on permeability are illustrated in (Figure 22).

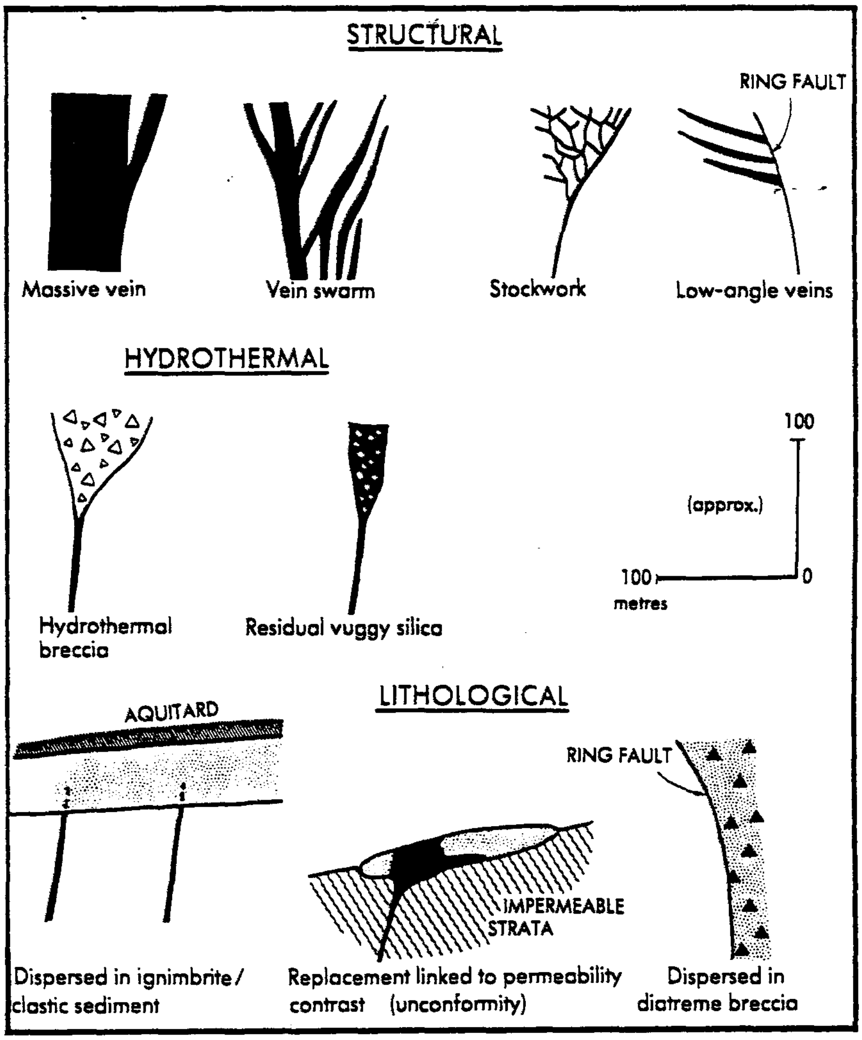


Figure 22. Selected styles and geometries of epithermal deposits to illustrate influence of structural, hydrothermal and lithological permeability. Typical examples: massive vein - Umuna (Papua New Guinea); vein swarm - Waihi (New Zealand); stockwork - Bellavista (Costa Rica); shallowly dipping veins abutting a ring fault - Emperor (Fiji); hydrothermal breccia - Choquelimpie (Chile); residual vuggy silica - Kasuga (Japan); dispersed in permeable rock beneath an aquitard - Round Mountain (Nevada); replacement linked to permeability contrast -L Coipa (Chile); dispersed in diatreme breccia - Montana Tunnels (Montana) (after Sillitoe, 1993).

The major factor in concentrating gold is the nature and efficiency of the hydrothermal system (Swindell, 1986), which is directly proportional to permeabilty. Permeability controls fluid flow and precious-metal deposition (Stoffregen, 1987; in Sillitoe, 1993), and has been crucial in controlling emplacement of several large, in part strata-bound orebodies, especially in shallow parts of epithermal systems. At Hishikari in Japan for example, ore deposition is controlled by contacts, commonly regional unconformities, between less permeable strata below and more permeable sequences above. Precious-metal deposition in the vicinity of the permeability boundary may be attributed to mixing of ascending, fracture-confined metalliferous fluids with cooler, more oxidised meteoric water, as well as boiling induced upon encountering the more permeable, lower pressure rock environment (Sillitoe, 1993). The evolution of the hydrothermal system responsible for forming a Nansatsu-type ore deposit incorporates at least two episodes of enhanced permeability (Figure 23).

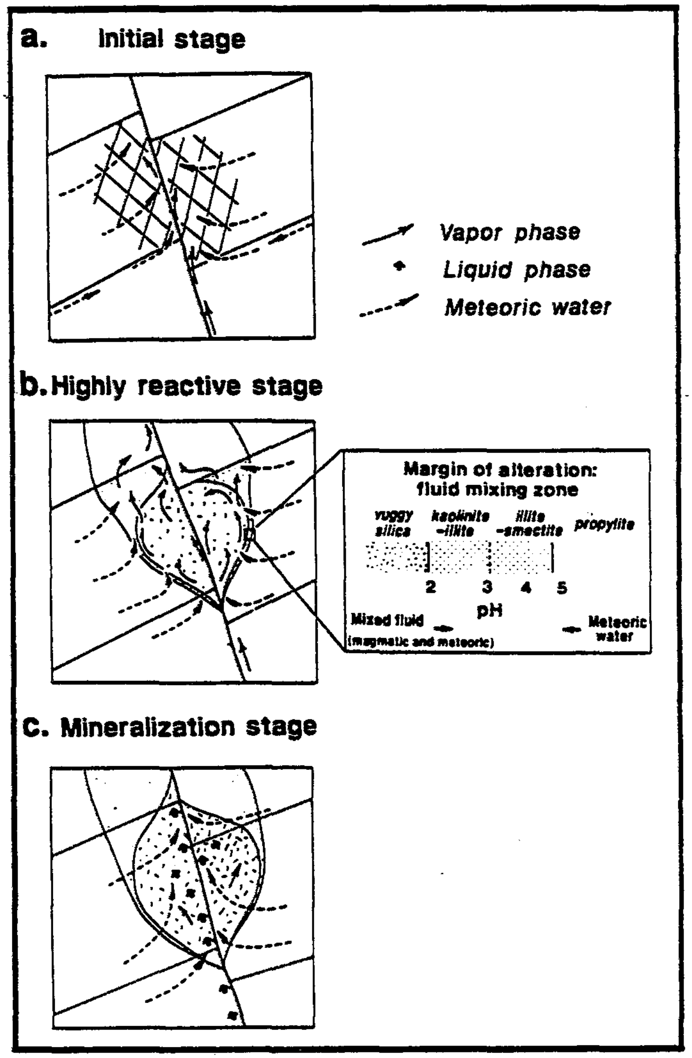


Figure 23. Modelling of the different stages of evolution of the hydrothermal system responsible for forming a Nansatsu-type ore deposit (from White, 1991). *(a)* Magmatic gases (HCl, SO_2, H_2S, etc.) initially rise along a fracture in a low-density vapour and condense into meteoric water, heating and acidifying the water. *(b)* This leads to the leaching of the host rock and development of a neutralization halo (inset), with increased permeability favouring more meteoric water involvement. *(c)* Subsequently a magmatic, metal-bearing saline liquid reaches the leached zone. Mixing with meteoric water causes mineral deposition and mineralisation. Activity in the system subsequently wanes, leading to supergene processes becoming dominant, locally remobilising metals (after Hedenquist *et al.*, 1994).

Most HS- and LS-type epithermal gold deposits form within the lower part and sometimes at the base of an eroding stratovolcano and its epiclastic products. Gold deposition in LS and possibly some HS systems in these settings, probably results largely from fluid mixing in the more permeable volcanic rocks (Mitchell, 1992). The stratigraphic position of the HS-type deposits referred to by Mitchell, 1992 - at or within a few hundred meters of an andesite-basement unconformity - suggests that they develop beneath the near-neutral pH hydrothermal water-table rather than in advanced argillic zones in perched aquifers high in the volcanic structure (Figure 24(b)). Stratigraphic reconstructions indicate that during mineralisation all or most epithermal systems intersected a regional unconformity (Mitchell & Balce, 1990).

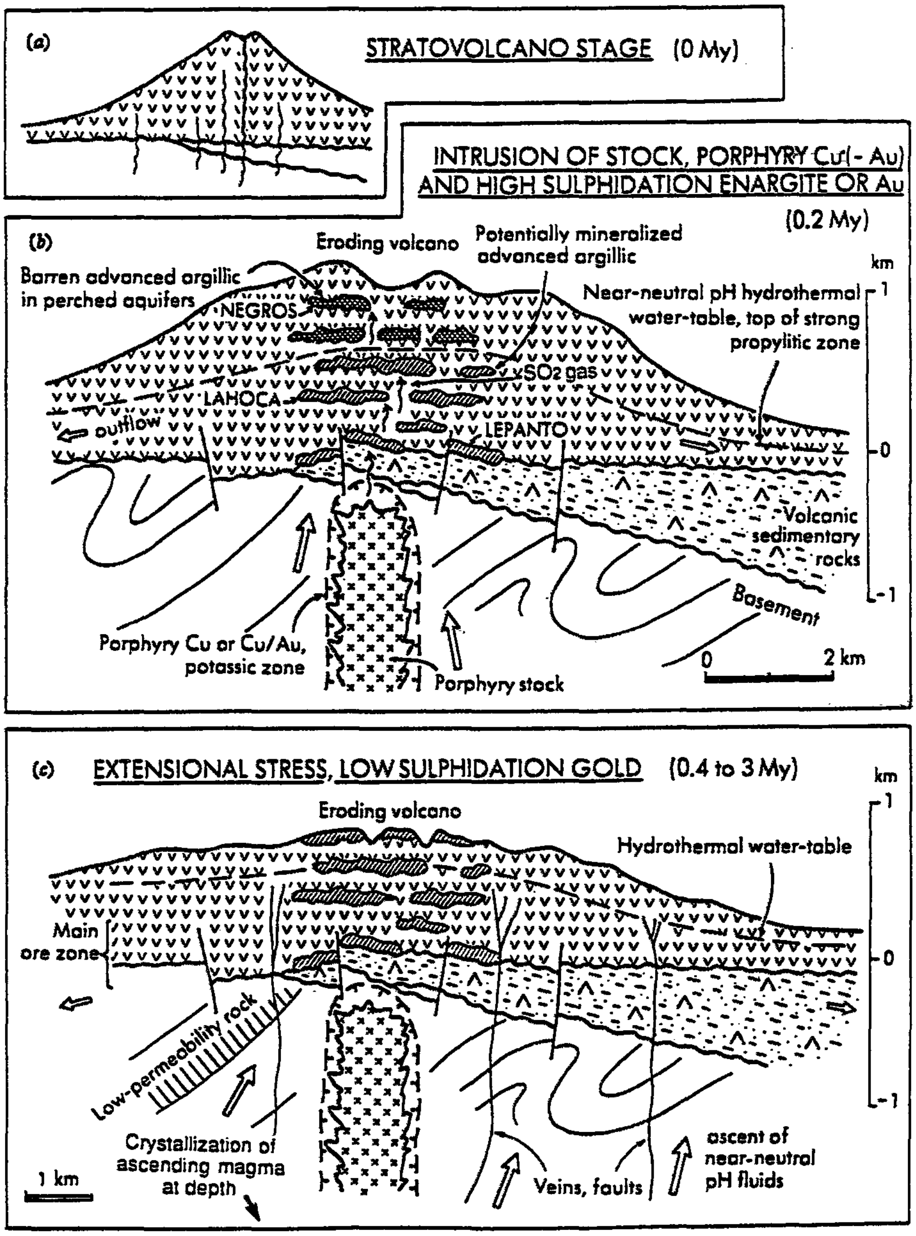


Figure 24. Schematic sequence of events in mineral district with both porphyry and epithermal mineralisation, showing *(a)* eruption of stratovolcano, followed by *(b)* advanced argillic alteration, high-sulphidation and porphyry mineralisation; and *(c)* subsequent low-sulphidation gold mineralisation. Time shown is from end of generation of stratovolcano (after Mitchell, 1992).

Epithermal mineralisation is confined to extensional or transtensional settings. A typical example of an extensional setting is presented by the conceptual block model of the depositional environment during formation of the Grouse Creek deposit (Figure 25).

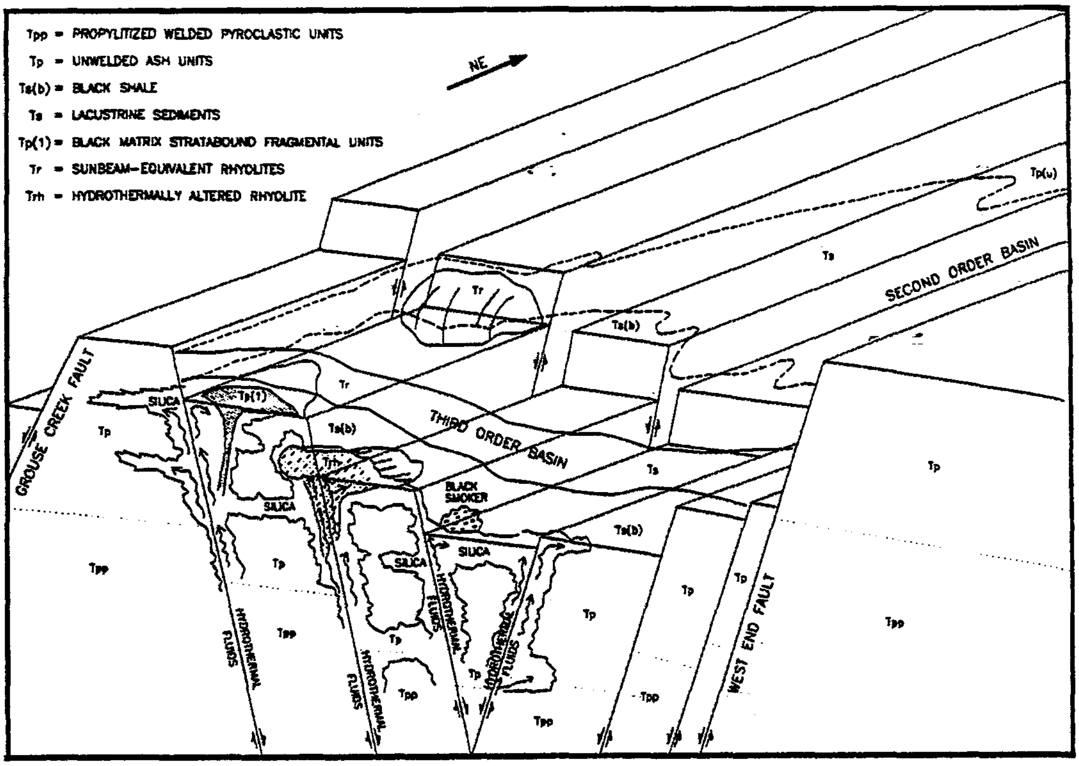


Figure 25. Conceptual block model of the depositional environment during formation of the Grouse Creek deposit. Tp=unwelded ash units, Tp(1)=black matrix strata-bound fragmental units, Tpp=propylitized welded pyroclastic units, Tr=Sunbeam-equivalent rhyolites, Trh=hydrothermally altered rhyolite, Ts=lacustrine sediments, Ts(b)=black shale (after Allen & Hahn, 1994).

Dilational fault jogs or releasing bends on strike-slip faults (Figure 26), provide especially favourable sites for epithermal mineralisation (Sibson, 1987; in Sillitoe, 1993). Ring faults that delimit calderas or diatremes also localise a number of epithermal deposits (Sillitoe & Bonham, 1984; in Sillitoe, 1993). Figure 27 presents a schematic representation of a typical mineralised caldera setting. Intersections between district-wide normal faults and diatreme ring fractures are particularly favourable; such intersections control the gold-bearing breccias pipes at Acupan in the Baguie district (Damasco & de Guzman, 1977; in Sillitoe, 1993) and the Lepanto enargite-gold replacement/breccia deposit (Sillitoe, 1983; in Sillitoe, 1993), both in the Philippines. In general, veins postdate the formation of any caldera and occur either inside, outside and/or across the caldera margin (Mitchell, 1992).

The major requirement for mineralisation in almost all of the Philippine deposits appears to be emplacement of andesitic to dacitic minor intrusions, with the two largest gold districts of Baguio and Paracale situated where magmatism coincides with basement antiforms. Hamilton (1988; in Mitchell & Leach, 1991) has pointed out that in island arcs the axis of the magmatic arc is a regional geoanticline, considered to reflect magmatic inflation and

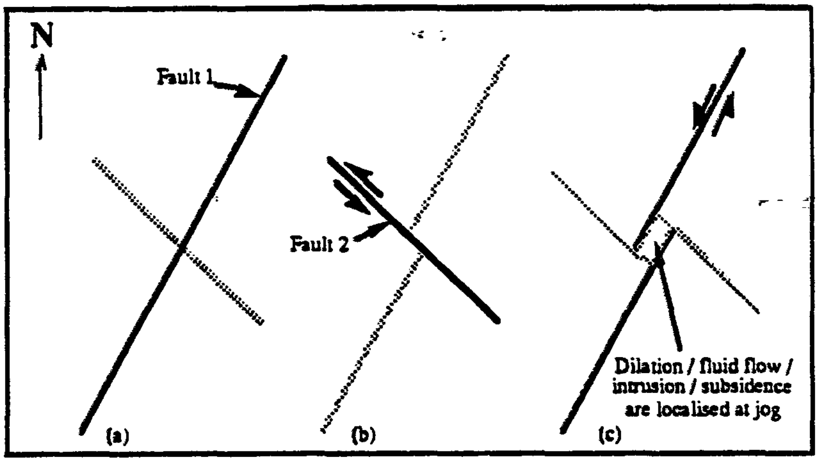


Figure 26. Reactivated fault intersections are commonly the locus of dilation, fluid flow, intrusion and/or local subsidence. In this sketch, Fault 1 is cut and offset by Fault 2 (Figure 26b). Subsequent reactivation of Fault 1 gives rise to dilational jog at the fault intersection (Figure 26c). Where both faults are vertical, the dilational jog forms a vertical "pipe". Volcanic/intrusive breccia pipes and associated deposits and a number of the so-called maar-type deposits may form in such environments (after Henley & Etheridge, 1995).

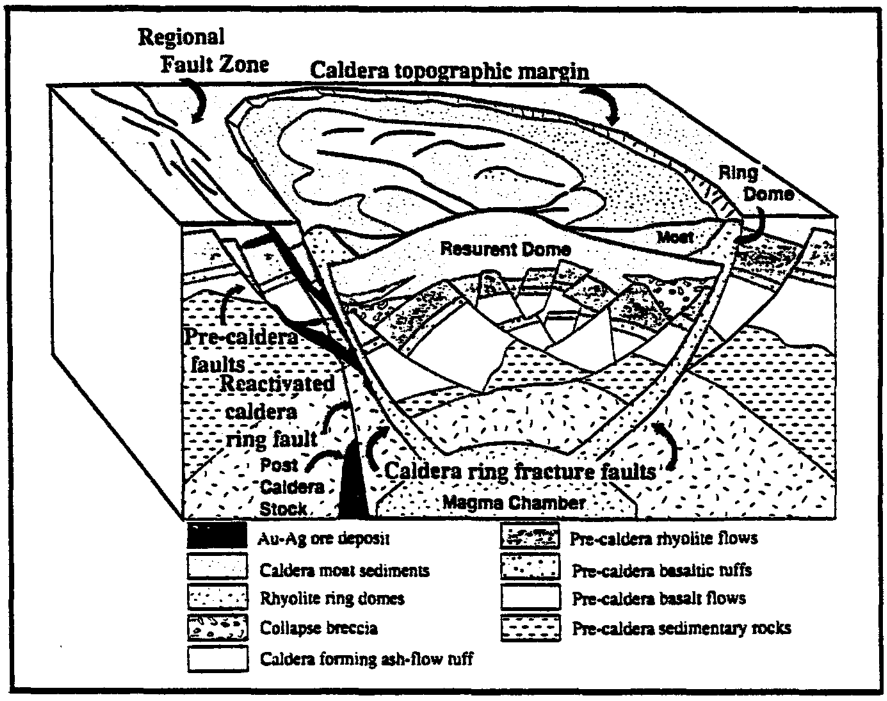


Figure 27. Schematic block diagram of reactivation of caldera ring fracture faults and precaldera faults which may be related to tumescence by a regional fault zone. Hydrothermal ore deposits are localised along these reactivated caldera structures and may extend outside of the ring fracture zone (after Rytuba, 1994).

thermal expansion. Tension faults or gashes are commonly present in the core of antiforms or anticlines, forming channels which focus fluid flow within the fold structure. Consequently, rising hydrothermal fluids are confined to fractures and focused towards the axis of the structure.

PART B: CLASSIFICATION AND CHARACTERISTICS OF EPITHERMAL PRECIOUS METAL SYSTEMS

2.1 INTRODUCTION

Epithermal ore deposits 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 (Silberman & Berger, 1985). They form in a dynamic environment in fluctuating and evolving geothermal systems, and it is not always appropriate to classify deposit types, but rather mineralisation types which might comprise a given deposit (Clarke, 1991). Ever since the term epithermal was defined (Table 2), (Lindgren, 1922; in Heald *et al.*, 1987), geologists have tried to divide the category into descriptive or genetic subsets.

Table 2: Characteristics of epithermal systems (from Silberman and Berger, 1986, after Lindgren, 1933; in Mitchell & Leach, 1991).

Depth of formation	Surface to 1 000m.
Temperature of formation	50-300°C.
Form of deposits	Thin to large veins, stockworks, disseminations, replacements.
Ore textures	Open-space filling, crustification, colloform banding, comb structure, brecciation.
Ore elements	Au, Ag, As, Sb, Hg, Te, Tl, U, Pb, Zn, Cu.
Alteration	Silicification, argillization, sericite, adularia, propylitization.
Common features	Fine-grained chalcedonic quartz, quartz pseudomorphs after calcite, brecciation.

Many attempts have been made to classify epithermal deposits based on mineralogy and alteration, the host rocks, deposit form, genetic models, and standard deposits. All have their strengths and weaknesses (White & Hedenquist, 1990). Classification schemes have been based upon plate tectonic concepts (Giles & Nelson, 1982; Bonham & Giles, 1981; Sawkins, 1984; in Silberman & Berger, 1985; and Sillitoe, 1981), association with volcanic landforms (Henley & Ellis, 1983; Sillitoe & Bonham, 1984; in Silberman & Berger, 1985), physical and mineralogical characteristics (Lindgren, 1933; Giles & Nelson, 1983; Heald-Wetlaufer *et al.*, 1983; Hayba *et al.*, 1985; in Silberman & Berger, 1985), and associated magma types and mineralogy (Bonham, 1986; in Silberman & Berger, 1985). Berger & Eimon (1983; in

Silberman & Berger, 1985) proposed conceptual models of volcanic-hosted epithermal systems based on a combination of physical, mineralogical, and hydrological characteristics. Some students of epithermal deposits fit all varieties of them into a generic model, flexible enough to fit all the variations (Buchanan, 1981; Silberman, 1982; in Silberman & Berger, 1985).

The two original characteristics used to classify deposits as epithermal were mineralogy (of both veins and hydrothermal alteration), and texture. Both these are readily observed, thereby providing an accessible basis for classification. Both also provide information pertaining to mineralising conditions, and from which temperature and, in some cases, depth may be inferred. For the explorationist, the most useful classification schemes should be brief, simple, descriptive, observationally based, and informative (White & Hedenquist, 1990).

Ultimately, no matter what classification scheme is utilised, there remain two distinct styles of epithermal gold deposits which are readily recognised. The best known style is variously referred to as adularia-sericite (Hayba *et al.,* 1985; Heald *et al.,* 1987), low sulphur (Bonham, 1986; in White *et al.*, 1995), or low-sulphidation style (Hedenquist, 1987; in White *et al.*, 1995). The second style is variously referred to as acid sulphate (Heald *et al.,* 1987), high sulphur (Bonham, 1986; in White *et al.*, 1995), high-sulphidation (Hedenquist, 1987; in White *et al.*, 1995), or kaolinite-alunite style (Berger & Henley, 1989; in White *et al.*, 1995), (Table 3). The terms high-sulphidation (HS) and low-sulphidation (LS) are based on the sulphidation state of associated sulphide minerals, which, along with characteristic hydrothermal alteration, reflect fundamental chemical differences in the epithermal environment. The distinguishing characteristics among epithermal deposits in volcanic rocks are tabulated in Table 4. Characteristics of both styles are summarised in Table 6 (Section 2.2).

Table 3: Recent Alternative nomenclature for epithermal types (after Sillitoe, 1981).

Enargite-gold		Ashley (1982)
High sulphur	Low sulphur	Bonham (1986, 1988)
High sulphidation	Low sulphidation	Hedenquist (1987)
Acid sulphate	Adulara-sericite	Hayba *et al.* (1985) Heald *et al.* (1987)
Alunite-kaolinite	Adularia-sericite	Berger and Henley (1989)

Table 4: Summary of Distinguishing Characteristics among Epithermal Deposits in Volcanic Rocks (after Heald *et al.*, 1987).

Acid-sulfate type	Andularia-sericite type
Higher sulfur mineral assemblage	Lower sulfur mineral assemblage
Hypogene alunite	No hypogene alunite
Alteration: advanced argillite →argillite (± sericite)	Alteration : sericite →argillite
No adularia	Adularia
Noteworthy Cu production	Variable base metal production
Rhyodactite host typical	Silicic to intermediate host
Similar ages, host and ore	Ages of host and ore distinct (>1 m.y.)
Relatively small physical size of productive area	Variable physical size; some very large

2.2 CLASSIFICATION OF EPITHERMAL PRECIOUS METAL DEPOSIT STYLES

Classification into high-sulphidation (HS-) and low-sulphidation (LS-) systems expresses a fundamental characteristic of the deposit, which in most cases can be easily inferred on the basis of simple observations, or at most supported by simple laboratory observations of mineralogy. It has implications for the mineralogy, texture, alteration and mineral zoning, as well as the genesis of the deposit, some aspects of which we are only beginning to understand (White & Hedenquist, 1990).

The classification, combined with a simple description of the form of the deposit (vein, stockwork, disseminated), conveys a large amount of information on mineralogy, alteration, and spatial characteristics of the mineralisation, and allows inferences to be drawn regarding likely regional controls, and the characteristics of the ore-forming fluids. Comparison with a relatively well-known example (if one exists) is an established and valuable way to convey a large amount of information by analogy. For example: (i) Low-sulphidation vein deposits comparable to Hishikari; (ii) High-sulphidation disseminated lode deposit, *cf.* Chinkuashih; (iii) High-sulphidation vein deposit, *cf.* El Indio; (iv) Low-sulphidation stockwork deposit, similar to McLaughlin; *etc* (White & Hedenquist, 1990).

Both the HS- and LS-styles of mineralisation are widespread (Table 5), occurring principally in convergent tectonic settings, and both have examples of major economic significance. Although the two styles show similar alteration mineralogies, the distribution of the alteration zones is different, and the economic mineralisation is associated with different parts of the system. The alteration zoning can be used as a pointer towards the most prospective part of the system, but only when the style has been correctly diagnosed (White & Hedenquist, 1995).

Table 5: Examples of epithermal gold deposits (after White & Hedenquist, 1995).

LOW SULPHIDATION	HIGH SULPHIDATION
McLaughlin, California, USA	Goldfield, Nevada, USA
Round Mountain, Nevada, USA	Summitville, Colorado, USA
Hishikari, Japan	Iwato, Kasuga and Akeshi, Japan
Emperor, Fiji	La Coipa, Chile
Golden Cross, New Zealand	El Indio, Chile
Waihi, New Zealand	Pueblo Viejo, Dominican Republic
Lebong Tandai, Indonesia	Chinkuashih, Taiwan
Kelian, Indonesia	Rodalquilar, Spain
Porgera Zone VII, Papua New Guinea	Lepanto, Philippines
Pajingo, Australia	Lahóca, Hungary

The general characteristics of both styles of mineralisation are summarised in Table 6, whilst the field characteristics of the two types are summarised in Table 7. HS deposits contain sulphide minerals with a high sulphur/metal ratio (*e.g.* enargite, luzonite, covellite) and are enveloped by advanced argillic assemblages, which are dominated by alunite but may include prominent pyrophyllite at deeper levels. In contrast, LS deposits contain sulphide assemblages indicative of a relatively low sulphidation state in association with sericitic, intermediate argillic or, uncommonly, chloritic alteration. The two deposit types are generated by fundamentally different, ascendant fluids: HS type from acid, sulphur-rich, oxidised fluids; and LS type from near-neutral, sulphur-poor, reduced fluids (Heald *et al.,* 1987). HS fluids are generated by condensation of SO_2-rich magmatic volatiles (Heald *et al.,* 1987) in the presence of varying amounts of meteoric water, whereas the meteoric-water-dominated LS fluids are suggested to possess a volumetrically lesser magmatic contribution of either acid volatiles or low-salinity brines (Sillitoe, 1993).

Table 6: Characteristics of Low- and High-sulphidation epithermal precious metal deposits (after Henley, 1991; Mitchell & Leach, 1991; White *et al.*, 1995; and White and Hedenquist, 1990).

	Low-sulphidation	High-sulphidation
Structural setting	Structurally complex volcanic environments, commonly in calderas.	Intrusive centres, 4 out of the 5 studied related to the margins of calderas.
Size	Variable; some large	Relatively small
length/width ratio	Usually 3:1 or greater	Equidimensional
Host rocks	Silicic to intermediate and alkalic volcanics; and underlying basement rocks of any type.	Rhyodacite typical
Localising controls	Any faults or fracture zones especially closely related to volcanic centres.	Major regional faults or subvolcanic intrusions.
Depth of formation	Mostly 0 to 1 000m	Mostly ?500 t0 ?2 000m
Timing of ore and host	Ages of host and ore distinct (>1 Ma)	Similar ages of host and ore (<0.5 Ma)
Character of mineralisation	Ore mineralisation characterised by open space and cavity filling, typically with sharp-walled veins. Layered vein fillings typical, commonly with multi-stage brecciation.	Ore mineralisation typically disseminated, either in white mica-pyrophyllite, or in massive silica. Open space and cavity filling not common.
	Near-surface may be stockwork or disseminated, depending on nature of local primary and secondary permeability.	Mineralisation usually associated with advanced argillic alteration, pyrite typically very abundant.
Characteristic textures	Crustification banding, fine comb texture, colloform banding, banded quartz-chalcedony, drusy cavities, vugs, vein breccia, sillica pseudomorphs after bladed calcite (lattice texture).	Vuggy silica (fine-grained quartz)
Mineralogy	Argentite, tetrehedrite, 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
Form of deposits	Open space veins dominant Stockwork veining common Disseminated ore mostly minor Replacement ore minor.	Veins mostly subordinate Stockwork veining minor Disseminated ore common Replacement ore common
Dominant metals	Ag, Au, As, Hg	Cu, Ag, Au, As
Minor metals	Zn, Pb, Sb, Se	Pb, Hg, Sb, Te, Sn, Mo

Table 6: Continued

	Low-sulphidation	High-sulphidation
Production data	Both gold- and silver-rich deposits, variable base-metals	Both gold- and silver-rich deposits, noteworthy Cu production
Alteration	Sericitic 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
Salinity	0-13 wt.% NaCl eq.	1-24 wt.% NaCl eq.
Source of fluids	Dominantly meteoric.	Dominantly meteoric, possibly significant magmatic component.
Source of sulfide sulfur	Deep-seated, probably 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
Examples	Pajingo, Australia Emperor, Fiji Lebong Donok, Indonesia Wapolu, Papua New Guinea Acupan, Philippines Golden cross, New Zealand	Temora, Australia Mount Kasi, Fiji Motomboto, Indonesia Nena, Papua New Guinea Lepanto, Philippines Nansatu district, Japan

Deposits of both types exist within single districts, but they are localised by different structures. For example, relatively minor LS veins exist alongside major HS precious-metal deposits (Figure 28A,D) at Lepanto in the Philippines (Garcia, 1991; in Sillitoe, 1993), Chinkuashih in Taiwan (Huang, 1955; in Sillitoe, 1993), and Choquelimpie in Chile (Sillitoe, 1991; in Sillitoe, 1993). LS epithermal deposits are found in association with a broad spectrum of volcanic rock compositions. In contrast, most HS deposits are genetically linked to subalkalic rocks ranging from andesitic to rhyodacitic in composition (Bonham, 1986; 1988; in Sillitoe, 1993).

As a corollary, epithermal deposit types and subtypes are broadly controlled by tectonic setting (Sillitoe, 1992; in Sillitoe, 1993). HS deposits are restricted largely to normal, subduction-related, volcano-plutonic arcs in island arcs and along continental margins. Sulphide-rich LS deposits tend to be found in comparable arc settings. In contrast, sulphide-poor LS deposits hosted by either subalkalic rhyolitic or alkalic rocks are generally generated in a variety of extensional settings, but commonly in arc terrains during or immediately following subduction.

Table 7: Field characteristics for distinguishing epithermal types (after Sillitoe, 1993)

	High Sulphidation (HS)	Low Sulphidation (LS)
Genetically related volcanic rocks	Mainly andesite-rhyodacite	Andesite-rhyodacite-rhyolite
Alteration zone	Areally extensive (commonly several km^2) and visually prominent	Commonly restricted and visually subtle
Key proximal alteration mineral(s)	Crystalline alunite; pyrophyllite at deeper levels	Sericite or illite ± adularia; roscoelite (V-mica) in deposits associated with alkalic rocks; chlorite in few cases
Quartz gangue	Fine-grained, massive, mainly replacement origin; residual, slaggy ("vuggy") quartz commonly hosts ore	Chalcedony and (or) quartz displaying crustiform, colloform, bladed, cockade and carbonate- replacement textures; open-space filling
Carbonate gangue	Absent	Ubiquitous, commonly manganoan
Other gangue	Barite widespread with ore; native sulphur commonly fills open spaces	Barite and (or) fluorite present locally; barite commonly above ore
Sulphide abundance	10-90 vol.%, mainly fine-grained, partly laminated pyrite	1-20 vol.%, but typically <5 vol.%, predominantly pyrite
Key sulphide species	Cu sulphosalts (enargite, luzonite) and Cu+Cu-Fe sulphides (chalcocite, covellite, bornite) common; generally later than pyrite	Sphalerite, galena and tetrahedrite common. Cu present mainly as chalcopyrite
Metals present	Cu, Au,, As (Ag, Pb)	Au and (or) Ag (Zn, Pb, Cu)
Metals present locally	Bi, Sb, Mo, Sn, Zn, Te (Hg)	Mo, Sb, As (Te, Se, Hg)

Note: Based on personal observations plus Bonham (1986, 1988), Berger and Henley (1989), Henley (1991), and White and Hedenquist (1990).

LS deposits show a wide variety of textures, including banded, crustiform quartz and chalcedony veins, druse-lined cavities, and spectacular, multiple-episode vein breccias (Berger & Eimon, 1983; in White & Hedenquist, 1995). Lattice-textured bladed calcite is common and formed as a result of boiling (Simmons & Christenson, 1994), although it may be replaced by quartz as the system cools. In areas that have experienced little erosion, distinctive silica sinters deposited at the palaeosurface by neutral-pH hot-spring waters may still be present (Vikre, 1985; White *et al.*, 1989; in White & Hedenquist, 1995).

The typical textures of HS deposits in contrast show relatively little variation, with the most characteristic texture being massive bodies of vuggy quartz typical of the Summitville (Gray & Coolbaugh, 1994) or Nansatsu deposits (Hedenquist *et al.*, 1994; in White & Hedenquist, 1995). Massive to banded sulphide veins consisting of pyrite and enargite may also cut the relatively oxidised, acidic fluids; or low-sulphidation (adularia-sericite) type, where assemblages containing illite or sericite ± adularia accompanied by low-sulphidation sulphides are generated by reduced, neutral-pH fluids (Hedenquist, 1987; Heald *et al.*, 1987; in Sillitoe, 1994).

2.3 LOW-SULPHIDATION STYLE EPITHERMAL PRECIOUS METAL DEPOSITS

Low-sulphidation (LS) deposits in general form distant from the inferred magmatic heat source[11] at temperatures of 200 - 300°C (Figure 29). Pressures are controlled by hydrostatic conditions, meaning that the maximum temperature at a given depth is constrained by boiling, a common process in these systems (Hedenquist & Henley, 1985; in Hedenquist & Lowenstern, 1994). Isotopic studies have shown that the mineralising hydrothermal fluids in the LS environment (White & Hedenquist, 1990), are similar to those typically tapped by drilling into active geothermal systems (Henley & Ellis, 1983; in White & Hedenquist, 1995). Figure 30 depicts the schematic cross-section of a LS model based largely on deposits in Western USA.

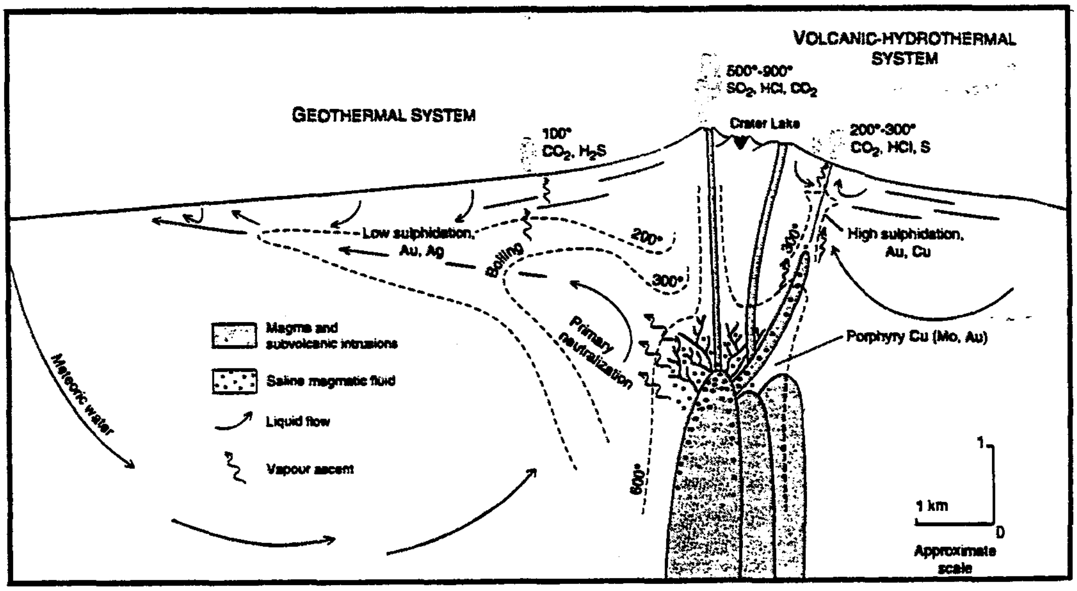


Figure 29. Schematic cross-section showing shallow, sub-volcanic intrusions (heat source), an associated stratovolcano, and the environments deduced for the formation of both porphyry Cu, and high- and low-sulphidation epithermal ore deposits. Active volcanic-hydrothermal systems extend from degassing magma to fumaroles and acidic springs, and incorporate the porphyry and/or high-sulphidation ore environments, whereas low-sulphidation ore deposits form from geothermal systems characterised by neutral-pH waters that may discharge as hot springs and geysers, such as those in Yellowstone National Park, USA (after Hedenquist & Lowenstern, 1994).

The principal ore minerals are native silver, argentite, parargirite, polybasite, stephanite, miargirite, native gold, electrum, tetrahedrite, chalcocite, and base-metal sulphides. Gangue minerals are quartz, calcite, dolomite, rhodochrosite, adularia, and pyrite (Ericksen, 1988).

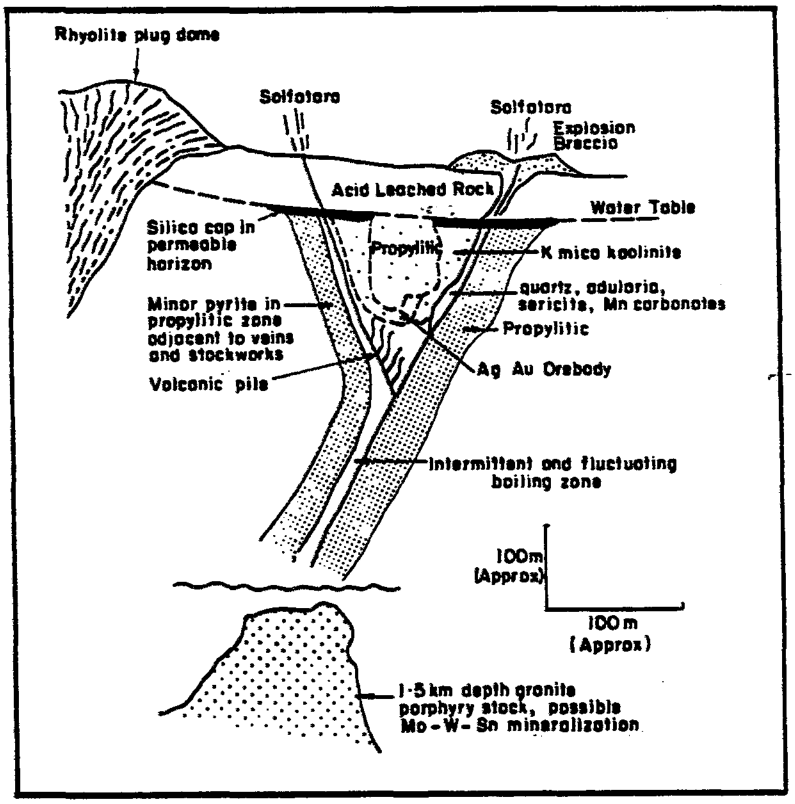
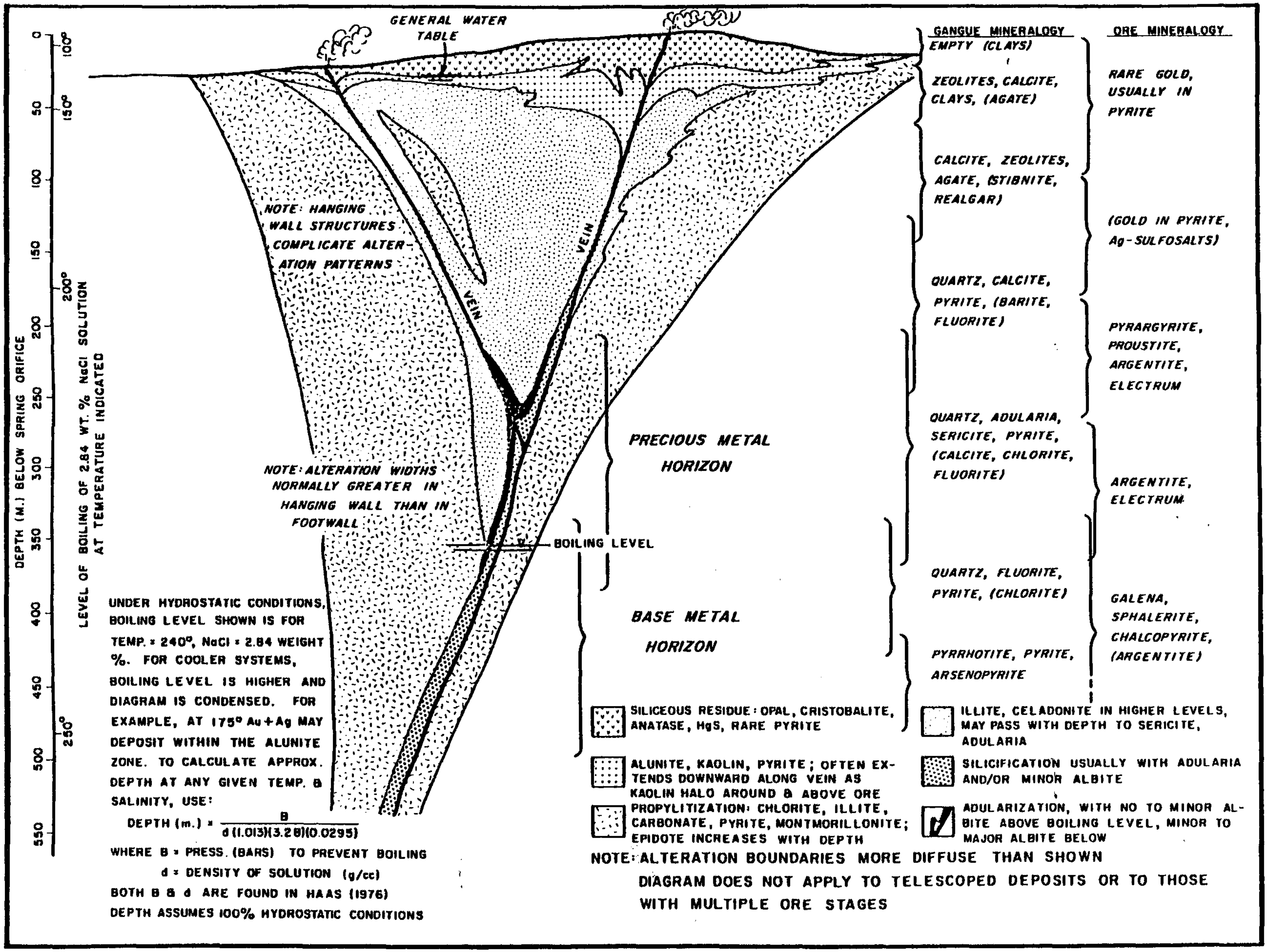


Figure 30. Schematic cross-section of low-sulphidation epithermal systems, based largely on deposits in Western USA (simplified from Bonham, 1986; in Mitchell & Leach, 1991).

LS precious-metal deposits are characterised by an alteration-mineral assemblage dominated by adularia and (or) sericite and a lack of hypogene acid-sulphate alteration and enargite (Heald *et al.*, 1987). A detailed schematic cross-section of the LS model showing the general form, alteration mineralogy, mineralised zones and ore and gangue mineralogy is presented in Figure 31. These deposits formed well after host-rock emplacement, by circulating, nearly neutral hydrothermal solutions that apparently are related to deep heat sources. Magmatic signatures in LS epithermal Au deposits are elusive (Hedenquist & Lowenstern, 1994). Examples of LS precious-metal deposits in the United States include Round Mountain, Nevada (Tingley & Berg, 1985; Shawe *et al.*, 1986; in Ericksen, 1988), Creede, Colorado (Steven & Eaton, 1975; Bethke *et al.*, 1976; Barton *et al.*, 1977; Bethke & Ryle, 1979; in Ericksen, 1988), and Bodie, California (Silberman, 1985; in Ericksen, 1988).

In the Western Pacific LS deposits are far more abundant than high-sulphidation deposits. There are more than 25 districts or deposits with past or planned production of more than 10 t Au and 10 with production plus reserves exceeding 1 000 000 oz Au. Many, but not all, epithermal districts include porphyry copper or copper-gold deposits, but no major LS veins are



UNDER HYDROSTATIC CONDITIONS, BOILING LEVEL SHOWN IS FOR TEMP. = 240°, NaCl = 2.84 WEIGHT %. FOR COOLER SYSTEMS, BOILING LEVEL IS HIGHER AND DIAGRAM IS CONDENSED. FOR EXAMPLE, AT 175° Au+Ag MAY DEPOSIT WITHIN THE ALUNITE ZONE. TO CALCULATE APPROX. DEPTH AT ANY GIVEN TEMP. & SALINITY, USE:

$$\text{DEPTH (m.)} = \frac{B}{d(1.013)(3.28)(0.0295)}$$

WHERE B = PRESS. (BARS) TO PREVENT BOILING

d = DENSITY OF SOLUTION (g/cc)

BOTH B & d ARE FOUND IN HAAS (1976)

DEPTH ASSUMES 100% HYDROSTATIC CONDITIONS

NOTE: ALTERATION BOUNDARIES MORE DIFFUSE THAN SHOWN

DIAGRAM DOES NOT APPLY TO TELESCOPED DEPOSITS OR TO THOSE WITH MULTIPLE ORE STAGES

Figure 31. A schematic cross-section of the typical low-sulphidation model showing the general form, alteration mineralogy, zones of mineralisation, ore and gangue mineralogy (after Buchanan, 1981).

superimposed on a porphyry deposit. Most LS-type deposits contain < 100 metric tons of gold (Sillitoe, 1988). The largest individual deposits are Ladolam (or Lihir), with 500 000 000 t at 4.5 g/t Au, and Porgera, which has reserves of 380 t Au and 800 t Ag, both in Papua New Guinea; Porgera includes an early stage of gold deposition with some features that are analogous to a porphyry system (Richards, 1992; in Mitchell, 1992). Hishikari in Southwest Japan, possibly the world's largest high-grade deposit, includes more than 2 000 000 t at 70 g/t Au (Izawa *et al.*, 1990; in Mitchell, 1992).

Kelian in Indonesian Borneo, a 75 000 000t deposit with a grade of 1.8 g/t, is classified as epithermal but has features that are transitional to a porphyry copper deposit (Van Leeuwen, 1990; in Mitchell, 1992). Other major producing regions include the world-class Baguio district (where production exceeds 800 t Au), the Paracale district and Eastern Mindanao Province, all in the Philippines; Central and Southern Sumatra; and the Highlands of Papua New Guinea. An example of the trend to large-tonnage, low-grade, open-pit mining of epithermal systems is provided by Baguio, where an open-pit mine recently opened on veins that have been mined commercially since 1917 (Mitchell, 1992).

At producing deposits the maximum vertical extent of ore-grade gold is about 800 m. High-grade bonanza zones, where they exist, usually lie in the upper part of the system within volcanic rocks. There is often a gradual decrease in gold grade and an increase of base-metal content with depth. In Europe base metals in the deeper parts of the veins often form lead-zinc orebodies, as in parts of the Carpathian arc (Ianovici & Romania, 1982; in Mitchell, 1992) and the Palaeogene Rhodope-central Serbian arc (Dakov, 1989; in Mitchell, 1992), but in the Western Pacific the base-metal content, which is predominantly lead-zinc, rarely exceeds 1wt%.

The typical LS-type deposits are base metal-poor veins, which commonly contain crustified chalcedonic quartz, manganoan carbonates (and rhodonite), and silver sulphides and sulphosalts. These deposits sometimes appear to form in geothermal systems where surficial ground waters mix with deeper, heated, nearly neutral pH, chloride brines in a lateral flow regime (Figure 32). The hydrology of these deposits is complex in that several different fluids may be involved in ore deposition and the characteristics of these fluids themselves change through mixing and boiling. At Creede, for example, three different fluids have been identified (Bethke & Rye, 1979; Foley *et al.,* 1982; in Heald *et al.*, 1987) and both boiling and mixing have been documented (Roedder, 1970; Hayba, 1984; Robinson & Norman, 1984; in Heald *et al.*, 1987).

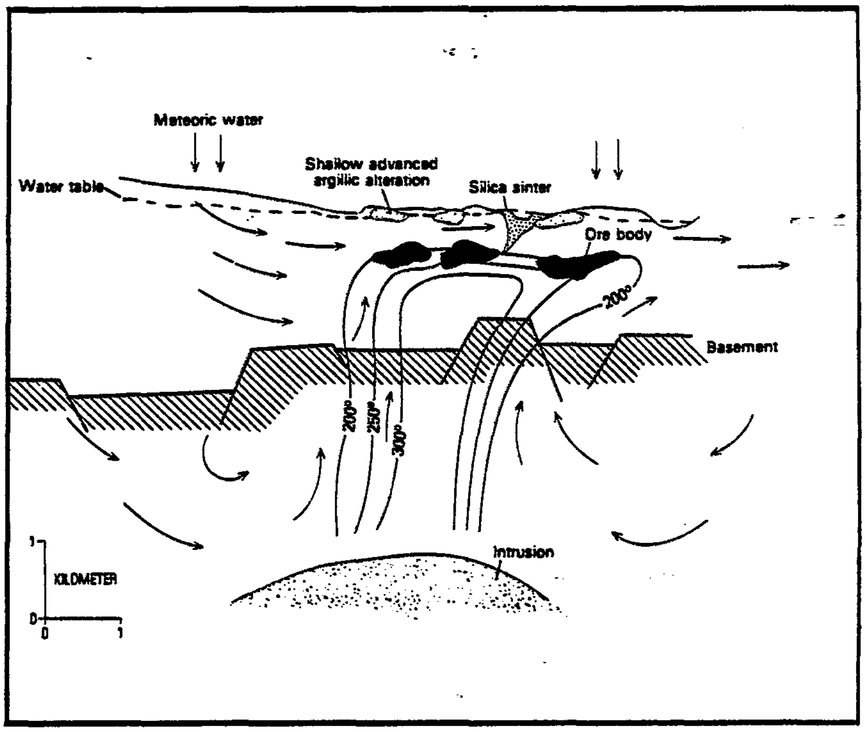


Figure 32. Occurrence model of low-sulphidation epithermal ore deposits in a geothermal system modified from models of geothermal systems from Henley & Ellis, (1983; Figure 5, in Heald *et al.*, 1987).

In general, both the vertical and lateral dimensions of the LS-type deposit are substantially more extensive than those of the HS-type deposit (Table 8), and some are extremely large (Guanajuato). The areal dimensions, especially of the larger deposits, require an extensive lateral flow regime, similar to those documented by drilling in many geothermal systems (Healy & Hochstein, 1973; Hedenquist, 1983; in Mitchell, 1992). The physical characteristics of an area, such as density of faulting, size of faults, and topography, are probably most significant in governing the size of the LS-type deposit.

Table 8: Approximate Physical Dimensions of 16 Epithermal Districts (after Heald *et al.*, 1987)

District	Length: main ore zone (km)	Width: main ore zone (km)	Lateral extent (km^2)	Vertical extent (m)
Red Mtn, CO	3.5	3	10.5	up to 500
Julcani, Peru	5	2.5	15	450
Lake City II, CO	9	3	27	at least 200
Summitville, CO	1.5	1	1.5	400
Goldfield, NV	2	1.5	3	410

Table 8: Continued

District	Length: main ore zone (km)	Width: main ore zone (km)	Lateral extent (km^2)	Vertical extent (m)
Lake City I, CO	13	3	39	at least 200
Colqui, Peru	8	2	16	400-480
Eureka, CO				up to 900
Creede, CO	8	3.5	28	400
Guanajuato, Mexico	21	9	189	at least 700
Pachuca, Mexico	11	8	88	up to 600
Tonopah, NV	4	1	4	180-300
Comstock, NV	10	4	40	1 000
Silver City, ID	6	3	18	750
De Lamar, ID	2	1	2	750
Round Mtn., NV	2.5	1	2.5	500
Oatman, AZ	6	2	12	90-400

CO = Colorado
NV = Nevada
ID = Idaho
AZ = Arizona

Two types of LS epithermal mineralisation may be recognised in volcanic terrains (Clarke, 1991). Chalcedonic Silica-Adularia-Illite type mineralisation is most common in rhyolitic terrains, whereas Crystalline Quartz-Illite-(Adularia-Kaolinite) type mineralisation is more typically formed in andesitic terrains. The Chalcedonic Silica-Adularia-Illite type of mineralisation typically shows open-space-filling, crustiform, banded veining with common chalcedony veins and quartz-adularia veins. Other common textures are coliform banding, fine comb quartz bands, banded quartz-chalcedony, quartz-filled vugs and cavities, and lattice-textured calcite with silica pseudomorphs. Hydrothermal brecciation is usual. Intense alteration adjacent to veins is represented by quartz, adularia and illite (sericite). Crystalline Quartz type deposits on the other hand mostly consist of coarse buck or comb, commonly amethystine, quartz (Gadsby *et al.,* 1990; in Clarke, 1991). Illitic clays dominate alteration, and kaolinitic clays are common. Intense alteration adjacent to veining is marked by pyrite-rich silicification. Adularia is not a common phase in altered wallrock. Zones of gold mineralisation approaching economic interest are restricted to small "shoots". The field characteristics of the LS epithermal subtypes are tabulated in Table 9.

Table 9: Field characteristics of low-sulphidation (LS) epithermal subtypes (after Sillitoe, 1993).

Principal volcanic host rock	Alkalic	Sub-alkalic, rhyolite	Sub-alkalic, andesite-rhyodacite
Sulphide content (principally pyrite, sphalerite, galena)	Low (<5 vol.%)	Low (<5 vol.%)	Relatively high (>10 vol.%)
Diagnostic alteration minerals	Roscoelite, adularia and fluorite (except in basic rocks)	Sericite or illite and adularia	Sericite (adularia generally absent)
Quartz type	Chalcedonic and crystalline	Chalcedonic and crystalline	Crystalline
Principal vein (gangue) texture	Crustiform	Crustiform	Massive, comb-textured, crudely banded.
Sulphide-gangue relations	Disseminated grains, concentrated preferentially in certain quartz bands	Disseminated grains, concentrated preferentially in certain quartz bands	Massive or semi-massive bands alternating with sulphide-poor gangue
Base-metal content	Low (say, <0.1%)	Low (say, <0.1%)	High (commonly serveral %)
Gold fineness	High	Low	Intermediate to high
Minor components	Te ubiquitous	Se common	Se and Te generally absent
Type examples	Emperor (Fiji), Cripple Creek (Colorado)	Round Mountain (Nevada), Mequite (California), Hishikari (Japan)	Comstock (Nevada), Umuna (Papua New Guinea), El Bronce (Chile)

The characteristics of many of the LS deposits of the Philippines and Indonesia are consistent with deposits formed in the andesitic stratovolcanic setting. Active geothermal systems in the Philippines, considered to be analogous to the stratovolcanic setting, have been described by Reyes (1985; 1990; in Heald *et al.*, 1987), Reyes & Giggenbach (1992; in Heald *et al.*, 1987), and Reyes *et al.,* (1993; in Heald *et al.*, 1987). A particular characteristic of this high relief setting is that there is a large degree of lateral flow in the geothermal systems, up to 10 km or more. In the andesitic stratovolcanic setting gold deposition occurs deep below the water table because the rising fluid is gas-rich (mainly CO_2 plus subordinate H_2S), and so boils at greater depth than gas-poor fluids (Henley *et al.,* 1984; Hedenquist & Henley, 1985). The most favourable sites for mineralisation will probably be within the upflow zone (White *et al.*, 1995), wherein deposition of ore almost always occurs more than 1 my subsequent to the formation of the host rocks (Heald *et al.*, 1987).

2.4 HIGH-SULPHIDATION STYLE EPITHERMAL PRECIOUS METAL DEPOSITS

High-sulphidation, acid sulphide or quartz-alunite systems constitute a class of epithermal deposit that is associated with advanced argillic alteration. The deposits are characterised by early copper (enargite)-gold mineralisation (Ashley, 1982; in Mitchell, 1992) with high contents of S, Se and Te and, sometimes, Hg. Although the main ore mineral is usually enargite-luzonite, sometimes with tetrahedrite-tennantite (and other LS-state sulphides), and commonly with gold, some deposits have only gold and negligible enargite, and some in the Eastern European arcs lack gold (Mitchell, 1992).

The characteristics of high-sulphidation (HS) epithermal ore deposits (Hedenquist *et al.*, 1994b; Arribas, 1995; in Hedenquist, 1995) indicate that the oxidised fluid deduced to be responsible for acidic leaching and alteration is similar in many respects to passively degassing volcanic systems and their associated acidic hot springs.

The principal ore minerals are generally enargite, tennantite-tetrahedrite, native gold, electrum, bournonite, dismuthinite, silver-bismuth sulphosalts, stibnite, and base-metal sulphides. Gangue minerals are quartz, pyrite, barite, siderite, realgar, and orpiment. HS gold deposits are characterised by an alteration-mineral assemblage dominated by hypogene alunite, commonly with associated enargite. These deposits usually formed shortly after the emplacement of their host rocks from acidic solutions closely related to magmatic systems (Heald *et al.*, 1987; in Ericksen, 1988), (Figure 29). Ore is hosted by fracture-controlled zones in which permeability was enhanced by early leaching, leaving only a silica-rich residue (vuggy silica), with alteration haloes of alunite, kaolinite, pyrophyllite, and diaspore (Hedenquist, 1995). Figure 33 presents a schematic cross-section of a HS-style deposit showing alteration, mineralogy and general location of ore zones. In most HS deposits the horizontal extent of ore exceeds the vertical. Enargite can also form massive sulphide bodies, as at Lepanto (Garcia & Bongolan, 1989; in Mitchell, 1992), where most ore occurs at the unconformable contact of the volcanics with basement (Gonzalez, 1956; Garcia & Bongolan, 1989; in Mitchell, 1992). At Chinkuashih the orebodies and the alteration lie within eruptive dacites and andesites at the top of the pre-volcanic basement, or within minor intrusions.

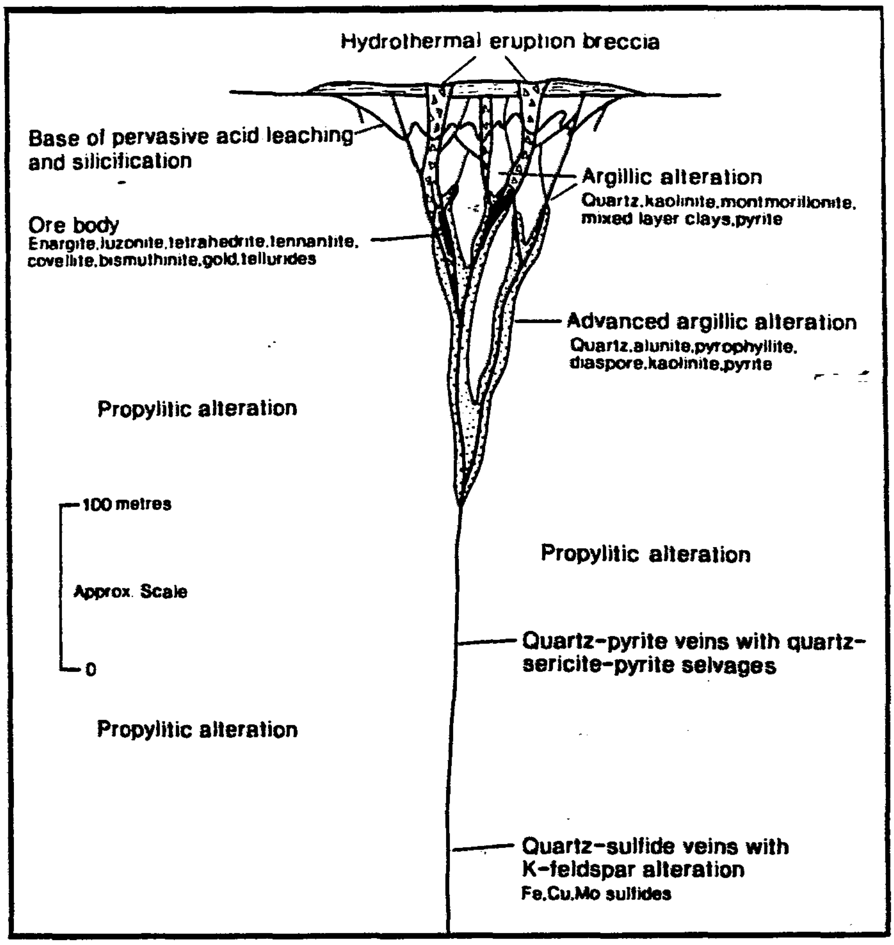


Figure 33. High-sulphidation style epithermal deposits. A Schematic cross-section of a high-sulphidation-style epithermal deposit showing alteration, mineralogy and general location of ore zones (after Henley, 1991).

Hydrothermal fluid temperatures, as measured in the Cordillera Occidental located in the Southern Peruvian Andes, range from 270° to 325°C. Salinities are 4 - 18 (Caudalosa) and 5 - 24 (Julcani) wt. percent NaCl equiv. Ericksen (1988), reported that the deposits are associated with:

a) Calderas: The resurgent caldera at Nevado Portuguesa (Rosario in Atunsulla);
b) Stratovolcanoes: Those at Ccarhuaraso and Palla Palla;
c) Volcanic domes: At Julcani and Castrovirreyna; and
d) Faults: (Cerro Anta in San Juan de Lucanas)

Examples of HS deposits in the United States include Goldfield, Nevada (Ashley, 1979; 1982; in Ericksen, 1988), Summitville, Colorado (Stoffregen, 1987), and Marisvale, Utah (Cunningham *et al.,* 1984; in Ericksen, 1988). In the Western Pacific HS deposits include Lepanto (Gonzalez, 1956; in Mitchell, 1992), (Figure 34) in the Philippines (production plus reserves of 900 000 t Cu and 122 t Au); Chinkuashih in Taiwan, with more than 92 t Au produced and a substantial potential (Li-Ping Tan, 1991; in Mitchell, 1992); and the relatively small Nansatsu deposits in Southwest Japan (Urashima *et al.*, 1987; in Mitchell, 1992). In Central and Eastern Europe productive HS deposits are largely confined to the Upper Cretaceous Srednogorie-Banat arc in Bulgaria and Serbia and to the Lahaca (Panto, 1951; in Mitchell, 1992) deposit in the Upper Eocene arc in Hungary.

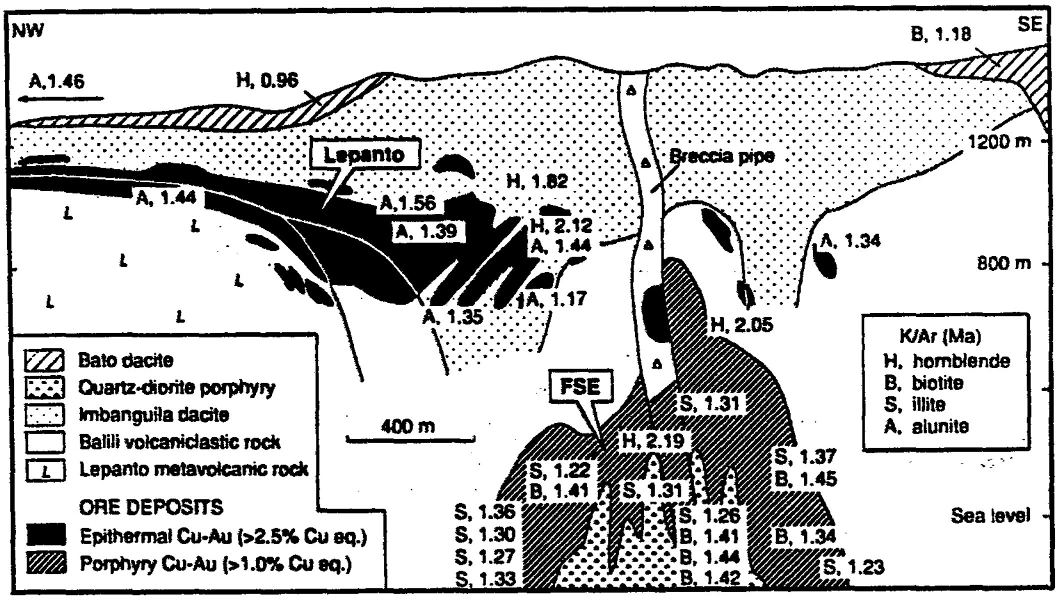


Figure 34. Schematic cross-section of Lepanto and Far Southeast (FSE) deposits along Lepanto fault (Garcia, 1991) showing major lithologic units, outline of enargite-Au and porhyry Cu-Au deposits, and most sample locations with K/Ar ages (after Arribas *et al.*, 1995).

A schematic section through an andesitic stratovolcano is presented in Figure 35, showing a stylised hydrologic regime and resultant hydrothermal mineralisation. In andesitic terrains like this, porphyry Cu deposits and Vuggy Silica type epithermal mineralisation may be common. In Vuggy Silica-Alunite-Kaolinite-Illite type mineralisation, silica commonly occurs as vuggy masses, apparently the product of rock destruction and replacement rather than open space filling. The vuggy silica forms irregular, pipe-shaped bodies or fracture-controlled discontinuous elongate lodes. Quartz-alunite alteration lies adjacent to lodes, within an envelope of quartz-illite-kaolinite alteration. Mineralisation is generally disseminated within the vuggy silica and massive silicification; gold is typically associated with a HS assemblage including covellite, luzonite-enargite, tennantite-tetrahedrite (Clarke, 1991).

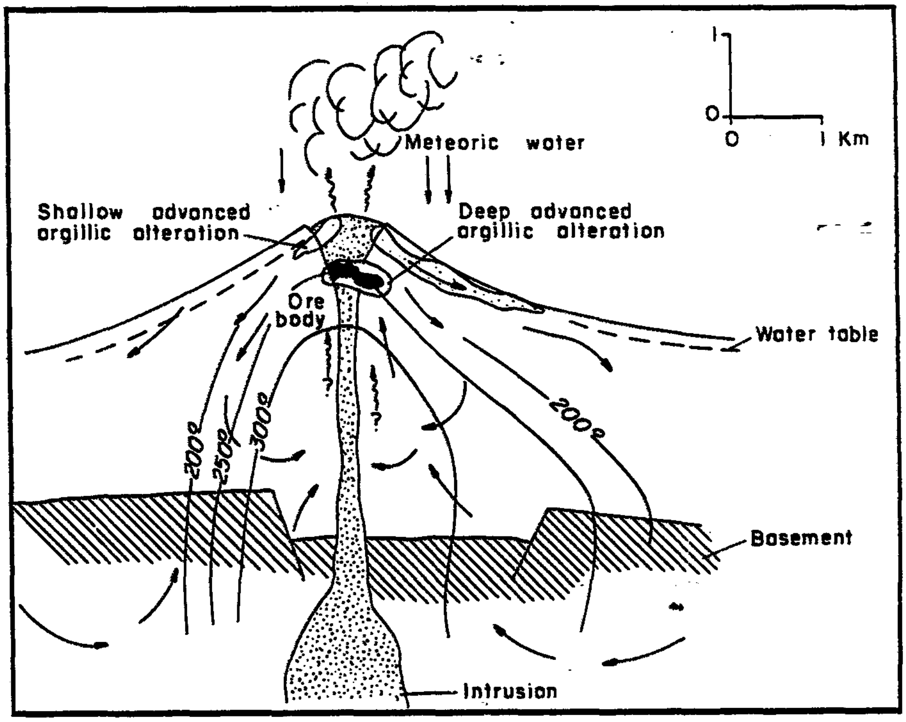


Figure 35. Occurrence model for high-sulphidation epithermal deposits envisioned by Heald *et al.* (1987), based partly on geothermal system of Henley & Ellis (1983). Note ore body lies well above level of hypogene boiling water table in Philippine geothermal fields. The wiggly arrows represent sulphur-rich emanations from the intrusion (after Mitchell & Leach, 1991).

Berger & Henley (1989; in Henley, 1991) have stressed some characteristics which suggest that two stages of alteration and mineralisation are involved in the HS-type deposit's formation. Stage 1 involves the intense acid alteration of host volcanics and stage 2 is the invasion of these alteration zones by a more-normal near-neutral pH fluid from which gold and associated minerals are precipitated by reaction with the alunite-kaolinite assemblage. Deen *et al.,* (1988; in Henley, 1991), using stable isotope data, have demonstrated this dynamic sequence in the Julcani district, Peru. The type of hydrodynamic environment for such mineralisation is illustrated in Figure 36 through analogy with the Hakone geothermal field in Japan.

The HS style of mineralisation shares many mineralogical and stable-isotope characteristics with the advanced argillic zone of alteration that caps porphyry Cu deposits (Hemley & Hunt, 1992; Rye, 1993; in Hedenquist & Lowenstern, 1994), and indeed there is commonly a close spatial relationship between these deposits (Sillitoe, 1989; in Hedenquist & Lowenstern, 1994).

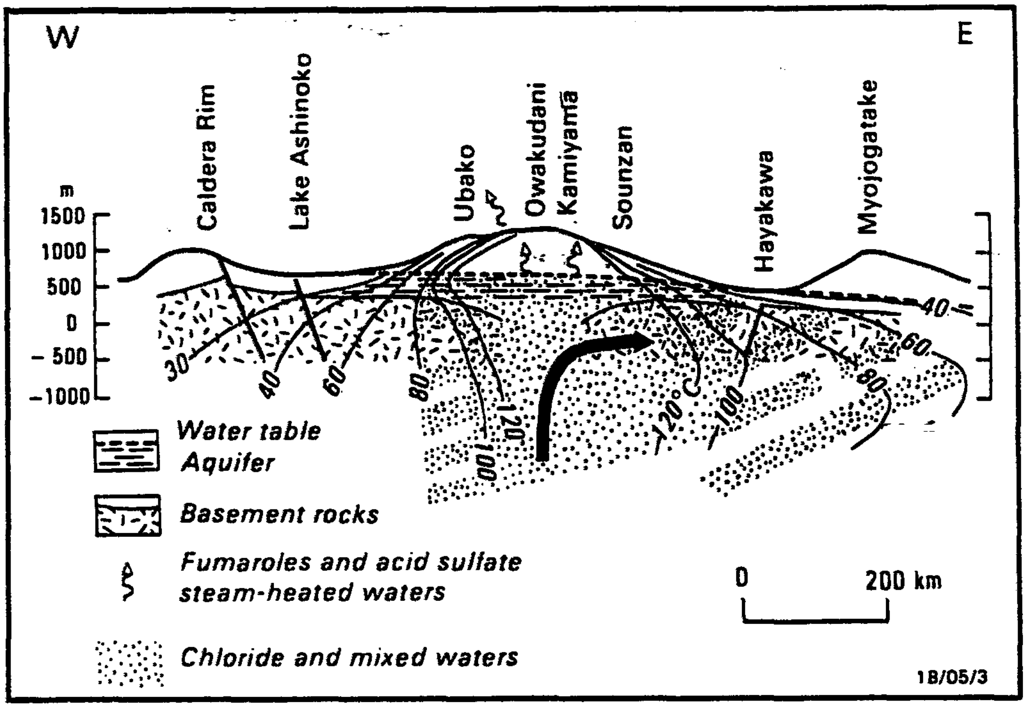


Figure 36. High-sulphidation-style epithermal deposits. Hydrology of the active hydrothermal system within the Hakone caldera and in relation to the occurrence of rhyolite domes. The regional hydraulic gradient controls groundwater-flow from West to East across the system. In such an environment, high-level magma degassing at an early stage may produce extensive alunite-kaolinite alteration which subsequently becomes the focus for ore deposition when invaded by the groundwater-based low-sulphidation system. Hakone is used here as an analogy of the hydrodynamic setting of some high-sulphidation systems without speculation on the possiblility of gold mineralisation in the system (after Henley, 1991).

2.4 PORPHYRY COPPER-GOLD ASSOCIATION

The much-disputed hydrothermal stacking hypothesis (Eidel & Meyer, 1987; in Mitchell & Balce, 1990) postulates that epithermal deposits form above underlying porphyry copper-gold deposits of similar age. There is commonly a close spatial relation between porphyry Cu (± Au) and HS epithermal Cu-Au deposits throughout the world, although a genetic association has not been proven (Arribas *et al.*, 1995). LS epithermal mineralisation also occurs in all districts with exposed porphyry copper ore bodies. This close spatial and temporal association of epithermal and porphyry deposits convinces many geologists that the two are genetically related, and formed above a common heat source at different levels (Mitchell & Leach, 1991).

Most proponents of stacking (Sillitoe, 1975; Sillitoe & Bonham, 1984; in Mitchell & Balce, 1990) argue that epithermal systems form at high levels in andesitic stratovolcanoes, while porphyry ore bodies develop 2 - 3 km beneath the base of the same volcanoes. HS epithermal gold deposits are generated typically in upper parts of porphyry copper-gold or copper-molybdenum systems, where they may be hosted by volcanic rocks roughly coeval with the mineralised porphyry stocks (Figure 28). Examples are displayed in Western Pacific Island

Arcs (*e.g.* Lepanto (Figure 34), Philippines: Garcia, 1991; in Sillitoe, 1993) and the Central Andes (*e.g.* El Hueso and La Pepa, Chile: Sillitoe, 1991; in Sillitoe, 1993). The LS-type most commonly occurs peripheral to, but also above or superimposed on, mineralised porphyry stocks. At Baguio, Marian, Paracale, Masara, Umuna, and Emperor, the gold-bearing veins or vein systems were localised by district-wide fault or fracture systems at roughly the same elevations as mineralised porphyry stocks, but up to 4 km from them (Figure 37) (Sillitoe, 1988).

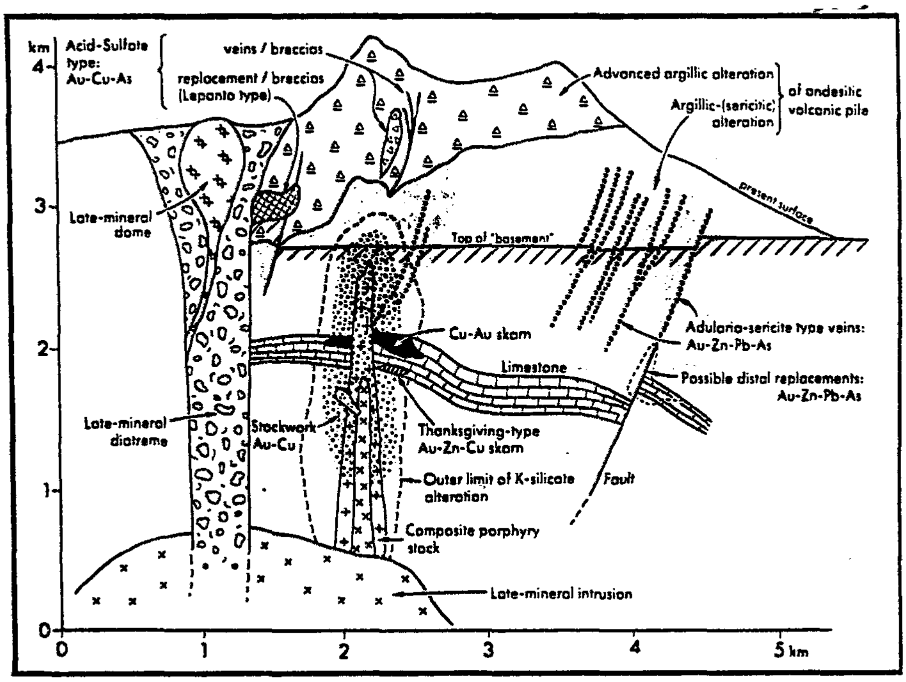


Figure 37. Styles of gold mineralisation in a typical Western Pacific porphyry copper system, with particular emphasis on epithermal deposits. A single system does not necessarily contain all mineralisation styles depicted (after Sillitoe, 1988).

Nowhere is the close spatial association between porphyry Cu (± Au) and HS epithermal Cu-Au deposits better seen than in Northern Luzon, Philippines, where the Lepanto epithermal Cu-Au deposit overlies the Far Southeast (FSE) porphyry Cu-Au deposit, both world-class orebodies (Figure 38). The metal associations in these deposits are consistent with a close genetic connection between porphyry and epithermal mineralisation since both deposits have similar Cu-Au ratios and high concentrations of Te, Bi, and Sn (Gonzalez, 1959; Calveria & Hedenquist, 1994; Arribas & Hedenquist, unpublished data; in Arribas *et al.*, 1995).

The establishment of this potential genetic link also has consequences for the source of metals in the epithermal environment. In porphyry Cu and intrusion-related Au deposits there is wide agreement on a dominantly magmatic fluid source for the ore-forming metals (Sillitoe, 1991; in

Arribas *et al.*, 1995), whereas in the epithermal environment, the evidence for a magmatic source of the metals weakens (Hedenquist & Lowenstern, 1994). Since there exists a metal association between the world class Lepanto enargite-Au HS deposit and the FSE Cu-Au porphyry deposits, and since the Cu, Au, Te, Sn, and other elements in the FSE porphyry deposit were derived from fluids released by diorite intrusions, it is most likely that the same metal suite in the Lepanto deposit was also derived from the said intrusions, either by a magmatic fluid or by remobilization of porphyry protore in an evolving hydrothermal system. Thus, the Lepanto epithermal deposit has good evidence for a magmatic source of Cu and Au. Further, K/Ar ages of alunite from Lepanto have the same range as those of hydrothermal biotite and illite from the FSE deposit, confirming that both epithermal and porphyry mineralisation formed from an evolving magmatic-hydrothermal system that was active for about 300 ka (Arribas *et al.*, 1995).

Additionally, a number of characteristics in the Wahmonie area of the Southwest Nevada Volcanic Field (Figure 39), suggest that exposed and near-surface mineralisation may be associated with an underlying porphyry system. Firstly, the high bismuth and tellurion concentrations of the Wahmonie district are not typical of LS-type precious-metal deposits (*e.g.* Creede, Colorado, Heald *et al.,* 1987); these elements are more commonly associated with porphyry-related gold deposits such as the Top deposit at Bald Mountain and the Fortitude and McCoy deposits in Nevada (Bonham, 1989; Brooks *et al.,* 1991; in Castor & Weiss, 1992). Secondly, subvolcanic stocks and rhyolitic dykes exposed in a central horst as well as geophysical evidence for a pluton beneath the district are consistent with the presence of a buried, perhaps composite, porphyry intrusion. Lastly, the presence of hypersaline fluid inclusions in quartz phenocryst and biotite ± tourmaline veins within the porphyritic granodiorite argue strongly for at least some porphyry-type magmatic-hydrothermal activity.

It remains an open question whether epithermal deposits are typically located above or peripheral to copper-gold porphyry deposits, whether they might be the near-surface, low-temperature equivalents of copper-gold porphyry systems, or whether they represent shallow deposits from a partially or totally reworked copper-gold porphyry system at depth (Heald *et al.*, 1987).

Although suggested (Sillitoe, 1989; in Arribas *et al.*, 1995), a genetic connection has not yet been proven, in part because no chronological data exist to support unambiguously a direct

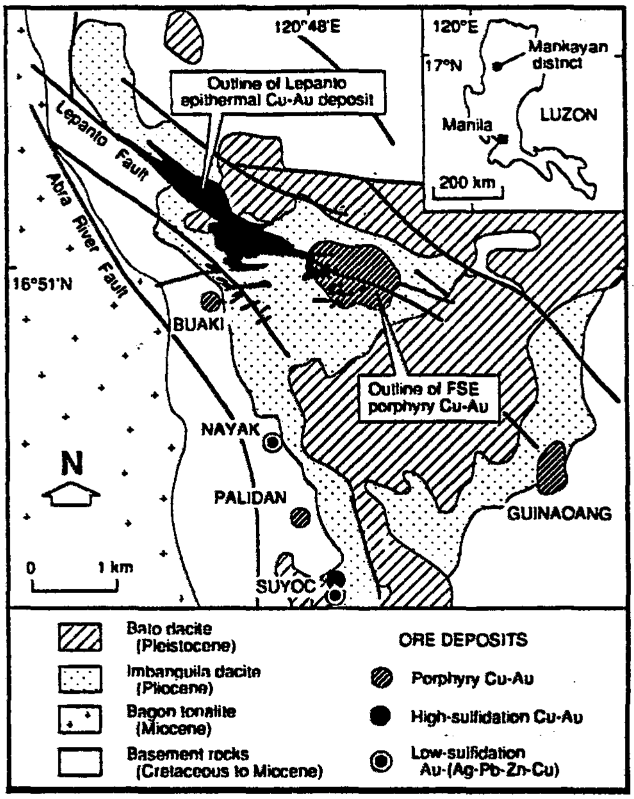


Figure 38. Map of Mankayan district in Northern Luzon (inset), Philippines, showing simplified geology and location and type of known hydrothermal deposits. Outlines of economically most important deposits (*i.e.*, FSE [Far Southeast], Guinaoang, and Lepanto) are shown projected to surface (based on Garcia, 1991; in Arribas *et al.*, 1995).

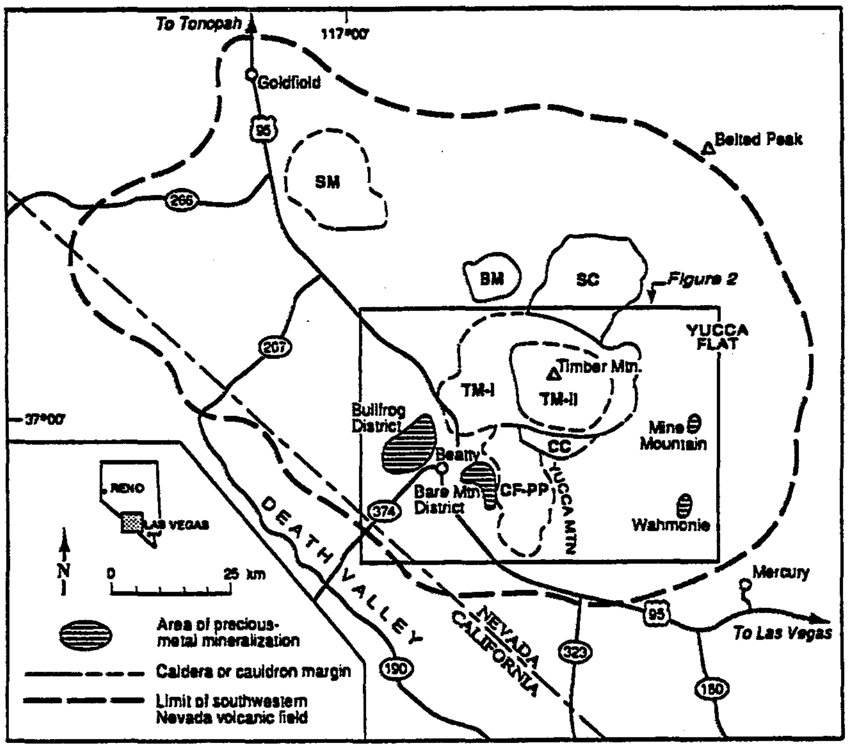


Figure 39. Map of the Southwestern Nevada Volcanic Field (SWNVF) showing major volcanic centers and mineralised areas (modified from Noble *et al.*, 1991; and Byers *et al.*, 1989). BM=Black Mountain caldera, CF-PP=Inferred Crater Flat-Prospector Pass caldera complex, CC=Claim Canyon cauldron, SC=Silent Canyon caldera, SM=Stonewall Mountain volcanic center, TM-1=Timber Mountain caldera complex, I,TM-II=Timber Mountain caldera complex II. Heavy dashed line is approximate limit of SWNVF (after Castor & Weiss, 1992).

temporal relation between porphyry and epithermal mineralisation. If a genetic connection can be demonstrated, this will have implications for the interpretation of ore-forming processes related to high-level intrusions (Arribas *et al.*, 1995). However, there is as yet no requirement that a porphyry Cu deposit is overlain by a HS deposit, or that a LS deposit will be rooted by porphyry mineralisation adjacent to the intrusive heat source (Figure 29) (Hedenquist & Lowenstern, 1994).

2.5 REGIONAL GEOTECTONIC SETTINGS OF EPITHERMAL DEPOSITS

Epithermal deposits are found in a variety of geological environments which reflect various combinations of igneous, tectonic, and structural settings (White & Hedenquist, 1990). These deposits occur principally in areas of significant magmatic and volcanic activity where localised and enhanced geothermal gradients related to magmatic activity develop (Swindell, 1986).

Both HS-type and LS-type epithermal deposits occur in tectonic settings associated on a continental scale with plate boundary subduction zones (Silberman *et al.,* 1976; Sillitoe, 1977; Mitchell & Garson, 1981; Clarke *et al.,* 1982; in Heald *et al.*, 1987), (Table 10), such as the West Pacific island arcs (Figure 40), the Western United States Cordilleras, and the Antofagasta magmatic arc in Northern Chile (Figure 41). The ores occur in structurally complex environments, typically with several generations of faults or fractures developed in two or more directions (Heald *et al.*, 1987).

Table 10: Some representative hydrothermal ore deposits associated with subduction-related magmatism (after Hedenquist & Lowenstern, 1994).

Ore deposit type	Relation to magma	Temperature Depth	Fluid	Associated metals	Example of active analogue
Porphyry	Adjacent to or hosted by intrusion	>600 to 300°C 2-5km	Hypersaline and immiscible vapour	Cu ± Mo ± Au, Mo, W or Sn	Shallow magma bodies beneath stratovolcano
Pluton-related veins	Fractures in and near intrusion	300 - 450°C Variable	Moderate to low salinity	Sn, W, Mo ± Pb-Zn, Cu, Au	Shallow magma bodies beneath stratovolcano
Epithermal (high sulphidation)	Above parent intrusion	<300°C Near surface to >1.5km	Moderate to low salinity, early acidic condensate	Au-Cu Ag-Pb	High-temp. fumaroles and acidic springs near volcanic vent

Table 10: Continued

Ore deposit type	Relation to magma	Temperature Depth	Fluid	Associated metals	Example of active analogue
Epithermal (low sulphidation)	Distant (?) from magmatic heat source	150 - 300°C Near surface to 1 - 2 km	Very low salinity, gas-rich, neutral pH	Au(Ag, Pb-Zn)	Geothermal systems with neutral-pH hot springs, mud pools
	Distant (?) from magmatic heat source	150 - 300°C Near surface to 1 - 2km	Moderate salinity	Ag-Pb-Zn(Au)	Not observed, transient brine?

The term 'fluid' is used to refer to non-silicate, aqueous liquid and/or vapour. The salinities (Na, K chloride) of fluids in these environments vary from hypersaline (>50 wt%) to moderate (10-20 wt%), low (<5 wt%) and very low (0.2-0.5 wt%) salinity.

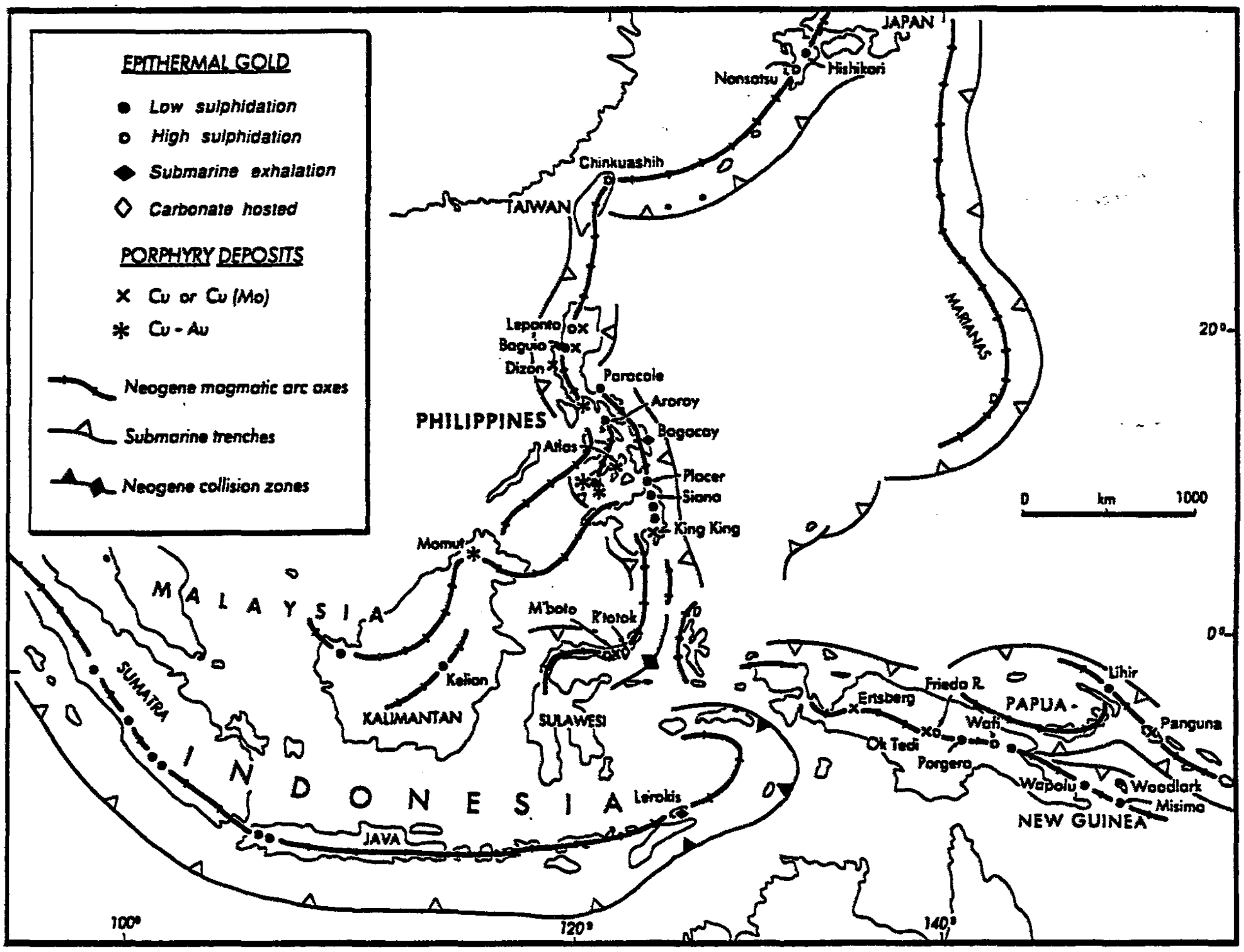


Figure 40. Neogene magmatic axes, subduction zones and related epithermal and porphyry Cu and Au deposits in the Western Pacific and Southeast Asia, excluding Myanmar and Fiji. Deposits shown are significant producers or else prospects or deposits under development (after Mitchell, 1992).

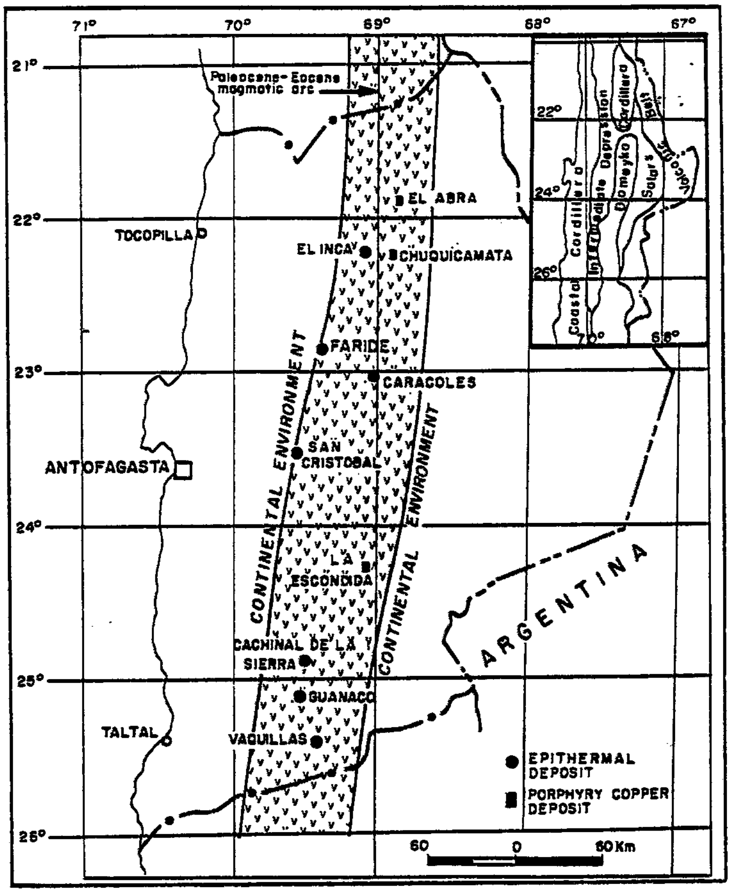


Figure 41. Palaeogeographic sketch of the Palaeocene-Eocene magmatic arc in the Antofagasta Province, Northern Chile. Also shown are the principal epithermal precious metal deposits, and as a reference, the major porphyry copper deposits (modified from Maksaev, 1984; in Camus & Skewes, 1991).

The Philippine Archipelago (Figure 42), with numerous epithermal gold deposits and prospects, comprises largely of arc systems built on basement composed mainly of ophiolite and metamorphic rocks. Deposits are mostly concentrated along the axes of late Cenozoic volcanic arcs (Mitchell & Balce, 1990). The large number of prospects and deposits, and the high gold potential of the Philippine arcs, typically reflects the presence of stratigraphic successions, basement anticlines and fractures favourable for focusing fluid flow at shallow depths following subaerial, predominantly andesitic, volcanism. These favourable rock sequences and structures can best be explained by Late Cretaceous to Early Palaeocene emplacement of ophiolite followed by arc reversal. This resulted in neutral or weakly extensional regional horizontal stress directed perpendicularly to the arcs - a situation favourable for epithermal mineralisation. Almost all subduction in the Philippines has associated active volcanoes, except for the Westward subduction in the East Luzon Trough (Mitchell & Balce, 1990).

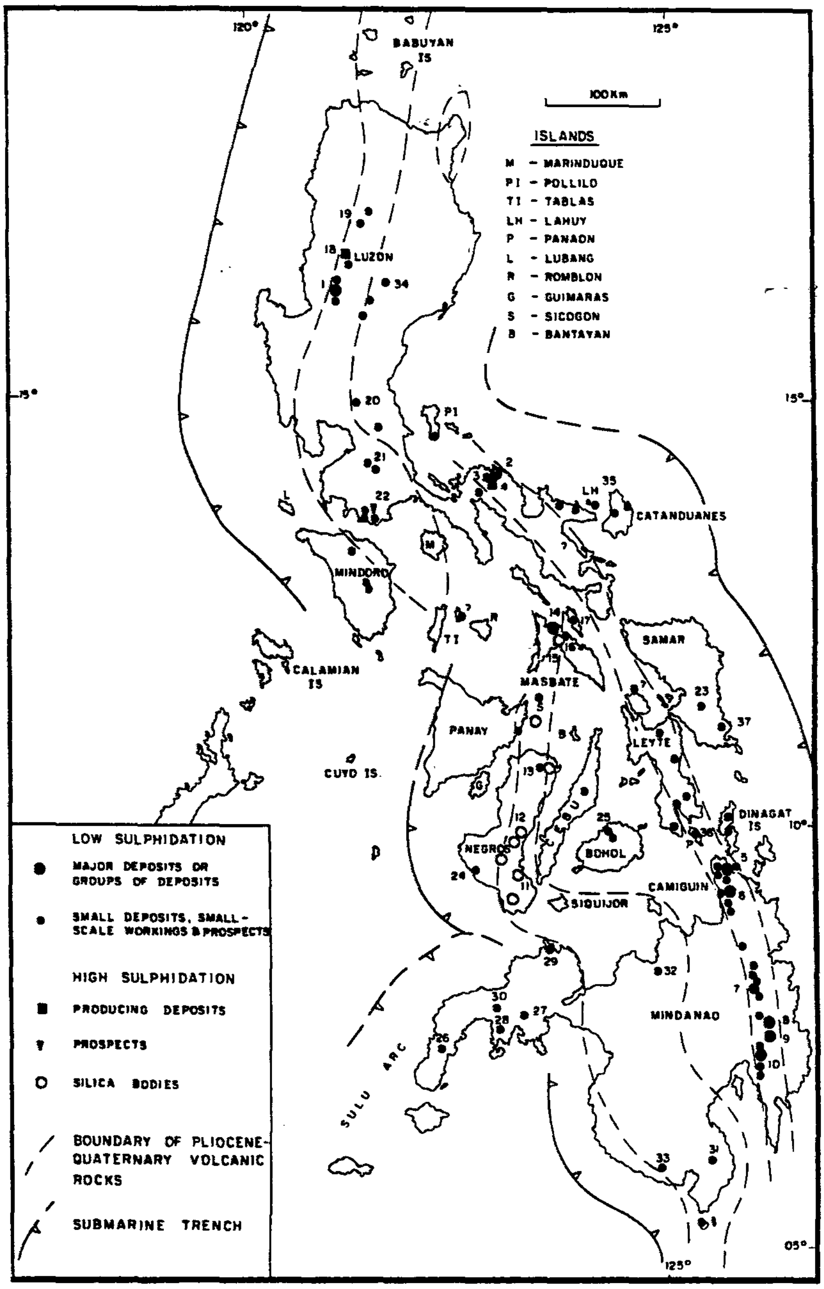


Figure 42. Distribution of epithermal gold mines and prospects and silica bodies in the Phillipines (after Mitchell & Leach, 1991).

The distribution and genesis of epithermal Au-Ag deposits in the Cordilleras of the Western United States is intimately tied to the plate-tectonic history of the region and those processes accompanying arc volcanism and plutonism. Of particular importance during intervals of back-arc volcanism have been deep-seated extensional fault zones and rifts, some reactivated from earlier periods of tectonism, that have provided access to deep sources of magmas, gases and metals (Berger & Henley, 1989; in Berger & Bonham, 1990).

From the above it will be appreciated that recent magmatic activity in continental-margins or island arcs is recognised as one of the principal geotectonic settings for the generation of metallic epithermal ore deposits (Sillitoe, 1977).

The Southwest Pacific is one of the most complex tectonically active regions of the Earth's crust (Howell *et al.,* 1985; Carlile & Mitchell, 1994; in White *et al.*, 1995), (Figure 3). The most common tectonic setting in the region results from subduction of an oceanic plate beneath another oceanic plate. Subduction of oceanic plates beneath continental plates is also occurring in several parts of the region (Barber, 1985; Stauffer, 1985; in White *et al.*, 1995). Thus the complex geology of the Philippines results from convergent subduction occurring on both sides of the archipelago, producing a zone comprised principally of obducted slices of ophiolites through which island arc volcanism has occurred (Gervasio, 1968; McCabe *et al.,* 1985; Hawkins *et al.,* 1985; in White *et al.*, 1995).

Most active island arcs in the Western Pacific are separated from either a continent or another island arc by back-arc oceanic basins that were generated by ocean-floor spreading in the Cenozoic. Pre-Cenozoic epithermal deposits are likely to be preserved in arcs that are located on the overriding plate during terminal continent-continent or continent-island arc collision (Mitchell, 1992). The geological features of the destructive plate margins of the Southwest Pacific also favour the formation of porphyry type deposits (Bogie, 1995). In these porphyry systems, HS-type gold deposits are commonly located above the mineralised stocks whereas LS-type deposits tend to occur distally around the stocks (Sillitoe, 1988).

Igneous activity has an important role in the formation of most epithermal deposits, if only in providing the heat necessary to generate a hydrothermal convection cell. The magmas may also contribute at least a component of the total gases to an overlying hydrothermal system (Giggenbach, 1986; in White & Hedenquist, 1990), and their possible contribution of metals has been speculated on (Hedenquist, 1987; Berger & Henley, 1989; in White & Hedenquist, 1990). Modern volcanic environments in which hydrothermal activity is occurring vary widely, and have been classified into silicic depressions (commonly calderas or grabens), andesitic stratovolcanoes, cordilleran volcanism, and oceanic volcanic islands (Bogie & Lawless, 1987; White *et al.,* in press; in White & Hedenquist, 1990), (Figure 43).

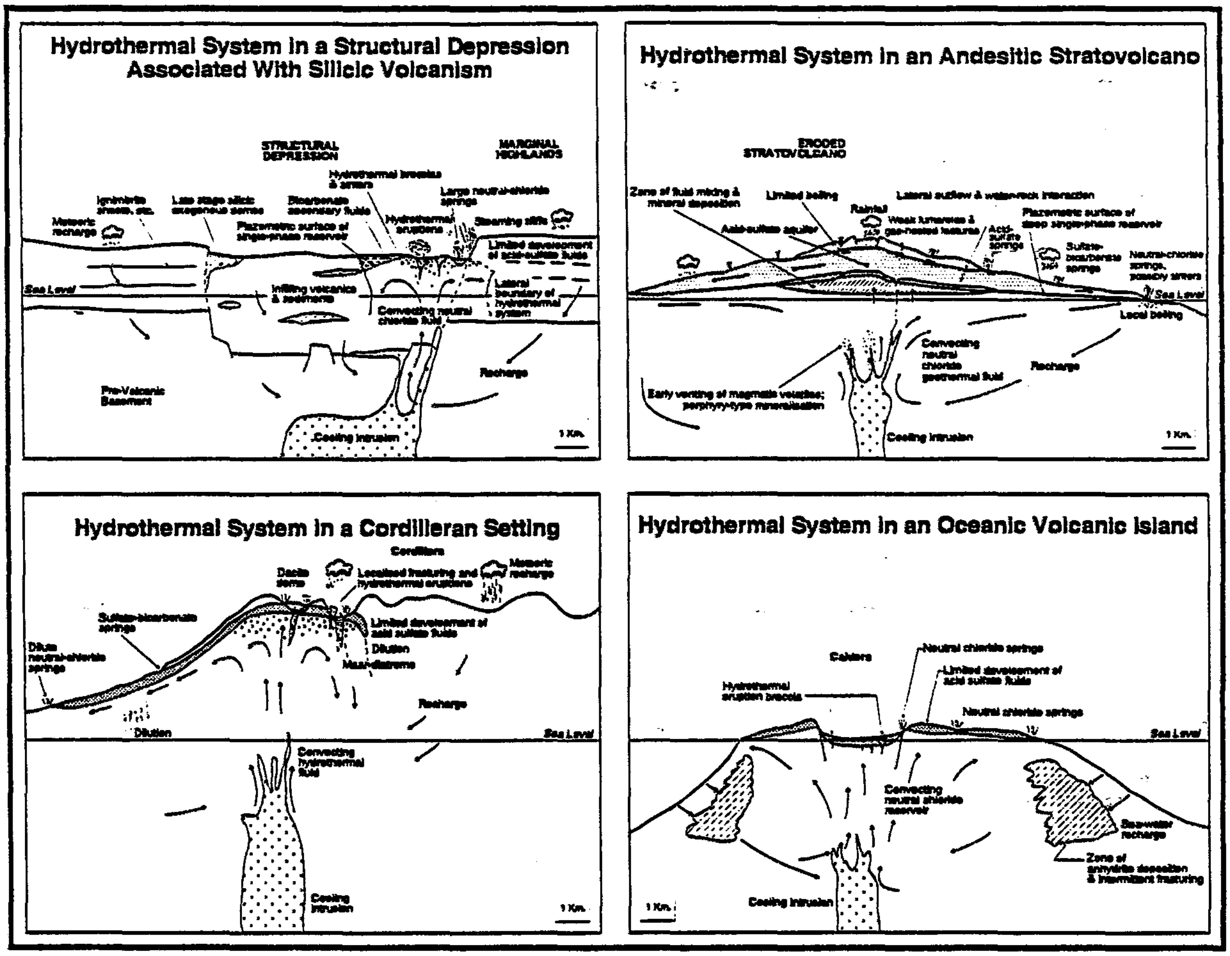


Figure 43. Epithermal settings (re-drawn after Bogie & Lawless, 1987; in White *et al.*, 1995).

Stratovolcanoes are the dominant volcanic landform in andesitic terrains on plate margins. These form volcanic islands or large subaerial edifices of high relief, and may eventually build up cordilleran mountain belts. In rhyolitic terrains, volcanism is marked by pyroclastic eruptions resulting in silicic depressions, either as downwarp basins or collapse calderas. Rhyolitic to dacitic flow dome structures are associated with resurgence within, or marginal to, these vent depressions (Clarke, 1991). Each of the above is characterised by a different hydrological regime which controls the discharge and recharge of the hydrothermal system, the distribution of conduits, the types and distribution of hydrothermal alteration products, and the potential sites of deposition of ore minerals. The characteristics of these different epithermal settings are reproduced in Table 11.

Table 11: Characteristics of different epithermal settins. Modified from Bogie and Lawless (1987; in White *et al.,* 1995). LS=low sulphidation, HS=high sulphidation

Characteristic	Silic depression	Andesitic stratovolcano	Cordilleran volcanism	Oceanic volcanic island
Relief	Low (0-300 m)	High (500-2000m)	High (500-3000m)	Moderate (200-500m)
Lithologies	Acid lavas surrounded by pyroclastic deposits and sediments. Local andesitic centres.	Andesitic lava flows and interbedded breccias, commonly with steep depositional dips. Local dacite domes.	Local andesitic centres and dacite domes on deformed basement.	Basaltic to andesitic lavas and fragmental rocks. Local limestone.
Calderas	Common, large.	Common, small.	Uncommon.	Common, small.
Intrusions	Not common, small, acidic.	Very common, mostly small. Diorite, some granodiorite.	Common, may be large. Diorite to granodiorite.	Common, small. Gabbro and diorite.
Relation of upflow to eruptive centres.	Not closely related.	Closely related.	Not closely related.	Closely related.
Surface expression	Neutral pH hot springs with sinters in depressions, hydrothermal eruption craters.	Fumaroles and solfataras at high relief, acid springs on flanks, neutral springs at distance. Sinters rare.	Hot springs with sinters in depressions. Acid springs on flanks, neutral springs at distance. Hydrothermal eruption craters.	May have hot springs with sinters in calderas. Hydrothermal eruption craters.
Phase separation	In upflow, limited alteration from separated gases.	In upflow and lateral flow, very extensive alteration from separated gases.	In upflow, minor alteration from separated gases.	In upflow, minor alteration from separated gases.
Hydrothermal alteration	Mostly illite-smectite, minor kaolinite-alunite.	Widespread propylitic below, very extensive illite-smectite and kaolinite-alunite.	Mostly illite-smectite, minor kaolinite-alunite.	Mostly illite-smectite, minor kaolinite-alunite. Sea-water recharge may cause anhydrite deposition.
Deposit styles	Low-sulfidation deposits.	Low- and high-sulphidation deposits, porphyry copper deposits.	High- and low-sulphidation deposits, porphyry copper deposits.	Low-sulphidation deposits, although a high-sulphidation to porphyry copper association is recognised.

Table 11: Continued

Characteristic	Silic depression	Andesitic stratovolcano	Cordilleran volcanism	Oceanic volcanic island
Gold deposition	In stockworks, discrete veins and breccias, especially at depth.	Dominantly in structurally controlled veins (LS), or permeable lithologies (HS).	In stockworks, discrete veins and breccias, especially at depth (LS), or permeable lithologies (HS)	In near-surface zones of high permeability
Localising controls	Permeable lithologies, faults, fractures, caldera margins.	Faults and fracture zones in competent lithologies (LS), or minor structures in crater, dome or maar settings (HS).	Faults, shears, fracture zones, mostly in basement rocks (LS), or in minor structures in crater, dome or maar settings (HS).	Permeable lithologies, faults, fractures, caldera margins.
Example of setting	Central Taupo Volcanic Zone, New Zealand (Cole, 1987)	Mt. Ruapehu, New Zealand (Hackett and Houghton, 1989)	Amacan, Philippines (Barnet *et al.*, 1985)	Curtis Island, Kermadec Group (Doyle *et al.*, 1979)
Deposits	Ohakuri Dam, New Zealand	Woodlark Island, Papua New Guinea.	Acupan, Philippines. Lepanto, Philippines.	Laddam, Lihir Island, Papua New Guinea.

Strong, regional, volcanic-related structural control is almost universally recognised for epithermal deposits (Henley, 1990; in White & Hedenquist, 1990), since these settings cause an increase in the presence of multiple fracturing, subsequently enhancing permeability within such a zone in the near-surface. Reactivated fault systems are particularly prone to developing the complex fracture arrays, extreme local dilation and dramatic increases in permeability that are necessary to efficiently focus hydrothermal fluid and produce large mineral deposits. At any given period in the evolution of an arc system, it may be dominated by either extensional, compressional or strike-slip deformation. Indeed, it is common for the tectonic environment to change repeatedly through the development of a single arc segment. As a result, structures formed in one tectonic environment are commonly complexly reactivated in quite different environments later in arc evolution. These complexly reactivated structures have greater potential to localise extreme fracturing, dilation and fluid flux than primary structures (Henley & Etheridge, 1995).

Deeply rooted rifting is also a key element in the genesis of epithermal systems within any volcanic terrain. In many volcanic fields, as volcanism progresses, a graben develops and the locus of volcanism changes to the depression. In such a progression, it is at this graben-forming stage that the ore-related epithermal systems begin to develop along the graben margin (Figure 25), possibly because of the increased access of meteoric waters to the magmatic heat sources and increased permeability that brings larger quantities of magmatic or deeply derived components to shallow crustal levels (Berger & Bonham, 1990). Thus deep-seated, sub-vertical basement faults are particularly important for localising high-level intrusives and volcanoes, together with their associated mineralisation (Henley & Etheridge, 1995).

Sibson (1987; 1992; in Henley & Etheridge, 1995) drew attention to the focusing power of step-overs, jogs and bends in strike-slip fault systems (Figure 26), taking as a possible example the Martha Hill, Waihi gold deposit (New Zealand). Strike-slip faults are a feature of oblique convergent margins. A well mineralised example of such a strike-slip fault system within an arc above an obliquely convergent margin is the Philippine Fault (Henley & Etheridge, 1995). Other examples include Hishikari (Japan), Camp Bird (Colorado) and Pajingo (Queensland) (Porter, 1988; in Henley, 1991).

A close association with felsic calderas and andesitic vent complexes has been observed in the San Juan Mountains of Colorado (Steven *et al.,* 1977; in White & Hedenquist, 1990), and in some parts of Japan (Kubota, 1986; in White & Hedenquist, 1990) and the Southwest Pacific. In fact, the most common regional structural setting for LS-type deposits is along the margins of calderas (Figure 27), although other tectonic settings (typically structurally complex volcanic environments) are not uncommon (Hayba *et al.*, 1985). The importance of the caldera setting lies in the excellent plumbing system it can provide for younger hydrothermal systems. Lipman *et al.,* (1976; in Heald *et al.*, 1987) concluded that the role of calderas with respect to mineralisation in the San Juan Mountains, Colorado, was primarily one of formation of zones of weakness above a magma chamber and that the caldera structures channelled distinctively younger hydrothermal fluids.

Other settings less frequently hosting epithermal gold deposits include those at Ladolam (Davies & Ballantyn, 1987; in Sillitoe, 1988), Kabang (Licence *et al.,* 1987), and Emperor (Anderson *et al.,* 1987), which are associated with domes or plugs of alkalic (trachytic) composition that were emplaced in summit calderas following collapse. Finally, with perhaps the best example of the dynamic, shallow environment represented by the term "maar-diatreme" (Figure 44), is the complex breccia-dominated setting at Ladolam, Lihir Island (Carmen, 1994; in Henley & Etheridge, 1995) as it evolved from a porphyry copper-gold style to epithermal gold style habitat.

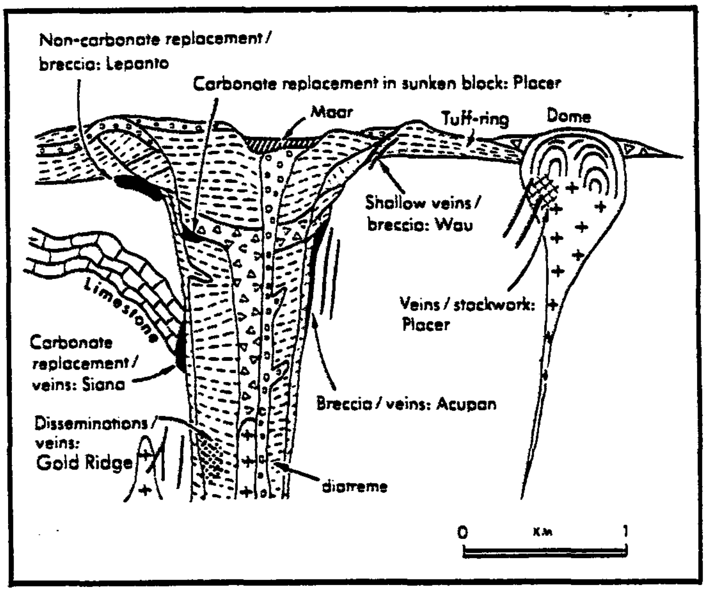


Figure 44. Idealised styles of gold mineralisation associated with maar-diatreme systems in Western Pacific island arcs. The presence of more than one style at a single locality is not neccessarily implied (after Sillitoe, 1988).

2.6 HOST LITHOLOGIES OF EPITHERMAL DEPOSITS

The host rocks for epithermal districts are rhyolitic to andesitic in composition (Heald *et al.*, 1987). Calc-alkaline andesite or dacite arc sequences, here referred to as andesitic, are characteristic of both continental margin arcs and island arcs. All known andesitic arcs are mineralised (Mitchell, 1992). Andesitic successions in epithermal districts are usually non-marine, although mineralised shallow marine andesitic clastics are locally present, while submarine mass-flow deposits accumulate away from the arc axis. Arc volcanism is either basaltic, lacking porphyry and epithermal deposits, or predominantly calc-alkaline andesitic to dacitic, forming porphyry- and epithermal deposit-hosting stratovolcanoes that rest unconformably on basement (Mitchell, 1992), (Figure 24). In a few districts, associated sediments (Creede, Guanajuato) and intrusive rocks (Silver City) are also mineralised, but mineralisation is mainly confined to the volcanics (Heald *et al.*, 1987).

2.6.1 HIGH SULPHIDATION DEPOSITS

The primary host rock for HS-type epithermal deposits is almost exclusively rhyodacitic and is also commonly porphyritic (*e.g.* Goldfield, Summitville, Red Mountain)(Heald *et* *al.*, 1987). For both HS- and LS-type precious metal deposits of the Southwestern Cordillera, the host rocks are largely Tertiary calc-alkaline extrusions with hypabyssal intrusions. Here andesites are the more common hosts to ore shoots. These volcanic piles commonly contain andesitic agglomerates, dykes, breccias and flows; rhyolitic tuffs, dykes and small plugs; latitic and rarely dacitic flows and breccias; and lake bed and fluvial volcanogenic sandstones and shales (Buchanan, 1981).

Although there is consistency in the host rock compositions of HS epithermal deposits, significant lithological differences do occur between the various individual precious metal deposits. Examples depicting these differences include:

a) The Rodalquilar HS-type deposit which is located in the area between Almeria and Cartagena in Southeast Spain. This is a region of widespread Neogene magmatic activity ranging from calc-alkaline andesites via ultrapotassic rocks toward alkali basalts (Lopez-Ruiz & Badiola 1980; Bellon *et al.*, 1983; Venturelli *et al.*, 1984; Bordet 1985; Torres-Roldan *et al.*, 1986; in Sänger-von Oepen *et al.*, 1989);

b) The Mankayan district which hosts the well documented Lepanto and FSE deposits in Northern Luzon, Philippines. The geology here can be divided into four main lithologic groups (Figure 38): (i) a volcanic to epiclastic basement consisting of several units (*i.e.* Lepanto metavolcanic rock, Apaoan sedimentary rock, and Balili volcaniclastic rock) of Late Cretaceous to Middle Miocene age; (ii) a large Miocene intrusion (Bagon intrusion) of tonalitic to gabbroic composition; (iii) a Pliocene dacitic pyroclastic and porphyry unit (Imbanguila hornblende dacite) that predates Cu-Au mineralisation in the Lepanto and FSE deposits; and (iv) an unaltered Pleistocene dacitic pyroclastic and porphyry unit (Bato hornblende-biotite dacite) (Arribas *et al.*, 1995);

c) The Kuril volcanic belt, which hosts numerous deposits/prospects (*e.g.* Prasolovskoye, Kunashir Island). The belt is composed primarily of calc-alkaline volcanic rocks of Miocene to Quaternary age (So *et al.*, 1995);

d) Lesbos which is part of a belt of late Oligocene-middle Miocene calc-alkaline to shoshonitic volcanism in the Northern and Central Aegean Sea and Western Anatolia (Fytikas *et al.*, 1984; in Kontis *et al.*, 1994). Here Au-rich quartz veins in the Megala Therma area (near Arghenos village) are hosted in andesitic volcanic rocks of Upper Miocene age (Figure 45); and

e) The McDermitt caldera complex located 50 km to the North of Winnemucca, Nevada, where associated rhyolitic ash-flow tuffs host the Sleeper deposit (Saunders *et al.*, 1988).

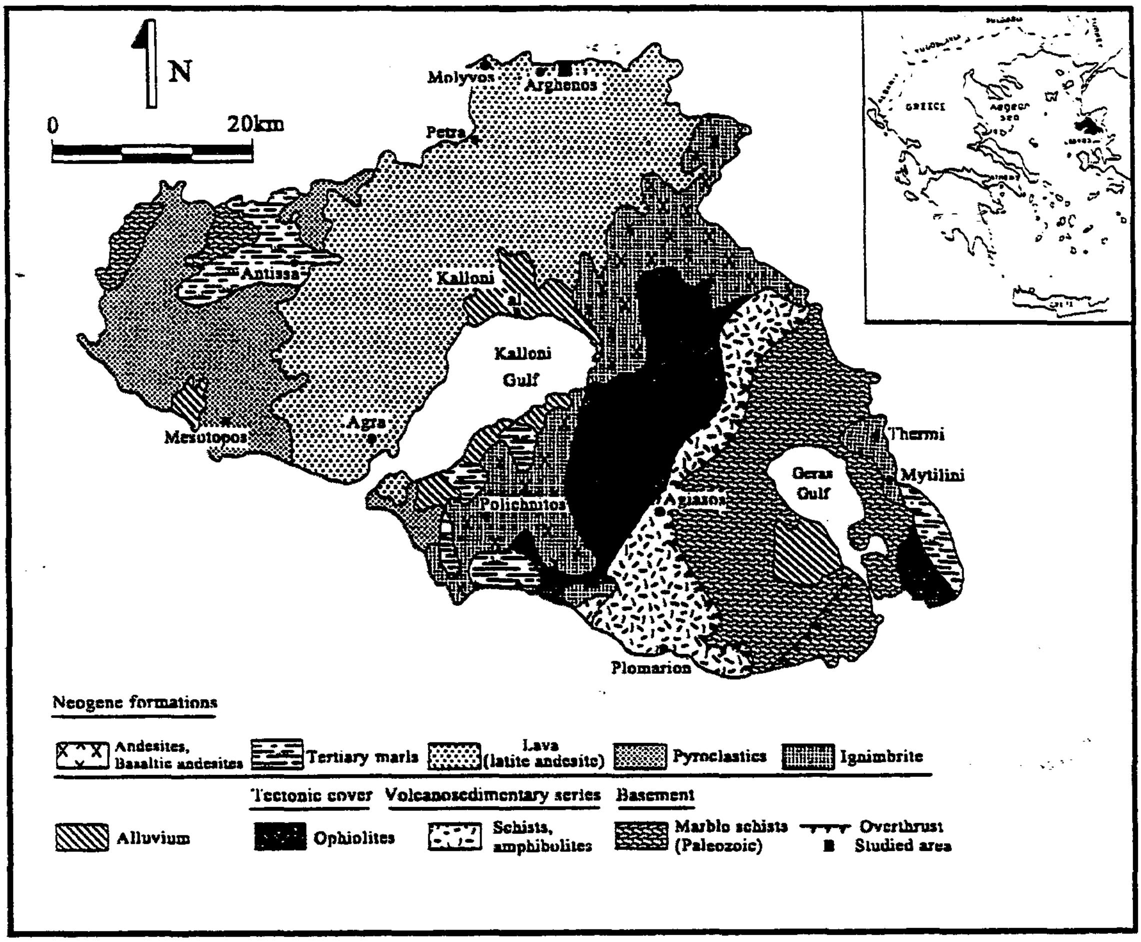


Figure 45. Geology of Lesbos (after Kontis *et al.*, 1994).

2.6.2 LOW SULPHIDATION DEPOSITS

Ores from LS-type deposits occur in host rocks of varying composition (generally rhyolitic to andesitic), and in several different compositional units within a district (Heald *et al.*, 1987). Ore is nevertheless mainly confined to volcanic rocks (Hayba *et al.*, 1985). For LS-type deposits, the more random occurrence of ore in several lithologies implies that the composition of the host rock(s) is not a controlling factor (Heald *et al.*, 1987).

Most LS-type deposits in both the Western Pacific and Central Europe are hosted by eruptive or volcaniclastic rocks that lie not more than a few hundred metres above underlying basement. In the gold districts of the Philippines, the stratigraphic succession generally comprises gently dipping mineralised andesitic or dacitic predominantly clastic rocks up to 500m thick, lying unconformably on folded "basement". Some ore-bearing host rocks in these districts include younger dykes, sills and small stocks of andesite, andesite porphyry, or dacite porphyry (Mitchell & Leach, 1991). At other deposits, ore-grade veins penetrate the basement and at Hishikari (Izawa *et al.*, 1990) they are concentrated at or just below the basement-volcanic unconformity (Figure 46). In a regional sense therefore, the deposits at Hishikari are unconformity-controlled (Mitchell, 1987; in Mitchell, 1992).

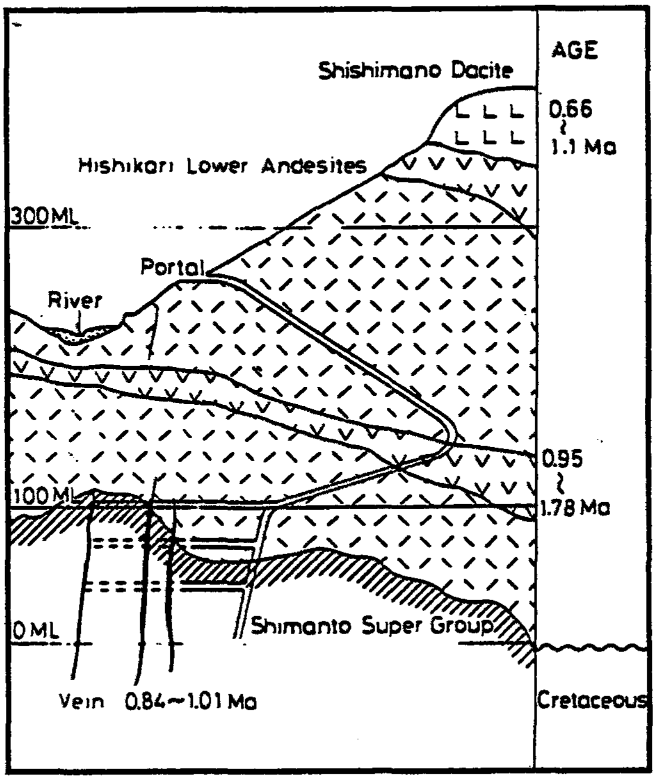


Figure 46. Schematic section through Hishikari low-sulphidation gold deposit, Southwest Japan, showing relationship of mineralisation to unconformity. Cross-hatching represents Lower Andesites (from Izawa *et al.*, in Mitchell, 1992)

Although mineralised volcanic rocks comprise mostly calc-alkaline andesites or dacites, there are four deposits in the Western Pacific, including Porgera (Figure 47) in Papua New Guinea (Richards, 1990; in Mitchell, 1992) and the Emperor mine in Fiji, where the mineralised rocks are porphyritic mafic alkaline intrusives or eruptives.

Examples illustrating the variety in compositional units hosting LS epithermal deposits include:

a) Java and Sumatra (Figure 48(a)) which form part of the Western Sunda-Banda Continental Margin (Charlie & Mitchell, 1994; in Marcoux & Milési, 1994), which developed along the Northern margin of the subducting Indian-Australian plate during

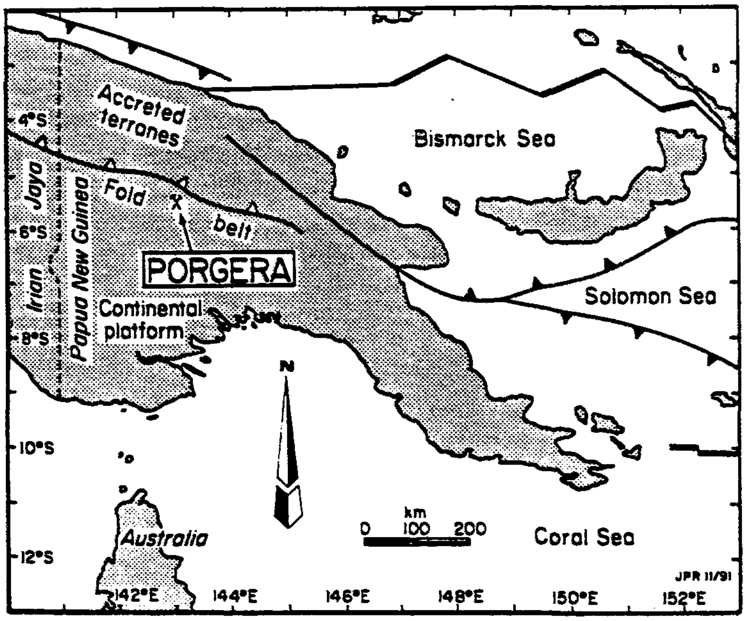


Figure 47. Sketch map showing the geographic location and tectonic setting for the Porgera Au deposit (after Reyes, 1990).

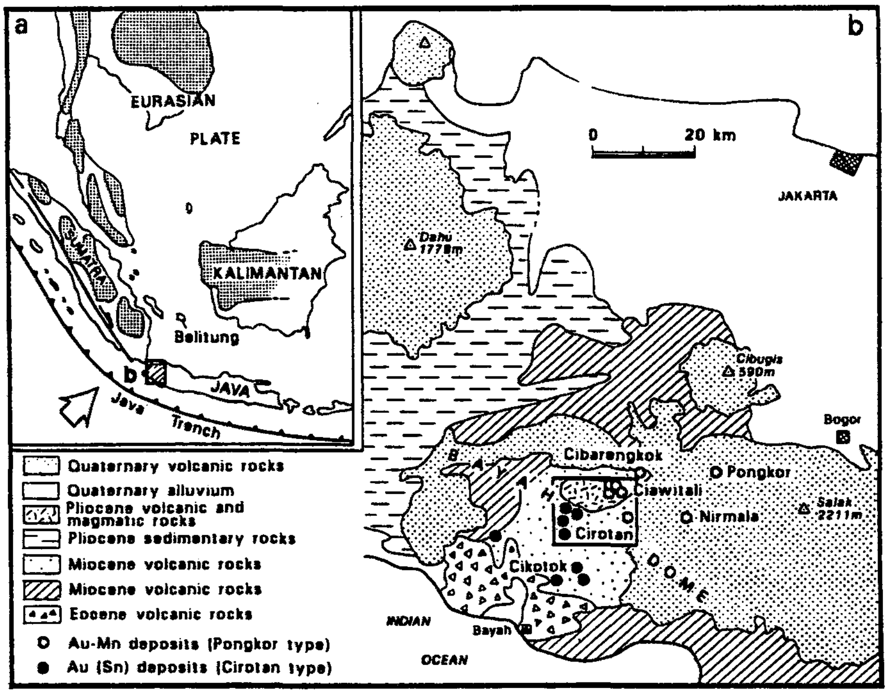


Figure 48. West Java gold deposits. (a) Exposure of Precambrian basement (stippled area) (b) Geology of North Bayah dome with main gold deposits (geology after Van Bremmelen, 1970; Hutchison, 1988; in Marcoux & Milési, 1994).

collision with the Eurasian plate in the Cenozoic. Epithermal ore deposits here are associated with abundant, still active, calc-alkaline volcanism;

b) The Masupa Ria district, located in Kalimantan, Indonesia, where mineralisation is hosted by a calc-alkaline volcanic centre consisting of a mixture of flows and fragmental rocks, predominately of andesitic composition. The predominant intermediate, calc-alkaline composition of both intrusive and volcanic rock here suggests that the most likely landform in the volcanic centre was a single or composite stratovolcano. Volcanic sediments interfingering with the adjacent Barito basin sediments indicate that the periphery of the centre was submerged (Thompson *et al.*, 1994);

c) West Java, where the gold deposits lie within and on the flanks of the Bayah Dome, Southwest of Jakarta (Figure 48(b)). This geological unit, exposed over an area about 40 x 80 km, consists of Oligocene to Quaternary calc-alkaline rhyolitic to andesitic rocks and small intrusive stocks with a few intercalations of Miocene limestone and sandstone (Van Bemmelen, 1970; Milési *et al.,* 1994; in Marcoux & Milési, 1994);

d) Four volcanogenic epithermal Au-Ag deposits which are located within the Budawang Rift of South Coastal New South Wales, Australia. These were investigated by Glaser (1988; in Glaser & Keays, 1988), and the mineralisation was found to be hosted within peraluminous rhyolites which comprise approximately 50 % of the bi-modal (rhyolitic and tholeiitic basalt) Budawang Volcanic Complex. All mineralisation occurs within the confines of the Budawang Rifts of early Late Devonian age, with which mineralisation is temporally related;

e) The Kerimenge prospect, in the Wau district, Papua New Guinea, where epithermal vein and breccia hosted Au mineralisation occurs within a block of multiply fractured dacitic porphyry of Pliocene age. The mineralisation is localised on intersecting regional fault structures, marginal to the eroded vent of a Late Pliocene-Pleistocene maar volcano (diatreme) (Syka & Bloom, 1990);

f) Togi, Japan, where detailed mapping has established a local basement of Miocene andesitic flows and pyroclastics which host mineralisation within zones of intense silicification and clay alteration (Smith & Corbett, 1990); and

g) The Faride Ag-Au epithermal deposit, which is associated with an eroded volcanic centre developed within a magmatic arc, composed of a series of volcanic centres (stratovolcanoes) of basaltic, andesitic, and dacitic composition. Extensive sheets of rhyolitic ignimbrite, possibly associated with calderas, are also present (Camus & Skewes, 1991).

2.7 FORM OF EPITHERMAL PRECIOUS METAL ORE BODIES

Epithermal deposits are extremely variable in form because of the low-pressure, hydrostatic conditions under which they are formed. Much of their geometrical variability may be attributed to the effects of permeability differences in the host rocks, since rock permeability controls the sites of fluid flow and resulting precious-metal deposition (Henley, 1985; in Sillitoe, 1993), and may be structurally, hydrothermally and (or) lithologically provided (Figure 22). At Hishikari (Ishihara *et al.,* 1986; in Sillitoe, 1988) and Paracale (Frost, 1959; in Sillitoe, 1988) for example, high-grade vein gold was precipitated beneath relatively impermeable lithologies.

2.7.1 LOW-SULPHIDATION DEPOSITS

Most LS deposits consist of cavity-filling veins with sharp boundaries, or stockworks of small veins (White & Hedenquist, 1995). Although steep veins or vein systems up to several kilometres in length are the commonest form of LS-type deposit, other geometries are also widespread. Some of the gold at Wau, Kerimenge, and Wapolu in the Philippines for example, occupies shallowly dipping veins (Sillitoe *et al.,* 1984; Billington, 1987; in Sillitoe, 1988). At Wapolu, the gold mineralisation was concentrated on a detachment fault bounding a metamorphic core complex. Also, extremely complex, conjugate vein systems and associated stockworks may be developed locally in strike-slip fault zones, especially in brittle lithologies (Figure 23), as at Mesquite, California (Tosdal *et al.,* 1991; in Sillitoe, 1993). Individual veins tend to splay and horsetail where they pass laterally into structurally less competent rocks (*e.g.* Acupan) (Mitchell & Leach, 1991).

In both the arc systems of Palau and Yap, epithermal mineralisation is characterised by quartz veins with open vugs, comb and crustification textures, and quartz-cemented breccias which are clast-supported (Rytuba & Miller, 1990). A vertical zonation of elements within lode structures is usually easily noticeable. Cu, Zn, and Pb (± Ag) tend to be enhanced lower in the systems, whereas Au and Ag occur in high concentration in the upper parts of the system (Kontis *et al.*, 1994).

2.7.2 HIGH-SULPHIDATION DEPOSITS

In HS-type deposits, although the majority of deposits consist of disseminated ores that replace or impregnate leached country rock, veins may also be important (White & Hedenquist, 1995). At Choquelimpie, two types of mineralisation have been worked - siliceous veins and hydrothermal breccia bodies. Mineralised veins are arranged in a crudely radial pattern around the central sector of a stratovolcano. The main gold-silver mineralisation occurs in a fault-controlled belt 2 km long and with a maximum width of 200 m (Gröpper *et al.*, 1991). At the Pao Prospect in North Luzon, styles of mineralisation include: quartz-enargite veins, vuggy silica alteration bodies, and pervasive low grade (0.5 ppm) gold mineralisation, associated with kaolinite-pyrite-quartz alteration, and geochemically characterised by high As and Hg (Kavalieris & Gonzalez, 1990).

At Summitville, highly silicified ore zones form irregular, steeply dipping pods, pipes, and tabular bodies. Ore zones are vertically and laterally continuous and individual zones may be traced laterally up to 500 m (Tewksbury zone) and vertically up to 200 m (Missionary). Mineralised zones closely follow intense acid sulphate altered zones consisting of vuggy silica, quartz-alunite, and quartz-kaolinite. Local joints, fractures, and unit contacts in the deposit often parallel regional faults (Gray & Coolbaugh, 1994).

Table 12 tabulates the distinguishing characteristic forms of the two styles of epithermal deposit.

Table 12: Form of epithermal deposits (after White & Hedenquist, 1995).

LOW SULPHIDATION	HIGH SULPHIDATION
(Andularia-sericite)	(Acid sulphate)
Open-space veins dominant	Veins subordinate, locally dominant
Disseminated ore mostly minor	Disseminated ore dominant
Replacment ore minor	Replacment ore common
Stockwork ore common	Stockwork ore minor

2.8 ORE AND GANGUE MINERALOGY OF EPITHERMAL PRECIOUS METAL DEPOSITS

Gold and gold-bearing minerals in epithermal gold deposits are mostly fine-grained and dispersed. What constitutes the bulk of "mineralisation" in these deposits is in fact the matrix, which is generally siliceous, and hosts the gold. They could technically be termed epithermal 'silica' deposits. Clarke (1991), implies that three forms of siliceous vein (lode) material can be recognised (Figure 49). These reflect different contributions of various types of hydrothermal fluids, and different chemical mechanisms, leading to silica and metal deposition.

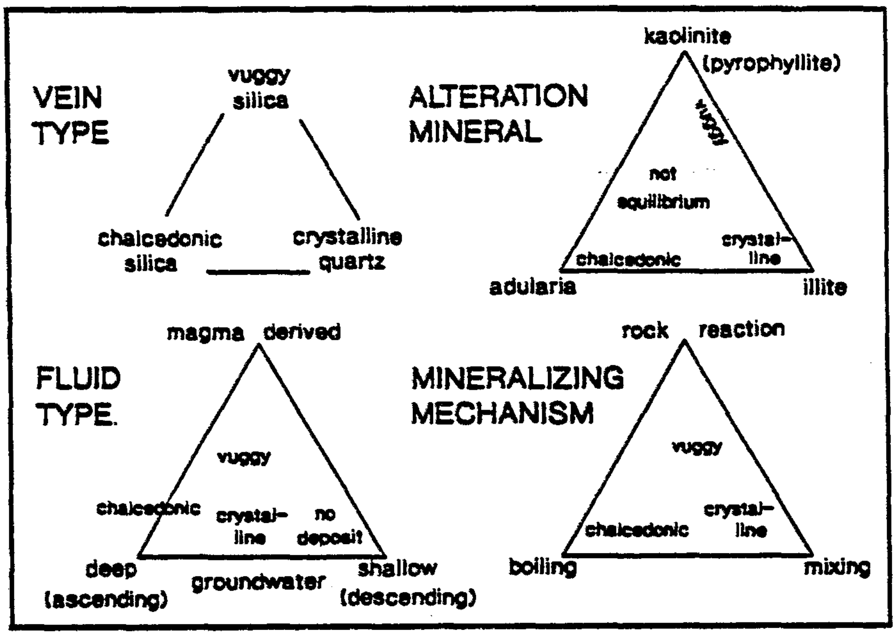


Figure 49. Schematic ternary plots illustrating the relative contribution of vein types, alteration minerals, fluid types, and mineralising mechanisms, to the three types of epithermal gold mineralisation (after Clarke, 1991).

Chalcedonic Silica type mineralisation typically contains adularia and illite (and bladed vein calcite). Crystalline Quartz type mineralisation has associated illite; locally there may be minor adularia, and kaolinite is common though usually minor. Kaolinite or pyrophyllite (in higher temperature environments) is the typical silicate mineral in Vuggy Silica mineralisation (with associated alunite). The above vein and wallrock alteration minerals (Figure 49) also reflect the hydrothermal conditions responsible for mineralisation (Clarke, 1991).

Within the precious metal ore horizon, vein mineralogy is generally a rather simple assemblage of argentite, adularia, quartz, pyrite, electrum, calcite, and ruby silvers. Tetrahedrite, stephanite, polybasite, base metal sulphides, naumannite, fluorite, barite, sericite and chlorite may occur in most deposits in small to large amounts. Even less commonly found are stibnite, realgar, rhodochrosite, rhodonite, bornite, boulangerite and a host of other minerals (Buchanan, 1981). Gold is invariably accompanied by Ag and As and, commonly, by Sb and Te. Gold:silver ratios in the Western Pacific and Metalifer Mountains of Romania are typically in the range 1:1 - 1:10, a common ratio being around 1:4 (Mitchell, 1992).

Precious metal mineralisation normally occurs in quartz or quartz-carbonate veins, veinlets or breccias, which can extend upwards into lower-grade stockworks or stockwork breccias (Mitchell, 1992). In the Grouse Greek deposit, Nevada, gold occurs as electrum and as native gold inclusions in pyrite. Silver minerals include electrum, native silver, miargyrite, pyrargyrite, acanthite, polybasite, stephanite, freibergite, tetrahedrite, aguilarite, owyheeite, and naumannite. Pyrite, arsenopyrite, galena cerussite, boulangerite, geocronite, bornite, and chalcopyrite are also present (Allen & Hahn, 1994).

There are several clear differences in ore mineralogy between the LS- and HS-styles of epithermal mineralisation (Table 13, White *et al.*, 1995). One distinction is the common occurrence of sphalerite and arsenopyrite in LS deposits, whereas sphalerite is scarce and arsenopyrite rare in HS deposits. Unlike LS examples, HS deposits commonly contain copper-arsenic minerals, especially the HS-state sulphosalts enargite and luzonite. Such sulphides, including the relatively HS-state minerals tennantite and tetrahedrite (Barton & Skinner, 1979; Hedenquist, 1995; in White & Hedenquist, 1995), are relatively rare or absent in LS gold deposits.

White & Hedenquist (1995), have noted that the gangue minerals associated with the two styles of epithermal mineralisation also show considerable overlap, but that there are clear differences as well (Table 14), differences that reflect the reactivity (pH) of the altering fluid. Quartz is common in both styles. Adularia, sericite and calcite, indicating near-neutral pH conditions (Reyes, 1990), are common minerals in LS deposits (the most common after quartz; Buchanan, 1981), but are absent from HS deposits. Minerals formed under relatively acidic conditions, such as kaolinite and alunite are common but minor in

HS deposits. In LS deposits alunite does not occur as a hypogene mineral, and hypogene kaolinite is rare, however both minerals are very common in areas affected by overprinting (Vikre, 1985) caused by downward percolation of steam-heated acid-sulphate water (Schoen *et al.*, 1974).

Table 13: Ore Minerals in Au-Rich Epithermal Ores shown as frequency of occurrence (abundance)

	LOW SULPHIDATION	HIGH SULPHIDATION
PYRITE	ubiquitous (abundant)	ubiquitous (abundant)
SPHALERITE	common (variable)	common (very minor)
GALENA	common (variable)	common (very minor)
CHALCOPYRITE	common (very minor)	common (minor)
ENARGITE-LUZONITE	rare (very minor)	ubiquitous (variable)
TENNANTITE-TETRAHEDRITE	common (very minor)	common (variable)
COVELLITE	uncommon (very minor)	common (minor)
STIBNITE	uncommon (very minor)	rare (very minor)
ORPIMENT	rare (very minor)	rare (very minor)
REALGAR	rare (very minor)	rare (very minor)
ARSENOPYRITE	common (minor)	rare (very minor)
CINNABAR	uncommon (minor)	rare (very minor)
ELECTRUM	common (variable)	uncommon (very minor)
NATIVE GOLD	common (very minor)	common (minor)
TELLURIDES-SELENIDES	common (very minor)	uncommmn (variable)

Based on compilation of mineral data from more than 130 epithermal deposits in the Southwest Pacific region (White *et al.*, in press; in White & Hedenquist, 1995), and 47 deposits in North and Central America (Buchanan, 1981; in White & Hedenquist, 1995).

Table 14: Mineralogy of Gangue in Epithermal Deposits shown as frequency of occurrence (abundance).

	LOW SULPHIDATION	HIGH SULPHIDATION
QUARTZ	ubiquitous (abundant)	ubiquitous (abundant)
CHALCEDONY	common (variable)	uncommon (minor)
CALCITE	common (variable)	absent (except as overprint)
ADULARIA	common (variable)	absent
ILLITE	common (abundant)	uncommon (minor)
KAOLINITE	rare (except as overprint)	common (minor)
PYROPHYLLITE-DIASPORE	absent (except as overprint)	common (variable)
ALUNITE	absent (except as overprint)	common (minor)
BARITE	common (very minor)	common (minor)

Based on compilation of mineral data from more than 130 epithermal deposits in the Northwest Pacific region (White *et al.*, in press; White & Hedenquist, 1995), and 47 deposits in North and Central America (Buchanan, 1981; in White & Hedenquist, 1995)

Chlorite is also characteristically present in LS-type deposits. The high silver-to-gold production ratios of most of the LS-type deposits reflect the abundance of native silver, silver sulphides, and sulphosalts (Hayba *et al.*, 1985).

Ore in HS-type deposits occurs primarily as native gold and electrum with sulphides, sulphosalts, and tellurides. Summitville and Goldfield have low silver-to-gold ratios (2:1) which reflect the high proportion of free gold and gold-bearing minerals. Julcani, Red Mountain and Lake City II have high silver-to-gold ratios (10:1) and are characterised by more abundant silver mineralisation, primarily in the form of silver sulphides and sulphosalts (Hayba *et al.*, 1985).

Precious metal values drop rapidly above the ore zone. Although the quartz vein filling extends well above the top of the ore zone, the quartz filling the vein gradually diminishes in width (Guanajuato, Pachuca, Oatman, Gooseberry, Silver Peak), and the crystalline nature of the quartz changes to an agate or chalcedony far above the ore shoot. As quartz and agate diminish in volume toward the vein tops, calcite becomes relatively more common (Buchanan, 1981).

2.9 ALTERATION MINERALOGY AND ZONING ASSOCIATED WITH EPITHERMAL PRECIOUS METAL DEPOSITS

2.9.1 INTRODUCTION

The boiling of ascending fluids in epithermal systems results in formation of H_2S-bearing steam that condenses into cool groundwater. H_2S oxidation to sulphate then takes place in vadose zones above palaeowaters (Figure 50). The resulting steamheated acidic fluids cause alteration of the host rocks (Sillitoe, 1993), (Figure 51).

Hydrothermal alteration, more or less directly attributable to vein-forming fluids, generally comprises local silicification, extensive and pervasive illitic (sericitic) assemblages, and propylitic alteration (Table 15). Illitic (sericitic) alteration is characteristic of all andesitic rocks hosting veins and stockworks, and forms the most obvious surface manifestation of epithermal systems. It surrounds veins forming zones commonly tens of metres, and in places up to several hundred metres, in width (Mitchell & Leach, 1991).

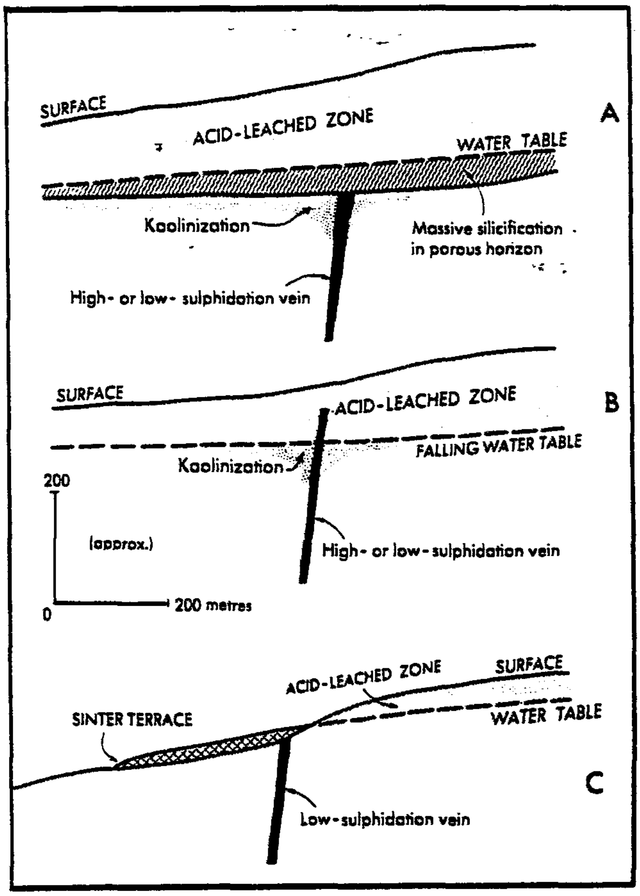


Figure 50. Shallow manifestations of epithermal systems: (A) with a stable water-table controlled by porous horizons (which may be silicified); (B) with a water-table falling during mineralisation, thereby overprinting an acid-leached zone onto a precious-metal-bearing vein; and (C) with a water-table intersecting the surface in a topographic depression, giving rise to hot springs and sinter accumulation. Note that processes involved in generation of silicified horizons at palaeo-water-tables (in A) may differ in LS and HS systems (after Sillitoe, 1993).

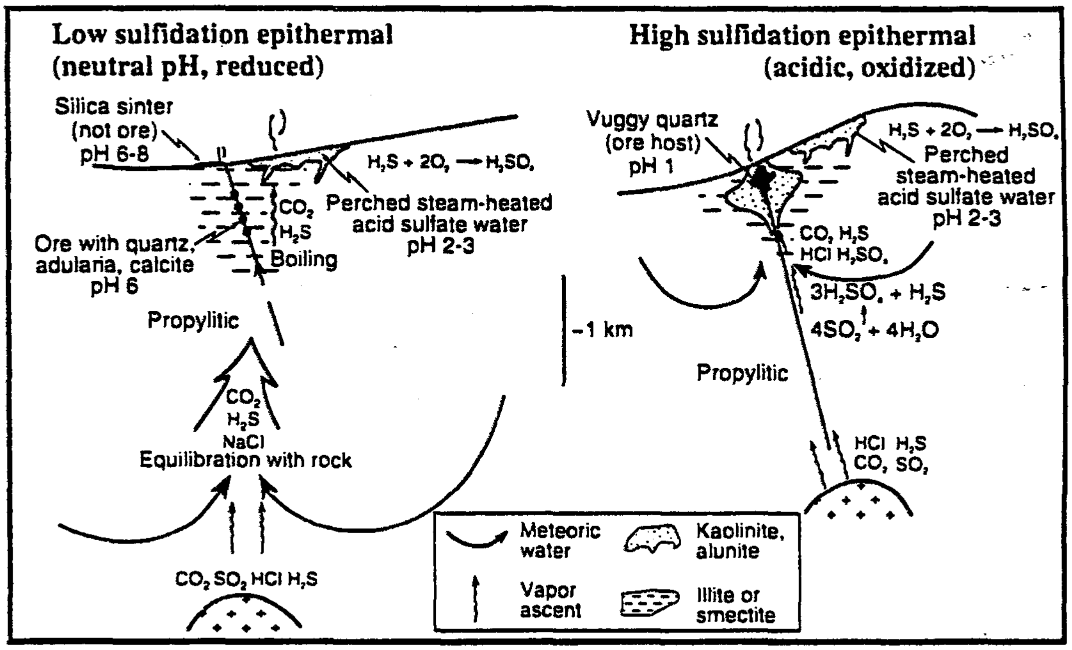


Figure 51. Generalized sketches showing the relation of fluid types to alteration zoning in the two styles of epithermal deposits. (a) In low-sulphidation sytems, the inferred magmatic input at great depth includes acid gases, NaCl, and possibly ore metals. Following wall-rock interaction and dilution, the fluid at depths of 1-2km is near-neutral pH and reduced, and in equilibrium with the host rocks at greater depths. The boiling fluid rises along permeable zones, depositing gangue and possibly ore minerals, and may discharge from near-neutral pH hot springs. The separated vapour with CO_2 and H_2S condenses in the vadose zone to form a steam-heated water, acidic from oxidation of H_2S. (B) In high-sulphidation systems, magmatic volatiles ascend with little or no modification until they reach the epithermal environment where they are absorbed by meteoric water, and the HCl and SO_2 form a highly acidic solution that leaches the rock outward from the fluid conduit. Ore metals may be introduced into this leached rock by later magmatic-meteoric fluid mixtures. Extensive propylitic alteration (including a variable assemblage of albite, chlorite, epidote, calcite, and pyrite) generally envelopes both styles of mineralisation (after White & Hedenquist, 1995).

Table 15: Petrographic Terminology for Alteration Assemblages in Epithermal Districts (modified after Heald *et al.,*1987).

Terminology	Characteristics of alteration terminology	Other terms used in literature	Notes
Silicic	Characterised by introduced silica.	Silicification	Wall rock silicified; amethyst or chalcedony typical in veins.
K-feldspar	Characterised by introduced potassium feldspar as vein material	Adularia, potassic selvedges	Typically adularia
K-feldspar-sericitic	Consists of both K-feldspar and white mica-type minerals ± pyrite.	Sericitic, potassic, potasium silicate.	Structurally controlled, disseminated about veins.
Sericitic	Consists of a mica-type mineral (e.g. illite) + quartz + pyrite; includes mixed-layered illite-smectite in which illite layers are dominant.	Phyllic quartz-sericite, illitic	
Sericitic-argillic	Consists of both white mica-type and kaolin-smectite-group minerals.		Argillic, intermediate argillic, sericitic, phyllic
Argillic	Characterised by kaolin- and smectite-group (e.g., montmorillonite) minerals; does not typically include mica-type minerals.	Intermediate argillic	Often zoned, with kaolinite nearer veins and montmorillonite farther from veins.
Advanced argillic	Characterised by minerals representing extreme base leaching (e.g., kaolinite) and sulphate or halogen fixation (e.g., alunite)	Argillic, alunitic, quartz + alunite	
Chloritic	Characterised by the introduction of a chlorite component (usually Fe-rich) into the vein; may occur alone or with hematite, pyrite, or other sulphides, or quartz.	Chloritic	A vein mineral or selvedge or, rarely, disseminated in wall rock.
Propylitic	Characterised by chlorite, albite, epidote, carbonate ± pyrite, Fe oxides, and minor sericite	Quartz-chlorite-pyrite	Typically a regional alteration
Potassium metasomatism	Characterised by introduced potassium resulting in recrystallisation of wall rocks to K-feldspar- and biotite-rich assemblages	Potassium silicate	Typically a broad-scale regional alteration

Permeability is an important controlling factor in alteration. The alteration mineral assemblage additionally varies with temperature, fluid composition, and primary mineralogy of the host-rock. Changes in geometry and (or) positioning of heat sources with time and sealing of permeable zones by deposition of silica, calcite, *etc.* will cause overlapping of alteration assemblages (Silberman & Berger, 1985).

2.9.2 LOW-SULPHIDATION EPITHERMAL SYSTEMS

Ore-associated alteration in LS deposits is produced by near-neutral pH thermal waters, with temperature decreasing both with decreasing depth and with increasing distance from the fluid conduits. With increasing temperature, smectite (stable at $<160°C$) gives way to interstratified illite-smectite, whereas illite by itself is generally stable at $>220°C$ (Reyes, 1990). This progression in thermal stability commonly results in a clear upward and outward zonation of minerals from LS ore bodies (Figure 52). The ore zone contains minerals indicating the highest pH (such as adularia and calcite), since boiling in the conduits causes CO_2 and H_2S loss, and consequent increase in pH (Simmons & Christenson, 1994; in White & Hedenquist, 1995).

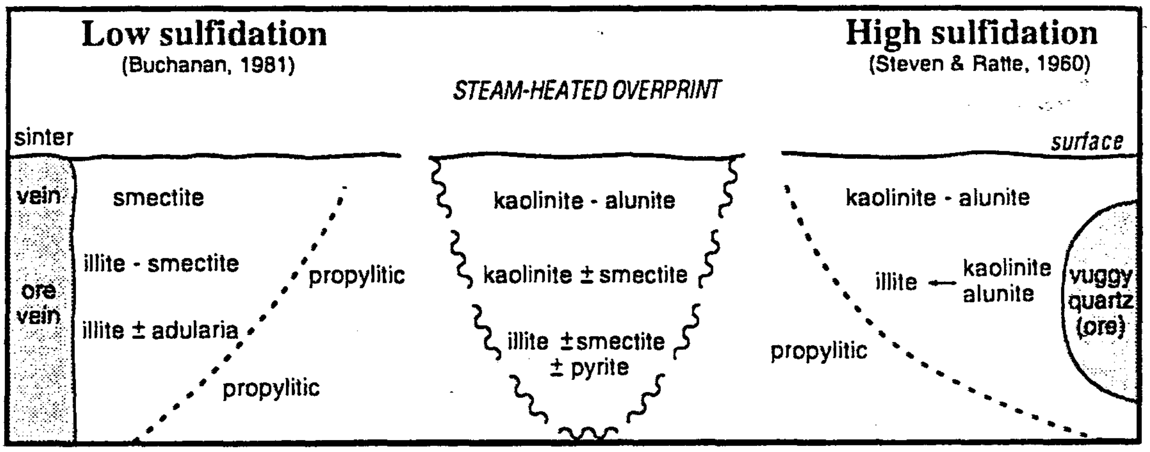


Figure 52. Distribution of hydrothermal alteration associated with high-sulphidation and low-sulphidation deposits. The alteration mineralogy varies both vertically and laterally. Quartz is stable in all areas. Propylitic alteration occurs in regions of low water rock ratios, i.e. outside conduit zones, and its mineralogy is controlled by rock composition. Typical minerals include albite, calcite, chlorite, epidote, and pyrite. Steam-heated overprint can occur in either low or high-sulphidation environments, though in the latter the hypogene and steam-heated alteration mineralogies are similar. The effects of the steam-heated overprint are most apparent in the low-sulphidation environment, as the alteration mineralogy is markedly different from that produced by hypogene fluids (after White & Hedenquist, 1995).

In general, LS-type deposits are characterised by the predominance of sericitic alteration that often borders a silicified zone near the vein (Figure 53). Also near the vein, fine-grained potassium feldspar and/or chlorite are often disseminated in the wallrock. The sericitic zone typically grades outward into a propylitic zone. An argillic zone between the sericitic and propylitic zones is sometimes present (Hayba *et al.*, 1985). Propylitic alteration forms within the zone of significant boiling or mixing. In some North American deposits, propylitic alteration is considered to have preceded fracturing, which subsequently facilitated fluid flow in conduits and mineralisation accompanying falling temperature (Sander, 1990; in Mitchell & Leach, 1991).

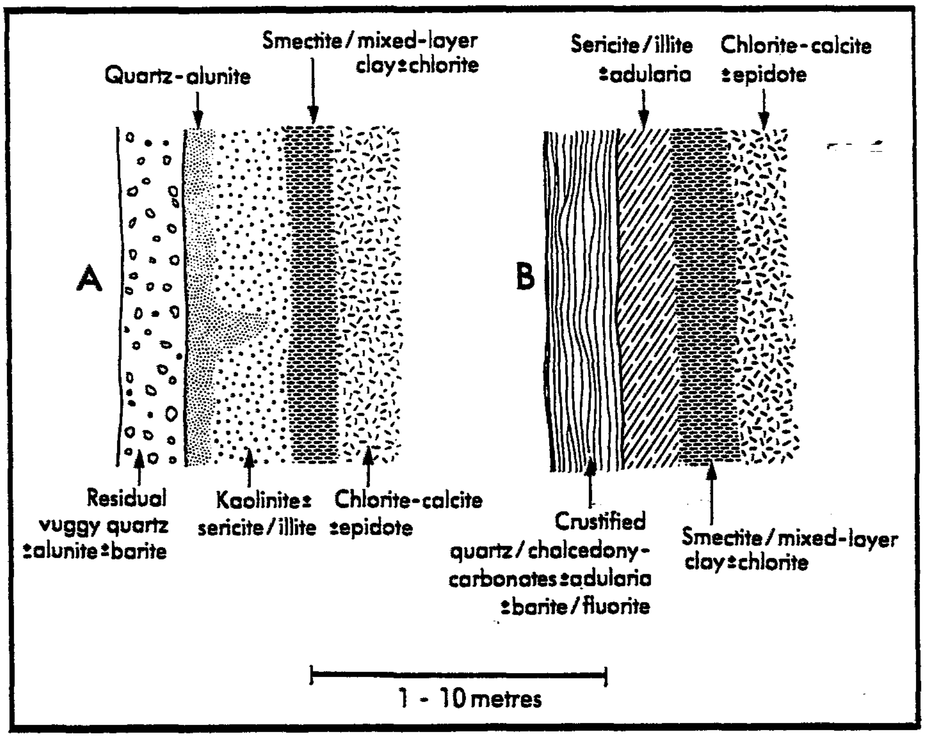


Figure 53. Schematic alteration zoning outward (left to right) from epithermal precious-metal veins (or other geometrical forms) of (A) HS and (B) LS types. The patterns result from fluid-rock interactions and declining fluid temperatures, and in (A) involve substantial rise in pH. The propylitic assemblages observed commonly as outermost zones are probably generated during early, higher temperature stages of the hydrothermal systems, possibly with much lower fluid/rock ratios (after Sillitoe, 1993).

In at least a few prospects the boundary between illitic and propylitic alteration is knife-sharp, although at some (*e.g.* Motherlode) there may be a "mixed zone" up to 100 m wide. The width of the propylitic zone can exceed 500 m, while at the surface in some systems it is virtually absent (Mitchell & Leach, 1991).

2.9.3 HIGH-SULPHIDATION EPITHERMAL SYSTEMS

A definitive characteristic of HS-type deposits is the association of advanced-argillic alteration with the ore (Figure 51). Kaolinite, usually accompanied by alunite, occurs close to the vein and is often coextensive with silicification (Hayba *et al.*, 1985), (Figure 52). Alunite $\{KAl_3(SO_4)_2(OH)_6\}$ is a characteristic mineral produced under conditions of acid-sulphate alteration of silicate rocks (Hemley *et al.*, 1969). Further from the vein, argillic alteration, sometimes intermixed with sericitic alteration, surrounds the zone of advanced-argillic alteration. The argillic alteration zone is commonly mineralogically zoned, with kaolinite nearer the vein and smectite farther from the vein (Figure 53). The outermost alteration zone consists of propylitic alteration (Hayba *et al.*, 1985).

Alteration due to acid-sulphate fluids is generally confined to permeable structures (Reyes, 1990). Exploration drilling to 3 km depth in neutral-pH geothermal systems in the Philippines (Reyes 1990; Reyes *et al.,* 1993; in Hedenquist, 1995) has penetrated zones of acidic alteration, generally associated with fractures, some yielding acidic fluid. Neutral-pH alteration can be divided into four progressively outward envelopes/zones on the basis of key clay minerals. These are: kaolinite (ambient to 120°C), dickite ± kaolinite (120 - 200°C), dickite ± pyrophyllite (200 - 250°C), and pyrophyllite ± illite (230 - 320°C). (Reyes, 1990) This overall mineral sequence reflects the progressive neutralisation of the acid by fluid-rock reactions (Hemley & Jones, 1964; in Berger & Henley, 1988). This pattern of alteration assemblages is particularly well displayed at Goldfield and at Summitville where it has been mapped in detail (Steven & Rattè, 1960; Harvey & Vitaliano, 1964; in Heald *et al.*, 1987). Here porous or vuggy masses of silica sometimes form branching, elongate bodies following joint patterns or faults and other times pipelike bodies of irregular outline. Along strike the silicified zones are characteristically discontinuous, and there is little continuity with depth (Berger & Henley, 1988). The above alteration sequence envelopes the array of silicic bodies.

The apparent proximity of HS-type deposits in time and space to a magmatic source may explain relatively high sulphur fugacities reflected by extensive acid-sulphate alteration and enargite-covellite-pyrite mineral assemblages. Thermodynamic calculations (Brimhall & Ghiorso, 1983; Whitney, 1984; in Heald *et al.*, 1987) indicate that acid, SO_2-rich solutions are expected products of degassing relatively oxidised magmas. SO_2 gas disproportionates below about 400°C in the presence of water to H_2S and H_2SO_4 (Holland, 1965; 1967; Sakai & Matsubaya, 1977; in Heald *et al.*, 1987), largely by the following reactions:

$$\begin{aligned} SO_2 + H_2O &= H_2SO_3 \\ 4H_2SO_3 &= 3H_2SO_4 + H_2S \end{aligned}$$

The interaction of the sulphuric acid solutions with the wall rocks produces the deep advanced argillic alteration associated with the HS-type deposit ores.

2.9.4 HYPOGENE AND SUPERGENE ALTERATION

Solfataric hypogene alteration and supergene alteration can overprint earlier patterns of alteration, particularly in the upper levels of systems (Silberman & Berger, 1985), (Figure 54). Acid-leached zones are thicker where palaeowater tables are deep, as in arid regions and (or) beneath topographic highs (such as stratovolcanoes in mountainous terrain). They tend to be more widespread than alteration zones associated directly with subjacent precious-metal mineralisation (Figure 50) (Sillitoe, 1993). One consequence of these complications is that they seriously affect geochemical patterns as the acid-leaching tends to remove many of the pathfinder elements to ore and disperses them laterally, or down to deeper levels, depending on the local hydrology (Silberman & Berger, 1985). Shallow acid-leached (advanced argillic) zones produced by condensation and oxidation of acid gases released by surficial boiling are, for example, still being formed above the Ladolam and Kabang deposits (Davies & Ballantyen, 1987; in Sillitoe, 1988; and Licence *et al.*, 1987).

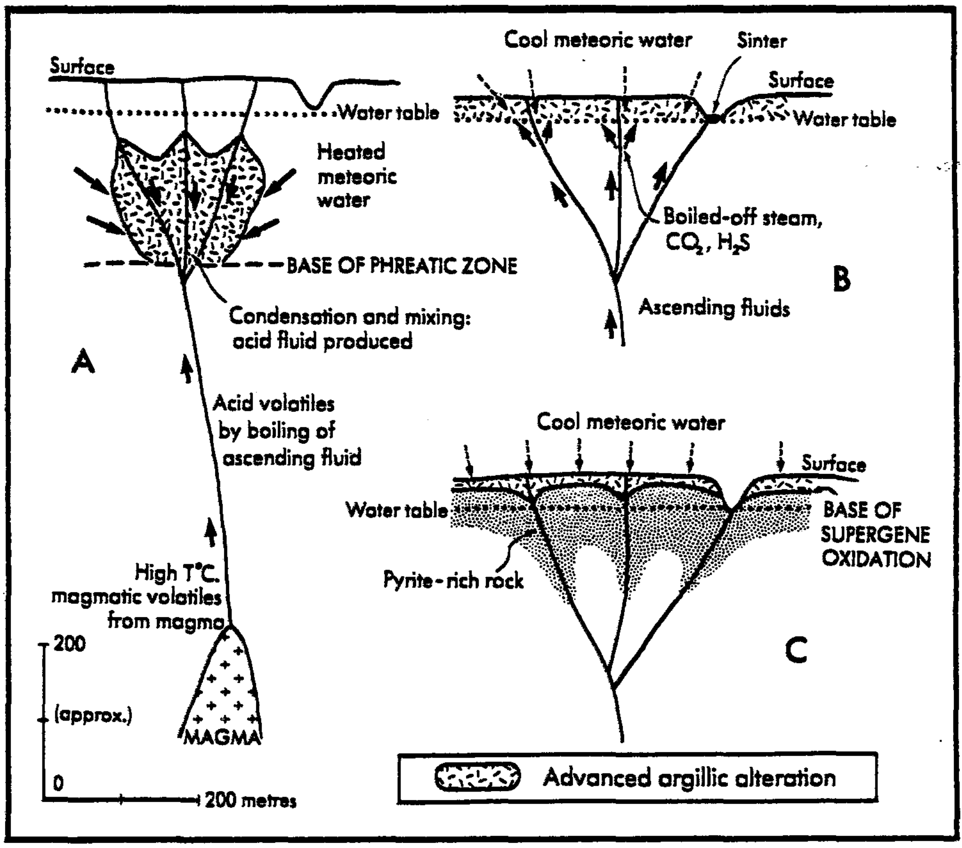


Figure 54. Origins of three types of advanced argillic alteration: (A) deep hypogene due to condensation of acidic magmatic volatiles and dissolution of condensate in meteoric water; (B) shallow hypogene (acid leaching) due to condensation and absorption of boiled-off volatiles and subsequent oxidation of H_2S in the vadose zone; and (C) supergene due to oxidation of sulphides above a water table. Only (A) is capable of generating precious-metal mineralisation (HS type), although both (B) and (C) may overlie and (or) overprint precious-metal mineralisation (HS and LS types). Note that the barren gap between advanced argillic alteration and the underlying stock in (A) may be eliminated by extreme telescoping (after Sillitoe, 1993).

At numerous epithermal prospects and in the hydrothermal systems in Negros, outcrops with kaolinite and in some cases alunite can be explained by supergene oxidation of hypogene pyrite and resultant acid attack. Since the pH of the supergene fluids is determined partly by the pyrite content in the oxidising rock, supergene acid sulphate assemblages are commonly confined to the surficial parts of pyrite-rich illitic alteration zones rather than those of propylitic alteration (Mitchell & Leach, 1991).

Sinter aprons or terraces are generated where upflowing hydrothermal fluids discharge at the Earth's surface and precipitate silica, *i.e.* where the water table intersects the surface in valleys of basins. Consequently, sinters may give way laterally to acid-leached zones (Figure 50). Sinters however have not been recognised above HS systems. This is attributed to the suppression of silica precipitation from low-pH fluids (Fournier, 1985); active acidic springs ($pH < 4$) in both HS and LS systems do not have associated sinters. In addition, the topographically prominent volcanic landforms and resulting deep water tables above most HS systems often preclude hot-spring discharge until there has been extensive lateral flow and dilution (Sillitoe, 1993).

PART C: GUIDELINES FOR EPITHERMAL EXPLORATION

3.1 INTRODUCTION

The dramatic increase in exploration for epithermal gold over the last few decades has resulted largely from new discoveries in the Western USA, from introduction of the heap-leach extraction process facilitating bulk mining of low-grade deposits, and from the realisation that most epithermal gold is associated with calc-alkaline volcanic arcs recognised since the late 1960's, as the most spectacular surface manifestation of ongoing plate tectonic processes. This resulted in reassessment of the gold potential of subduction-related magmatic arcs, with exploration increasing in the American Cordillera and Caribbean, and progressing Northward through the Western Pacific margin as Australian entrepreneurs ventured further from home. This led to the discovery of the world's largest epithermal deposit on Lihir Island in New Guinea. The Lihir deposit, with nearly 600 tonnes of recoverable gold, lies within the breached crater of a statovolcano (Mitchell & Leach, 1991).

The modern understanding of the dynamic structural and chemical settings of epithermal and sub-volcanic deposits allows us to quite effectively pinpoint the most probable foci of mineral deposition in any given district. This approach is fundamentally different from the historically more reactive mode of exploration, which generally involved indiscriminate testing of each and every occurrence of elevated gold resulting from 'blanket geochemical' surveys. Structural understanding, combined with understanding of the chemical processes involved, provides a major tool for exploration particularly in terrains obscured by deep weathering. Geological interpretation based on integration of field mapping with high resolution magnetic, radiometric and other geophysically and remotely sensed imagery, provides a powerful complementary tool to the traditional geochemical and geophysical targeting techniques. Thus, timely and effective geological input is capable of reducing exploration risk and providing a more cost-effective basis for drilling and development (Henley & Etheridge, 1995).

The challenge of exploration in high level volcanic environments lies in unravelling the complexity brought about by the myriad of influences on the kinetics of gold deposition in reactive rocks undergoing brittle deformation (Henley & Etheridge, 1995). Genetic models are attempts to explain the origin of these features by physicochemical processes. These mineral deposit models for epithermal gold deposits can be efficient tools for exploration and resource assessment, since modelling of the chemical environment of fluid flow and ore transport and deposition in fossil epithermal systems, may be achieved with greater confidence (Ericksen, 1988). The epithermal environment is extremely diverse in character, as a variety of physical and chemical processes occur within a complex and dynamic geological environment. Consequently, the features observed, and their spatial relationships, vary widely. The diversity of features observed, and their significance in exploration, can only be understood with the aid of a strong conceptual understanding of the processes which occur in hydrothermal systems (White & Hedenquist, 1990).

Guidelines for exploration vary according to the scale at which work is conducted, and are commonly constrained by a variety of local conditions. On a regional scale tectonic, igneous and structural settings can be used, together with assessment of the depth of erosion, to select areas for project area scale exploration. At project area scale, direct (*i.e.* geochemical) or indirect guidelines may be used. Indirect methods involve locating and interpreting hydrothermal alteration as a guide to ore, with the topographic and hydrologic reconstruction of the system being of high priority. These pursuits may involve mineralogical, structural, geophysical or remote sensing methods. On a prospect scale, both direct and indirect methods may be used; however, they can only be effective in the framework of a sound conceptual understanding of the processes that occur in the epithermal environment, and the signatures they leave (White & Hedenquist, 1990). For example, provided care is taken to recognise the presence or absence of CO_2, and to correct apparent inclusion salinities accordingly, fluid inclusions, at an early stage of exploration, may be used to determine the salinity of the original hydrothermal system. The salinity of the hydrothermal system has an important outcome on the solubility of gold in epithermal systems (Section 1.5). Similarly, through comparison with alteration zones of active geothermal systems, the exploration geologist may determine the potential distribution and types of ores in a fossil geothermal system (Ericksen, 1988).

The concentration of gold in economic deposits (usually measured in parts per million) is too low to allow its direct detection by geophysical techniques. However, gold mineralisation is commonly accompanied by vein quartz or disseminated sulphides which cause significant contrasts in electrical properties between mineralisation and surrounding altered rocks, making mineralised areas amenable to geophysical detection. Surveys are generally most useful when carried out relatively early in the history of a project to assist in semi-regional exploration and to help guide initial drilling (Irvine & Smith, 1990). Although the extremely variable nature of the epithermal environment and rather subtle characteristics of the physical signatures of epithermal systems may make geophysical surveys appear to be a frustrating exercise (Allis, 1990), the integrated geological-geophysical interpretation derived from airborne geophysics provides a basis of targeting potential ore environments (*e.g.* calderas, prospective structures, alteration zones, *etc*). The aim of follow-up geology, geochemistry and surface geophysics is to localise mineral concentrations within these target areas (Webster *et al.*, 1989). The choice of the appropriate geophysical methods depends on the stage of the exploration programme, and the degree of access in the prospect area (Allis, 1990).

Historically, compared to widespread evidence of extinct hydrothermal activity, the occurrence of epithermal ore deposits has been significantly scarce, indicating that only a small proportion of hydrothermal systems actually form ore, either now or in the past (Hedenquist & Lowenstern, 1994). The strategy thus employed is to utilise available exploration techniques to identify prospective 'epithermal deposit'-hosting regions of the earth's surface. Subsequently, in order to accurately delineate the underlying potential for mineral wealth, the explorationist needs to effectively probe into and penetrate through the extremely complex surface features of the epithermal environment. He is ultimately required to establish a model of both the hydrological system and mineralisation controls intrinsic to the particular prospect. Only then can delineation and evaluation procedures commence.

3.2 EXPLORATION CONCEPTS IN EPITHERMAL ENVIRONMENTS

The following is a selection of specific aspects of exploration pertinent to the epithermal environment. These features should be considered whilst planning an exploration strategy designed to investigate primarily the potential for epithermal-type precious-metal ore deposits.

3.2.1 SETTING

a) GEOLOGY, TOPOGRAPHY AND ALTERATION

When designing an exploration programme within a volcanic terrain, it is essential to consider the geological setting and hydrological conditions (*e.g.* relief) that prevailed at the time of mineralisation (White & Hedenquist, 1995). Attention should also be given to the structural configuration of the volcanic units overlying strata, if any, and subsequent geologic events such as weathering or metamorphism. Exploration for epithermal deposits should start with recognition of a potential gold-bearing volcanic unit that has the physical permeability to be a host, whatever the origin of that permeability (Worthington, 1981). In relatively low-relief areas, major shallow conduits are likely to be distributed above a feeder zone extending into the basement. The resulting alteration pattern commonly has an approximately symmetrical distribution in this setting due to mushrooming of the ascending thermal water. In contrast, in high relief areas (*e.g.* andesitic stratovolcanoes), there is a large degree of lateral flow in geothermal systems, up to 10km or more, resulting in strongly asymmetric alteration zones (Figure 5) relative to the upflow zone (White & Hedenquist, 1995).

b) PALAEOISOTHERMS AND LEVEL OF EROSION

As many alteration minerals are stable over limited temperature and/or pH ranges (Reyes, 1990), they provide information with which to reconstruct the thermal and geochemical structure, and hence the hydrology of the extinct hydrothermal system (White & Hedenquist, 1995). During exploration of epithermal prospects, this information allows general palaeoisotherms to be deduced from the distribution of alteration minerals, which in turn helps to locate conduits of palaeoflow and to determine the level of erosion. The erosion level is significant as most epithermal ore is deposited over the temperature range of ~180 to 280°C (Figure 55), equivalent to a depth below the palaeo-watertable of about 100m to 800-1500m (Hedenquist & Henley, 1985). Prospects with indications of low palaeotemperatures are encouraging, whereas indications of palaeotemperatures >280°C suggest the depth interval with potential for epithermal ore has been eroded (White & Hedenquist, 1995).

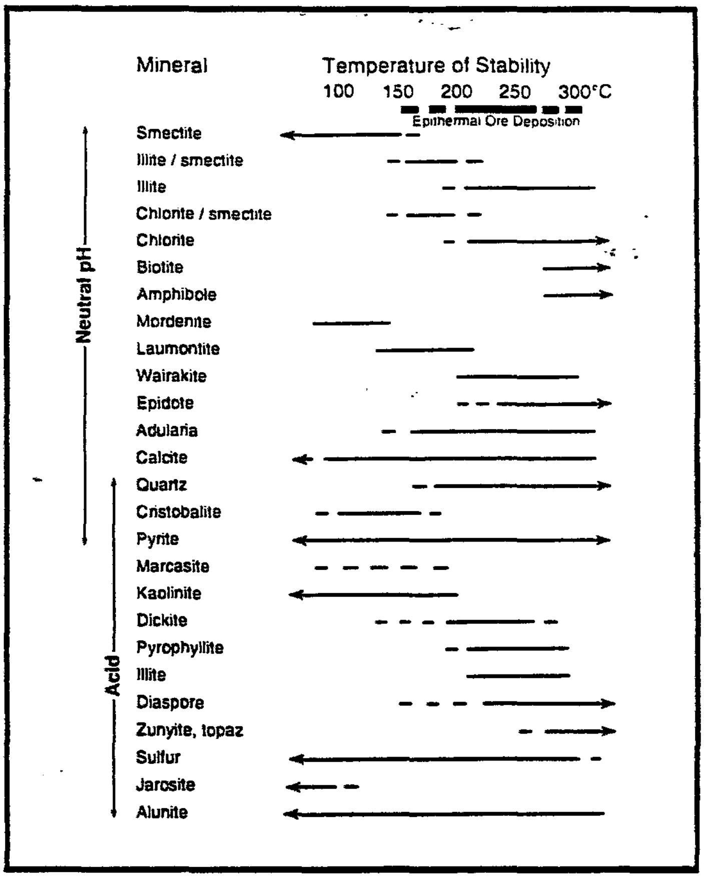


Figure 55. Temperature stability of hydrothermal minerals common in the epithermal environment. Some workers disagree on the absolute temperature of first appearance of some minerals, and the temperature of transition from one day to another, but the relative stability is similar in geothermal systems throughout the world. Identifying zones of mineral assemblages may be more meaningful in indicating palaeotemperatures than single minerals (Reyes, 1990). In low-sulphidation systems the principal gangue minerals are quartz, calcite, and adularia; in high-sulphidation systems the principal gangue mineral is quartz (after White & Hedenquist, 1995).

c) STRUCTURE, ALTERATION AND PALAEOFLOW CONDUITS

Structural studies probably still provide the most effective means of identifying the locality(s) of palaeoflow conduits. Structure, particularly fracture analysis, provides the main clue to flow design in the fossil system. Alteration associations in active geothermal systems can likewise be used since the distribution of alteration minerals in these systems are relatively similar. Resulting alteration assemblages (as in (a) above) may thus provide key data on the level of exposure within a system (Browne, 1978; Henley & Ellis, 1983; in Henley, 1985). Recognition of other depth indicators such as hydrothermal eruption or vent breccias is also a powerful targeting tool. This is especially so where there is evidence that multiple brecciation episodes have occurred.

3.2.2 GEOCHEMISTRY

a) POST- MINERAL COVER

Epithermal gold-silver deposits occur in young volcanic terrains where post-mineral cover commonly impedes the search for mineralisation (Smith & Corbett, 1990). A comprehensive understanding of the regional geology in a prospective area may thus be hampered by extensive vegetation, a regolith or residual soil blanket, deep weathering and poor bedrock exposure, as well as by outcrops which only occur in drainage and/or as a few competent peaks (Simmons & Browne, 1990). The decision to use primary or secondary geochemical dispersion in exploration, or drainage, soil, vegetation, or rock survey, will be determined by local conditions and the nature of the exploration programme (Clarke & Govett, 1990).

b) DISPERSION PATTERNS

An understanding of primary dispersion patterns for trace elements will aid the design of appropriate drainage or soil surveys. For example, an awareness of the probable rock-geochemical signature beneath areas of residual soils, permits subdued anomalies derived from the primary rock halo (Figure 56), to be sought - rather than relying on dispersion from lode material to provide an adequate target. A basic understanding of the prevailing rock-geochemical response is thus regarded as indispensible for designing appropriate stream sediment and soil geochemical surveys (Clarke & Govett, 1990). An example of the surface trace element distribution of the Summitville area is depicted in Figure 57.

c) MINERALISATION STYLE

Effective assessment of an epithermal gold prospect requires establishing whether it is LS- or HS-style. The distinct origins of the two styles result in differences in the geochemical signatures of the ores (Table 16). With the use of these generalised chemical differences to guide the design of geochemical surveys, and correct application of alteration zoning to focus detailed assessment towards ore potential, we can explore these systems more efficiently (White & Hedenquist, 1995).

Table 16: Geochemical Associations of Epithermal Deposits (After White & Hedenquist, 1995).

	LOW SULPHIDATION	HIGH SULPHIDATION
ANOMALOUSLY HIGH	Au, Ag, As, Sb, Hg, Zn, Pb, Se, K, Ag/Au	Au, Ag, As, Cu, Sb, Bi, Hg, Te, Sn, Pb, Mo, Te/Se
ANOMALOUSLY LOW	Cu, Te/Se	K, Zn, Ag/Au

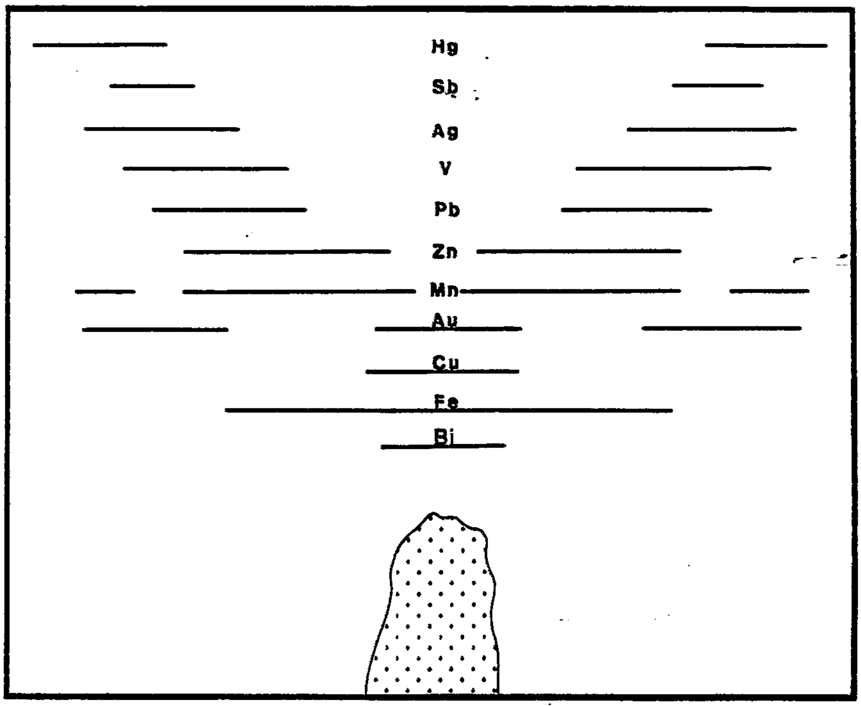


Figure 56. Classical hypogene metal and mineralogical zoning scheme of Emmons (1927) showing central and peripheral gold zones. The peripheral gold zones relate to epithermal mineralisation (modified after Jones, 1992).

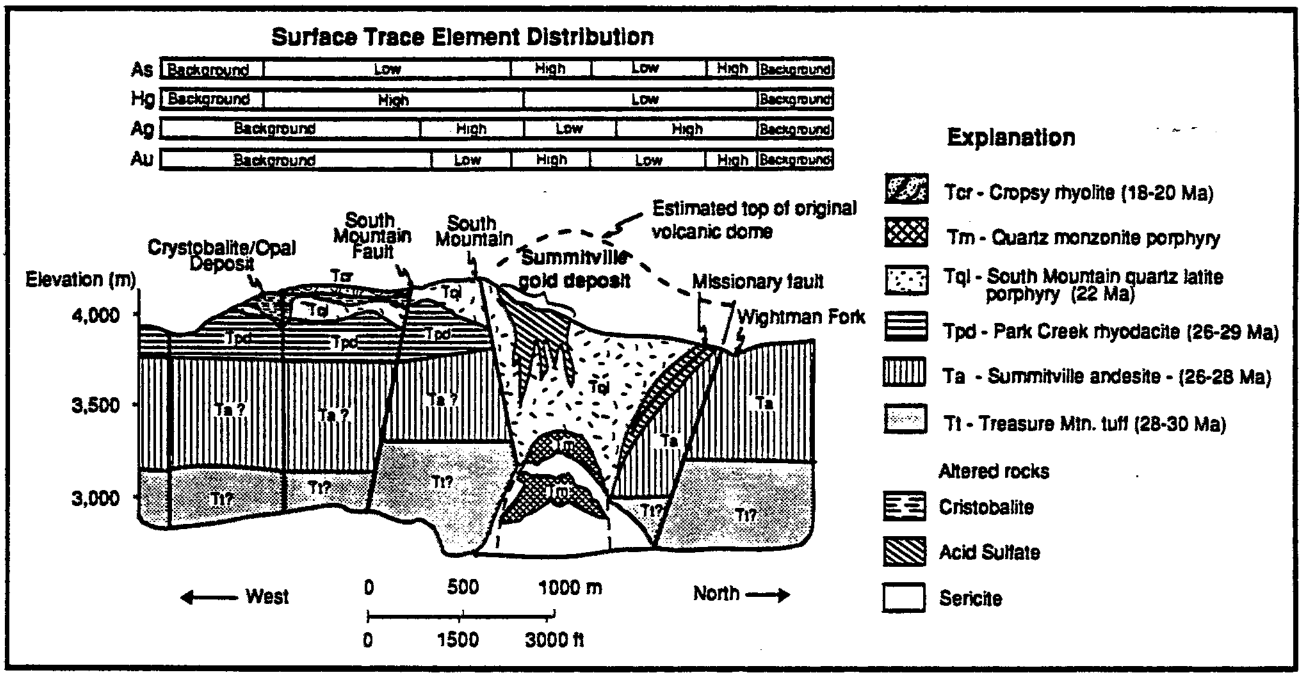


Figure 57. Schematic cross section of the Summitville area showing the acid sulphate Au deposit, cristobalite- and opal-bearing hot spring deposit, and the sericite-pyrite altered quartz monozite pophyry at depth. The surface trace element distribution is projected above the cross-section (after Gray & Coolbaugh, 1994).

d) TYPICAL ELEMENTS

As no single element, except possibly Au itself, is a definitive indicator, a multi-element approach reduces the possibility of erroneous interpretation (Clarke & Govett, 1990). The typical suite of elements in anomalous concentrations associated with epithermal Au deposits includes Au, Ag, Sb, As, Hg, Te/Se, Cu, Pb and Zn (Henley, 1993; in Kontis *et al.*, 1994; Clarke & Govett, 1990).

e) ^{18}O & BASE METAL ANOMALIES

"Negative" ^{18}O anomalies may serve as an additionally useful guide to mineral exploration. These are isotopically unambiguous and areally extensive, and they occur in altered sedimentary and volcanic host rocks of both epithermal deposits and geothermal systems. These rocks are conspicuously depleted in ^{18}O relative to their peripheral and unaltered equivalents (Field & Fifarek, 1985). Also, it may be anticipated that areas of significant mineralisation may have relatively high base-metal concentrations in waterways, since the same geothermal fluids that are responsible for epithermal gold deposits, carry significant quantities of base metals in chloro- and bisulphide complexes (Hodder, 1987).

f) ALTERATION

The effects of hydrothermal alteration associated with epithermal gold deposits differ from prospect to prospect, depending on factors such as geological age, rock type, nature of the hydrothermal fluids and level of erosion (Irvine & Smith, 1990). Regionally pervasive propylitic alteration appears to be the areally significant target for a regional-scale bedrock exploration programme (Clarke & Govett, 1990). In HS deposits, the ore is typically closely associated with the zone of MOST acidic alteration. In contrast, in LS deposits, the ore is associated with the LEAST acidic alteration (*i.e.* adularia and calcite or illite) (White & Hedenquist, 1995). It should be noted however that this alteration (and silicification) can be due to hydrothermal activity that is not part of an ore-forming system. Confidence that the alteration did in fact result from an ore-forming system may be provided by recognition of the association of structural and lithologic parameters (forming part of an exploration model), or if geochemical samples reveal anomalous concentrations of Au and Ag or the characteristic pathfinder elements - Sb, As, Hg, Tl, Mo, W, and Mn (Clarke & Govett, 1990). There are mechanisms that tend to enhance the signatures of Hg, As, and W in soil survey (Readdy, 1988).

g) CLAY MINERALOGY

Another useful exploration guide is provided by clay mineralogy. The careful determination of clay mineralogy can provide a guide to temperature and, used carefully in conjunction with an appropriate boiling point vs depth curve (Figure 58), to depth of formation (Henley, 1991). Clays and other OH- and SO_4-bearing minerals have the most potential to indicate palaeotemperature zoning. The recent availability of field portable infrared spectrometers (such as the PIMA II) has revolutionised the ability to map alteration correctly and in detail (*e.g.* interstratified clays can be identified) (White & Hedenquist, 1995). Integration of aspects of photogeology and remote sensing into geochemical exploration programmes for epithermal gold deposit-related clays can improve the potential discovery rate of geochemical surveys (Readdy, 1988).

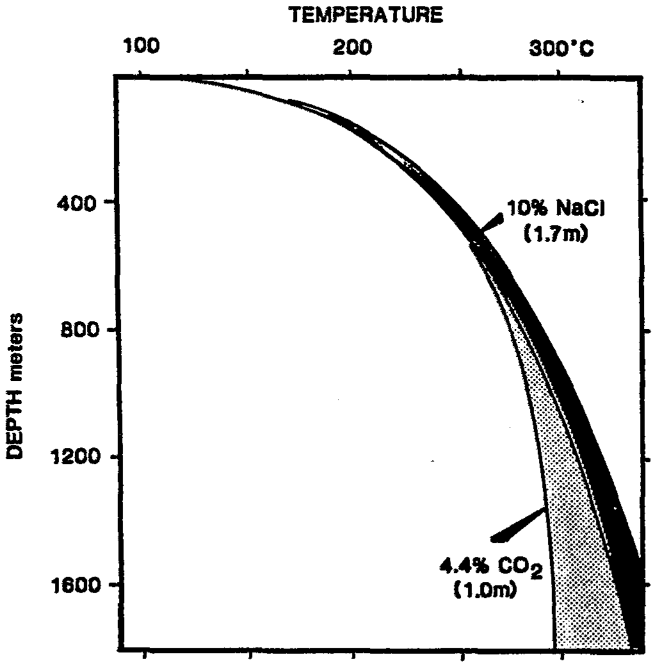


Figure 58. Temperature-depth relations for 'boiling fluids' containing NaCl and CO_2 in relation to boiling point vs. Depth relations for pure water (after Henley, 1991).

3.2.3 GEOPHYSICAL SURVEYS

a) REGIONAL SCALE AIRBORNE SURVEY

When geophysical surveys are used in conjunction with geological and geochemical data (White *et al.*, 1995), combined with an understanding of mineralising processes, they can be an important aid in exploration (Irvine & Smith, 1990). On a regional and semi-regional scale, high-resolution airborne magnetic, radiometric and gravity surveys can assist in locating alteration systems and the structures that control deposits (White & Hedenquist, 1995). Filtering and image processing techniques are valuable tools for extracting the maximum information content from geophysically acquired data (Irvine & Smith, 1990).

b) DEEP-SEATED STRUCTURE

In most cases of hydrothermal alteration there are significant changes in physical parameters which are detectable by geophysical techniques, thereby facilitating delineation of alteration systems and specific gold-related components within them (Irvine & Smith, 1990). The most common geophysically targeted geological characteristic is that of deep-seated structures, since these provide zones of weakness for subsequent intrusion or channel-ways for hydrothermal, mineral-bearing solutions. This factor dictates the scale of regional geophysical surveys used in exploration. The data set must cover a sufficiently large area to recognise disconnected segments of structures, cross-cutting features and localised splays - all of which can control mineralisation. Structural interpretations may localise dilational positions (*e.g.* Figure 26) within such deformation zones (Webster *et al.*, 1989). These are highly prospective.

c) PROSPECT-SCALE GROUND SURVEY

Ground geophysical methods for prospect-scale exploration include high-resolution resistivity mapping (*e.g.* Induced Polarisation (IP)/Resistivity, Controlled-source Audio-magneto-telluric (CSAMT), Transient Electromagnetics (TEM), and Schlumberger Surveys) which can locate changes in rock properties (*i.e.* pyritization, silicification and hydrothermal clay alteration) that may be related to ore-forming processes (White & Hedenquist, 1995).

d) HYDROTHERMAL FOCUS

For prospect-scale exploration - defined as responses of up to a few 100m from a deposit - within an extensive zone of alteration, the most prospective site for mineralisation will be associated with the focus of the hydrothermal or geothermal activity (Silberman & Berger, 1985) and may be provided by an internal zonation of the alteration. For gold deposits in the Southwest Pacific, the focus is commonly marked by high values of K with depletion of

Na (potassic alteration) except, of course, when hosted by K-rich bedrock. Silicification may also be most intense in this area (*e.g.* Hishikari; Irvine & Smith, 1990). Gold mineralisation is usually associated with quartz veins and/or quartz-vein systems and associated silicified rocks, which show high resistivity and are one of the most common targets of electrical exploration methods (Nishikawa, 1992). However, if the targets are situated at depth and are small, they are difficult to detect directly. Peak levels of Au, Ag and the pathfinder elements (in particular As, Sb and Tl) during prospect evaluation should also coincide with the focus of past hydrothermal activity. For gold-telluride deposits, elevated Hg and Te (and V) can also be characteristic (Clarke & Govett, 1990).

e) INDUCED POLARISATION

Although pyrite is widely distributed in some alteration systems and rare in others, the examples of IP surveys in two quite different geological environments at McLaughlin and Rhyolite Creek demonstrate that the technique has useful application where knowledge of the geology is sufficient to allow reliable interpretation of anomalies (Irvine & Smith, 1990).

f) SEISMIC REFLECTION

There are no known cases of applying modern seismic reflection methods to delineate the quartz-adularia alteration zones of epithermal systems, although in low-density host rocks, the density contrast (and presumably velocity contrast) compared to surrounding argillic-propylitic altered rock may make such zones an attractive seismic reflection target (Allis, 1990).

g) EMPIRICAL MODELS AND GEOPHYSICAL RESPONSES

Figure 59 shows the elements of a high-level precious-metal deposit based on the empirical models of Buchanan (1981) and Silberman & Berger (1985). Table 17 lists the anticipated responses from various geophysical methods for surveys conducted at four different erosion levels indicated on the model. It is important to note the limitations of this simplistic portrayal: (i) a wide diversity of geological characteristics cannot be represented in a single model, and (ii) geophysical responses may change with even minor variations in the geological model. Depending principally on the level of erosion, the geophysical signature of the target illustrated by Figure 59 varies from a resistivity low with associated magnetic and gravity lows, to a resistivity high with a magnetic low, but no distinct gravity expression. Clearly, exploration at levels B and C are situations where geophysical methods can contribute significantly, along with geological and geochemical input (Irvine & Smith, 1990).

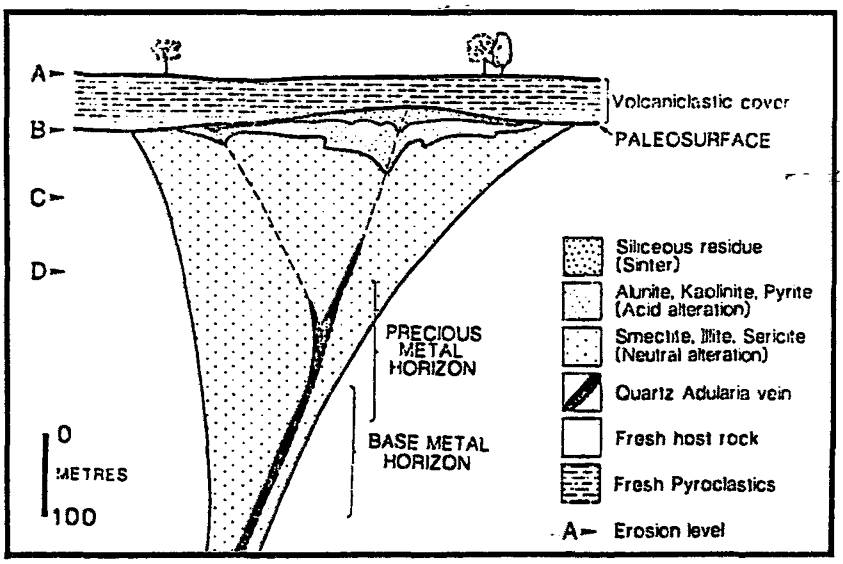


Figure 59. Diagram of a hypothetical epithermal gold deposit, modified after Buchanan (1981) and Silberman & Berger (1985), to illustrate the effect of erosion level on geophysical responses. The anticipated responses are listed in Table 17 (after Irvine & Smith, 1990).

Table 17: Anticipated responses from various geophysical methods (after Irvine & Smith, 1990).

Erosion level of prospect	Gravity	Magnetic	Gamma-Ray Spectrometer	E.M.	CSAMT	IP	Resistivity
A-Blind exploration through young volcanics	Weak broad low	Weak broad low	No response	Nil or weak response	Weak broad conductor	Nil or weak response	Nil or weak conductor
B-Exploration at hydro-thermal palaeo-surface	Strong broad low	Strong broad low	Strong potassium anomaly *	Strong broad conductor	Strong broad conductor *	Moderate broad response*	Strong broad conductor *
C-Target is non-outcropping vein system at 50-150m depth	Weak sharp low (requires close stations)	Sharp low	Strong potassium anomaly	Distinct conductor	Strong conductor within which narrow resistor may be resolved	Moderate response	Strong conductor within which narrow resistor is unlikely to be resolved
D-Survey over outcropping vein system	Unlikely to be detected	Very sharp low	Narrow potassium anomaly	Narrow multiple conductors	Narrow strong resistor flanked by strong conductors	Weak narrow response	Narrow strong resistor flanked by strong conductors

* Results may be strongly affected when a sinter is present at or near the present topographic surface.

h) RESISTIVITY MODEL

Figure 60 depicts a resistivity model for the application of geophysical electrical methods to epithermal gold deposits. In the first instance, it is relatively easy to detect the stockwork-type vein systems, owing to their large resistivity contrasts and large scale, as seen in the Hikiji deposit, Japan. In the second target type, it is also relatively easy to detect the outline of the argillic zone, as it usually shows low resistivity, and is large in scale. If depth penetration is required, as in the third target type, the Schlumberger method is harder to set up (Figure 61). In this case, other methods, such as CSAMT or TEM are more efficient and economical (Nishikawa, 1992).

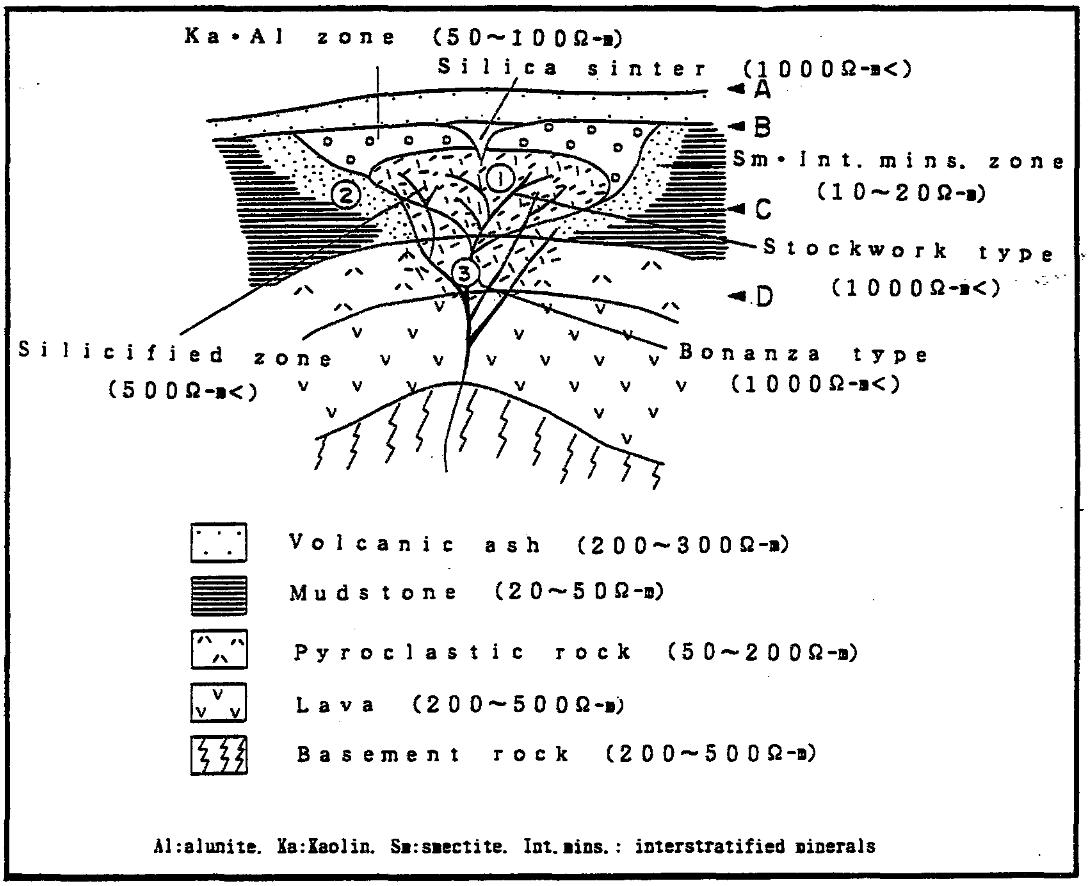


Figure 60. A resistivity model for an epithermal gold deposit (after Readdy, 1988).

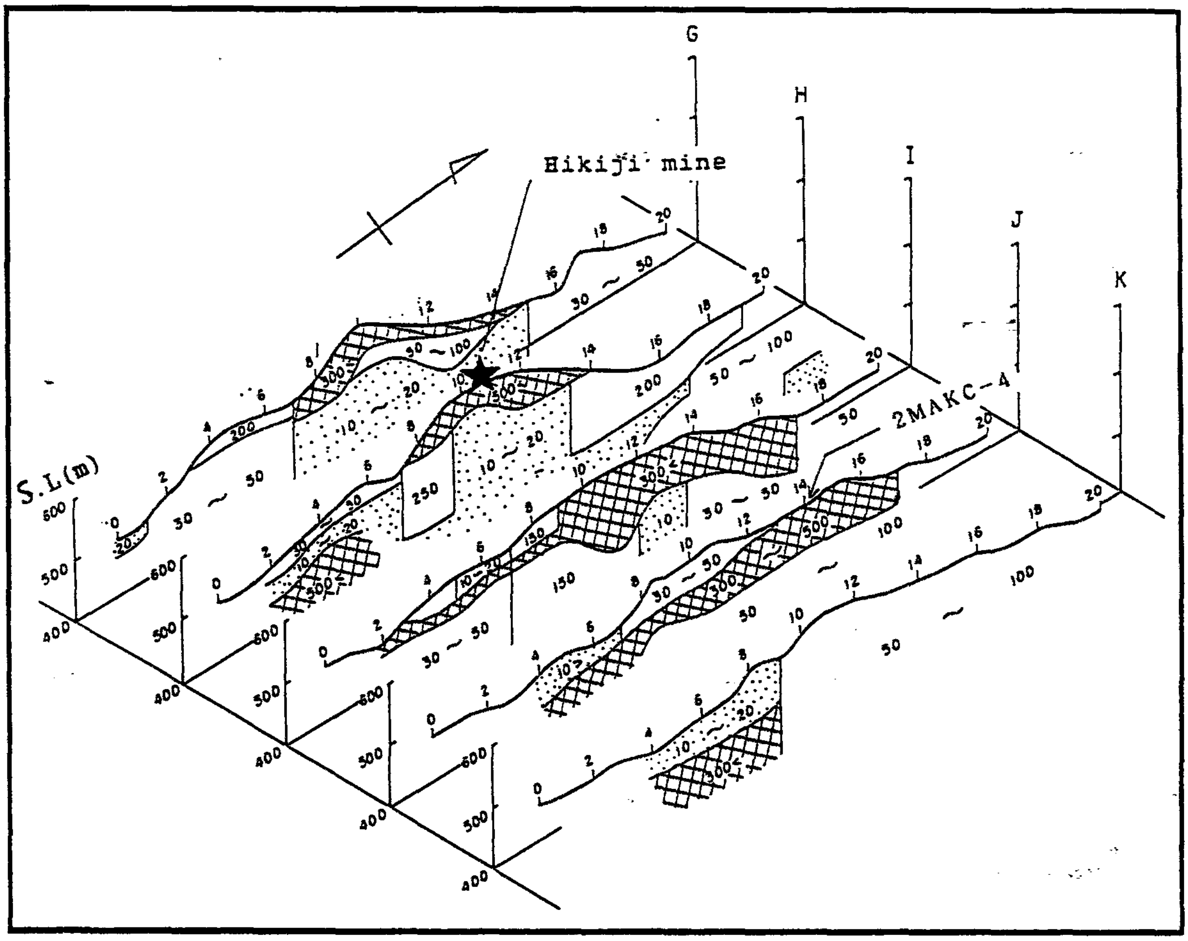


Figure 61. Resistivity section maps from the Schlumberger survey of the Hikiji area (after Readdy, 1988).

i) MASKED RESPONSES

In applying geophysical methods to epithermal gold exploration, local factors may reduce the effectiveness of a particular technique, or generate an ambiguous data set. For example: (i) Radiation from potassic minerals may be masked by young ash units or recent sediments; (ii) There are many electrically conductive features other than alteration zones; (iii) Unweathered, unfractured rock may be as resistive as quartz vein material; (iv) Highly conductive alteration zones or highly resistive siliceous sinters at surface may make mapping of deeper silicified zones impractical; (v) Pyrite is commonly widespread in regions of hydrothermal alteration and/or mineralisation, reducing the effectiveness of induced polarisation surveys; *etc.* Consequently, successful application of geophysical methods requires both a clear appreciation of the survey objectives and a good understanding of the local geology (Irvine & Smith, 1990).

j) SUMMARY OF GEOPHYSICAL TECHNIQUES

Table 18 summarises some of the distinctive physical features of the epithermal environment, and the geophysical techniques that can be used to detect them (modified after Irvine & Smith, 1990).

Table 18: Summary of the distinctive physical features of the epithermal environment and the geophysical techniques that can be used to detect them (after Irvine & Smith, 1990).

Physical feature	Geophysical approach
Magnetite destruction	Ground and airborne magnetics Magnetic susceptibility measurements
Enhanced potassium content	Airborne or ground spectrometer surveys
Increased or reduced density	Gravity surveys
Clay alteration (reduced resistivity)	Ground and airborne EM surveys Ground resistivity surveys Magnetotelluric surveys
High silica content (increased resistivity)	High resolution resistivity Controlled source audiomagnetotellurics Piezoelectric surveys
High sulphide content	Induced polarisation Electromagnetic surveys
Major structures	Airborne magnetics Gravity Magnetotellurics
Deep intrusives	Airborne magnetics Gravity Magnetotellurics

3.2.4 CONCEPTUAL MODELS

a) DATA SYNTHESIS

In order for all acquired reconnaissance and follow-up exploration data to be effectively utilised in interpreting the potential for and location of epithermal mineralisation, reconstruction of the geological setting, topography and palaeohydrology of the prospective district must be combined with available information on alteration mineralogy and zoning, geochemical anomalies and geophysical results (Irvine & Smith, 1990).

b) COMPLEX ZONING VARIATIONS

Conceptual models of epithermal ore-deposit types are of necessity generalisations of the geometry, alteration, structural controls, ore and gangue mineralogy, and trace-element geochemistry of these deposit types. Variations in alteration patterns due to duration, structural complexity, effects of host-rock properties such as permeability and chemical reactivity, the overprinting of features by changes in the position of active vents, elevation changes in the water table, and changes in the levels of boiling make the actual zoning relationship in deposits very complex (Silberman & Berger, 1985).

c) SECTOR COLLAPSE

An additional consideration is that in volcano-plutonic arcs subject to rapid uplift under pluvial, tropical conditions, such as the well-mineralised arcs of the Western Pacific region, seismically induced landsliding is a major process of landform degradation (Löffler, 1977; in Sillitoe, 1994). Sector collapse appears to be a normal stage in the evolution of large, composite volcanoes, especially those >2km high and having steep upper slopes due to more felsic, andesitic to dacitic compositions (Siebert, 1984; Francis & Wells, 1988; in Sillitoe, 1994). During caldera collapse, the entire system is likely to sink and the consequent hydrothermal perturbation is difficult to predict but could be relatively passive. In contrast, when sector collapse occurs, the hydrologic regime is certainly catastrophically disturbed. Fluid pressures will drop, possibly from lithostatic to hydrostatic, with the resulting production of widespread hydrothermal (phreatic) fracturing and brecciation, as at Ladolam (Moyle *et al.*, 1990; in Sillitoe, 1994). There is likely to be a concomitant major ingress of meteoric water causing dilution of the exsolving magmatic fluids, as documented for transformation from porphyry-type to epithermal conditions (Gustafson & Hunt, 1975; in Sillitoe, 1994). Ocean water may also gain access to some systems in Island-arc settings, as inferred at Ladolam (Moyle *et al.*, 1990; in Sillitoe, 1994). Boiling of deeply circulated meteoric or ocean water-dominated fluids, induced by reductions in confining pressure, is an effective means of precipitating Au in the epithermal environment (Hedenquist, 1987; in Sillitoe, 1994) and seems to have been instrumental in accumulation of the giant Au resource at Ladolam (Figure 62).

mineralisation is concealed at much shallower depths than otherwise might be anticipated (Figure 62). Field evidence for telescoping may be provided by either (i) a porphyry intrusion and/or late magmatic, vitreous quartz veinlet stockworks (generated originally as part of a K-silicate assemblage) overprinted by advanced argillic alteration and HS sulphide assemblages or (ii) preservation of volcano remnants (*e.g.* Marte and Ladolam) (Sillitoe, 1994).

e) CONCEALED EXTENSIONS

Thus in districts where one style of mineralisation is found, evidence should be sought for the presence of the other in an appropriate spatial relation (Sillitoe, 1991; in Arribas *et al.*, 1995). Epithermal and subvolcanic styles of mineralisation may be juxtaposed in highly telescoped systems, and either one (*e.g.* Marte, Chile: Vila *et al.*, 1991; Ladolam, Papua New Guinea: Moyle *et al.*, 1990; in Sillitoe, 1993) both (*e.g.* Lepanto, Philippines: Garcia, 1991; in Sillitoe, 1993), or neither (*e.g.* Guinaoang, Philippines: Sillitoe & Angeles, 1985; in Sillitoe, 1993) may constitute ore.

f) FLEXIBLE APPROACH

Appreciation of the broad spectrum of epithermal deposit styles and geometries is fundamental to efficient prospect evaluation (White & Hedenquist, 1990). Deposit geometry is the principal factor influencing bulk mineability. The explorationist must be aware of the variety of structural, hydrothermal and lithological ways of focusing hydrothermal fluid flow and, often using only sparse evidence, be able to predict concealed mineralisation styles (Sillitoe, 1993). The most generalised and yet potentially most broadly important concept is that during the early reconnaissance stages in a "new" exploration province, rigid adherence to a single geologic model should be avoided. Few, if any geologic terrains, should be excluded from preliminary examination and reconnaissance sampling (Larson & Erler, 1993).

3.3 ISSUES RELATING TO THE IDENTIFICATION OF PROSPECTIVE EPITHERMAL DEPOSIT-HOSTING REGIONS

While the selection of the overall region for an exploration programme is a corporate decision, influenced by mainly non-geological factors, the next step in narrowing down can be the integration of existing data (geological & geochemical) into a "prospectors map", highlighting areas of potential (Tischler, 1990).

The approach adopted towards exploration in any area depends to a large extent on the nature of the available data. In the broadest context, prospective areas for exploration must include: (i) evidence for a relatively near-surface heat source; (ii) the availability of large volumes of meteoric water; (iii) fracturing suitable for the movement of large volumes of meteoric water over extended periods of time; (iv) a host rock of sufficiently low permeability to allow the focusing of hot fluids within specific fracture sets; and (v) evidence for the transportion and precipitation of metals having occurred. The coexistence of more than one of the above criteria significantly increases the likelihood that a discovery will host potentially economic mineralisation, rather than the lesser, yet more likely result, of a barren hydrothermal system (Swindell, 1986).

To consider regions of the earths surface under the following broad topics is a suggested approach towards establishing or identifying areas which potentially might host epithermal mineralisation:

3.3.1 IGNEOUS PROVINCES (AND ASSOCIATED ALTERATION)

Igneous Provinces associated with convergent settings present attractive targets when establishing prospective epithermal gold-hosting areas on the surface of the Earth. The extent of the igneous province should first be determined; this includes all the igneous rocks (volcanic and plutonic) related to the igneous phase. Calc-alkaline to alkaline provinces are most prospective (White & Hedenquist, 1990), since these magmas are characteristic of the early stages of extension in arcs and back arcs of convergent plate margins (Smellie, 1994; in Elston, 1994). Satellite imagery, air photo interpretation and air-borne geophysical surveys will locate intrusions and extrusive centres in the sialic portions of convergent plate boundaries; such environs are inherently prospective (Tischler & Keyte, 1988; in Tischler, 1990). Regional controls on their ore potential include the level of erosion (and, hence, age) and geological features in the basement that focus the ascent of fluids (Berger & Henley, 1988). The correlation between regional tectonism and gold prospectivity in the Coromandel Peninsula for example, suggests that the best targets for gold exploration in general may be those fossil subduction settings, where volcanic activity has been associated with the fracture and collapse of basement rocks, thereby allowing the penetration of late-stage magmatic fluids into upper levels of the crust (Hodder, 1987). The pervasive alteration accompanying this process is obviously a prime target in identifying prospective regions.

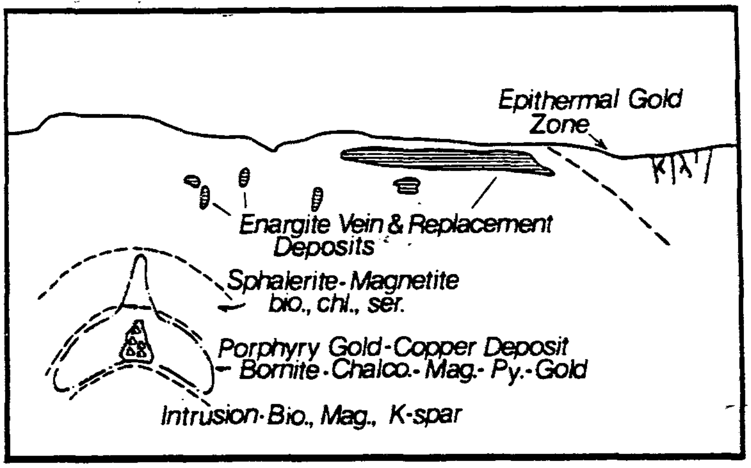


Figure 64. Schematic cross-section showing location of gold-enriched deposits in the Lepanto district, Phillipines. After Lowell (1988; in Jones, 1992).

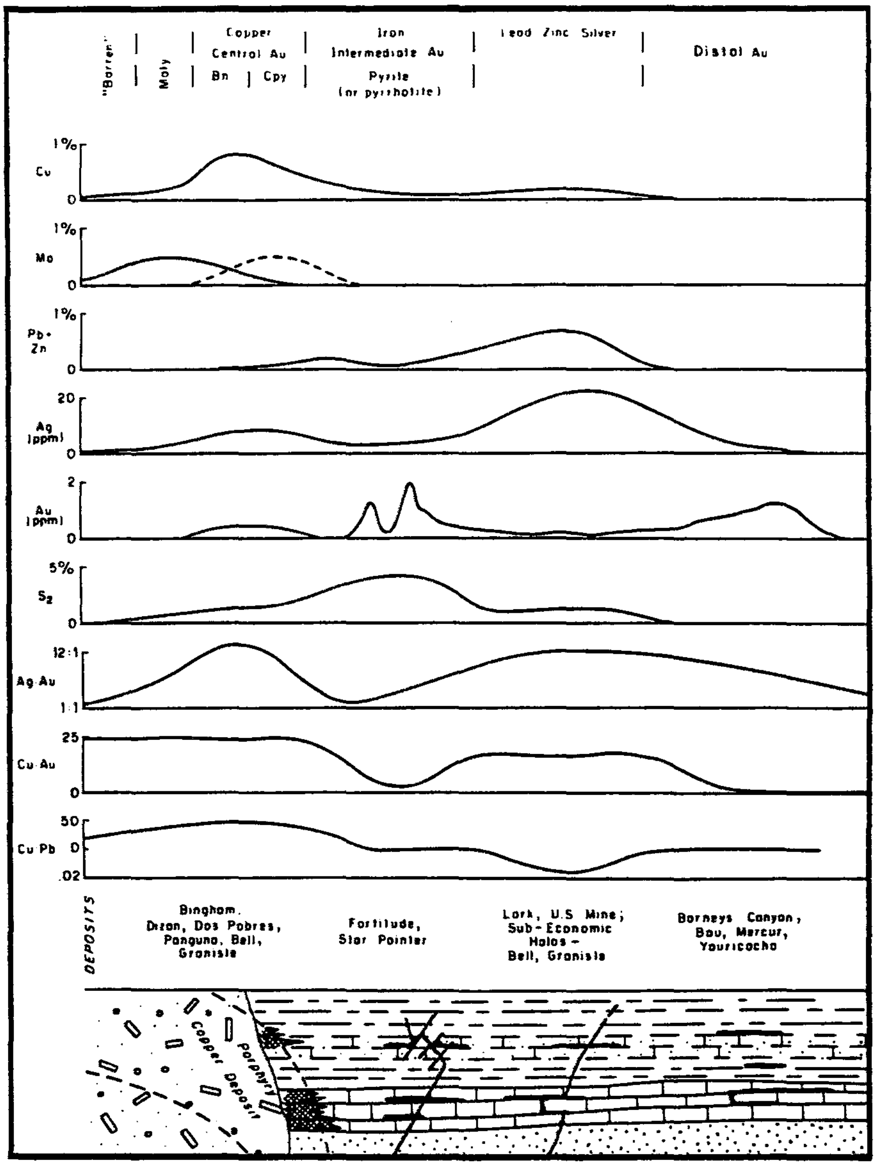


Figure 65. Generalised zoning model for gold-enriched porphyry copper systems showing changes in metal concentrations and metal ratios with distance from the source intrusion. All element concentrations are in ppm, unless "%" shown. Ag:Au is ppm Ag/ppm Au; Cu:Au is %Cu/ppmAu; and Cu:Pu is %Cu/%Pb. Distal gold and intermediate epithermal gold are clearly indicated by extremely low Cu:Au ratios (after Jones, 1992).

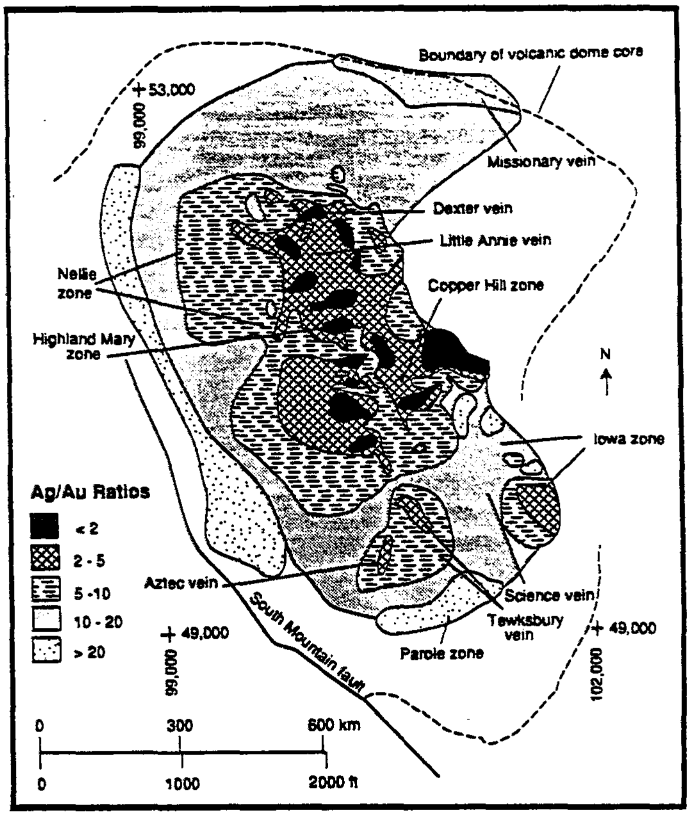


Figure 66. Map of the Summitville deposit showing Ag/Au ratios. Summitville has a low Ag/Au ratio center and much higher Ag/Au ratios on the deposit periphery. Data are based on analyses of exploration drill holes, ratios are independent of elevations between approximately 3 750 and 3 500m (after Gray & Coolbaugh, 1994).

3.3.3 STRUCTURAL SITES

The most favourable prospective setting in any district includes the intersection of a predicted gold zone with major pre-mineral faults (fluid conduits), areas of high fracture density, such as folded strata (permeability-porosity), and favourable host- and cap-rocks (*e.g.* limestone beneath diorite sills) (Jones, 1992).

Structures which penetrate deeply into the Earth's crust are commonly associated with gold deposits, as they focus the flow of mineralised fluids and localise deposition. Many deposits lie close to major regional faults or to splays off them. Thrust faults are favourable hosts where they have associated crush zones which enhance permeability. Caldera settings, are also common for epithermal deposits (such as at Lihir Island, Papua New Guinea), and caldera collapse may provide favourable sites in marginal structures (as at the Emperor Mine, Fiji) (Irvine & Smith, 1990).

Mineral exploration within the caldera environment has focused primarily on ring and radial fractures. Such fractures, related to caldera collapse, resurgence, and subsequent rifting may serve as conduits for mineralising solutions in later hydrothermal episodes, unrelated to caldera-forming events (Elston, 1994). Potential for mineralisation is highest where caldera-related fractures are intersected and reactivated by younger regional faults (Figure 27) (Rytuba, 1994). Only regional mapping can document calderas tens of kilometres in diameter. Small-scale aerial photographs or satellite images can help (Elston, 1994).

3.3.4 HIGHLY POTASSIC VOLCANICS

With the phantastic discovery of Lihir, attention focused onto highly potassic volcanics. They appear to be linked to magma generation within very deep portions of Benioff Zones (dying subduction). The island of Lipari (Aeolain Island Arc/Mediterranean) is similarly potassic, exhibits similar argillic alteration, siliceous breccias and banded chalcedonic sinters. It would seem to be a marvellous target for exploration (Tischler, 1990).

3.3.5 REMOTE-SENSING CAPABILITIES

Advances in Thermal-Infrared-Multispectral-Scanning (TIMS), which collects data in 6 channels in the mid-infrared atmospheric window (8-14 micrometers), permits rock-forming minerals such as carbonates and silicates to be remotely identified since they display characteristic emission features that allow discrimination for mapping. Colour composite images using standard decorrelation image-analysis techniques (principal components and Munsell) and both visual inspection and digital mapping can be used to compare the images to published geologic and alteration maps. Integration of information derived from the thermal infrared images with other geological, geophysical, and geochemical data can add to the overall geologic knowledge of an area and facilitate rapid evaluation of prospects (Kruse *et al.*, 1988).

Thus, during the process of identifying prospective epithermal environments, it is clearly necessary to first establish the nature of the geological environment at the time of mineralisation. The hydrological conditions that probably prevailed at that time can subsequently be assessed, and consequently the likely forms of and controls on mineralisation. Favourable geological environments of any stage should be considered prospective (White *et al.*, 1995).

3.4 RECONNAISSANCE CONSIDERATIONS IN AN EPITHERMAL TERRAIN

3.4.1 INTRODUCTION

Epithermal deposits, whether concealed within the crust or exposed at surface, occur within an extremely complicated and dynamic geological environment. Genetic processes in individual deposits take place in an exceptionally complex manner, so much so that the effects of hydrothermal alteration, due to a variety of physical and chemical processes, differ from prospect to prospect. Factors which affect alteration zoning patterns include amongst others: i) The geological age of the occurrence; ii) The duration of exposure to hydrothermal/epithermal fluids; iii) The degree of exposure and extent of erosion; iv) The nature of the mineralising hydrothermal fluid; v) The lithology of the host rock (degree of chemical reactivity); vi) Variations in the elevation of the watertable and the zone of boiling; vii) Variation in and extent of permeability; viii) The degree of structural complexity; ix) Variations in the positions of active vents and other hydrological conduits; and x) Vertical and lateral effects of temperature variation.

Additional factors to be considered in designing an exploration strategy include the scale at which the prospecting excercise is to be conducted, the degree of accessibility to the designated target area, the nature and density of the vegetation, the depth of the soil or regolith blanket, the climate, including the extent of precipitation and its seasonal variations, the degree of consolidation of younger lithological cover, *etc.*

Considering the above fundamental physicochemical variations, it is easy to appreciate that it is well-nigh an impossible task to develop a standard approach towards exploration for epithermal precious metal deposits. There are however a number of exploration procedures which frequently have more success than others and are therefore more regularly utilised in achieving the overall objective of reconnaissance exploration. That objective in epithermal terrains is to locate prospective regions hosting major regional (and related) structures, which at the same time reveal zones of pervasive alteration displaying favourable heat-flow characteristics - these should be preferably associated with intrusions into island arc or continental margin crust. Such areas are considered to be highly prospective but should not have been too deeply eroded and ideally should yield anomalous geochemical responses to reconnaissance sampling.

A number of exploration concepts relating specifically to epithermal deposits have been discussed in Section 3.2 above. Some of these do find application in the reconnaissance domain as well. Additional options available to the explorationist are discussed below. These are presented in such a manner as to generate a number of 'thought starters' concerning techniques available and factors to consider in establishing an overall approach towards reconnaissance in the designated target area. The data below should be reviewed in conjunction with the aspects presented in Section 3.2.

3.4.2 GEOPHYSICS (ALTERATION, MINERALOGY AND STRUCTURAL SETTING)

a) DETECTION OF ALTERATION

The passage of hot, saline, reactive fluids through volcanic rocks causes pronounced changes in their physical properties (Allis, 1990). The resulting contrasts in physical properties aid the delineation of hydrothermally altered zones by geophysical methods. Hydrothermal alteration in epithermal systems results in the destruction of magnetite, thus reducing the magnetic susceptibility effectively to zero, and also eliminating remanent magnetisation (both normal and reversed). In the vicinity of mineral occurrences there are thus large areas of quiet magnetic response, constituting a distinctive geophysical signature (Webster *et al.*, 1989). High temperature hydrothermal solutions also usually introduce potassium (including the useful indicator ^{40}K component, detectable by radiometric surveys) in the form of minerals such as alunite, commonly widespread in the upper levels of epithermal alteration systems, and adularia which is normally found within the zone of maximum gold deposition (Irvine & Smith, 1990). Airborne gamma-ray spectroscopy thus potentially provides a relatively cheap tool for reconnaissance over large areas of country-side, particularly where access is difficult (Killeen, 1979), since the radiometric data also allow extrapolation of poorly outcropping geology and provide a cost-effective mapping technique, even at exploration scales (*e.g.* 1:25,000) (Webster *et al.*, 1989).

A distinctive characteristic of hydrothermal clay minerals containing aluminium (*e.g.* montmorillonite, pyrophyllite, kaolinite, alunite) is an intense absorption of solar radiation around the 2.2 mm wavelength (Hunt, 1979; in Allis, 1990). This has led to the potential application of airborne and satellite reflectance spectroscopy methods as an epithermal exploration method. This topic is reviewed by Watson (1985; in Allis, 1990).

Aerial photographs, satellite images, airborne magnetic, radiometric and resistivity surveys, and airborne remote sensing techniques, can thus all assist in locating areas of hydrothermal alteration (Allis, 1990), which is an indirect approach to exploring for epithermal deposits (White & Hedenquist, 1990). Such surveys also provide basic geological information (Henley, 1991).

b) DETECTION OF STRUCTURE

Having defined regions likely to have been subjected to significant episodes of heat-flow, and which have not been too deeply eroded, structure should be the next regional guideline considered. Major structural zones may be recognised on a variety of regional data sets, from geological maps to airborne gravity, magnetic and radiometric surveys and satellite images. Other favourable structures such as those that occur around calderas may also be located from these data sets. The recognition of regional structures is also aided by satellite thematic mapper and side-looking aerial radar (SLAR) imagery (Henley, 1991). The distribution of known mines and prospects may directly indicate structures, as well as indicating favourability of other structures. On the regional scale only structural zones should be distinguished, as the actual site of mineralisation is commonly on a subsidiary structure within the structural zone, rather than on a major regional fault (White & Hedenquist, 1990). Depending on correct flight line orientation, airborne surveys can locate major regional and related structures which may provide a fundamental control on the location of ore districts (Henley, 1991). Under the right conditions, structural subtleties like lineaments, joints, and faults can be recognised from SLAR imagery and corroborated by regional field data (Kemp & White, pers.comm., 1987; in Simmons & Browne, 1990).

Zones of extensive rock alteration may in some cases be detected through gravity survey (Locke & De Ronde, 1987; in Henley, 1991). Additionally, in low-density host rocks, gravity highs of several mgal may occur over areas of mineral precipitation. (These anomalies may vary from hundreds of metres in characteristic length if the precipitation is focused on a fault zone, to anomalies similar in size to the entire upflow zone of the original geothermal system (5-10km^2); Allis, 1990).

c) STRUCTURAL DOMAINS

On a regional scale (*e.g.* 1:100,000) magnetic data reflect the regional tectonics and divide the area into domains for the application of different genetic models (Webster *et al.*, 1989).

Satellite imagery, detailed air photo interpretation (which can detect late movements on faults beneath cover formations), and airborne geophysical surveys will locate intrusions into island arc crust; such environments are inherently prospective (Tischler, 1990). In low-density host rocks, gravity highs of several mgal may occur over areas of mineral precipitation. These anomalies may vary from hundreds of metres in characteristic length if the precipitation is focused on a fault zone, to anomalies similar in size to the entire upflow zone of an original geothermal system (5-10km^2), (Allis, 1990). For exploration under younger regolith cover, high resolution aeromagnetics and seismic reflection data (Johnson & Henderson, 1991; in Henley & Etheridge, 1995) provide an additional, powerful, targeting tool.

d) STRUCTURAL ANALYSIS

It is worth noting that structural analysis by detailed airphoto interpretation, combined with alteration mapping by high resolution electrical resistivity surveying, enables (with close geological control), synthesis of structural and alteration characteristics of the Togi geothermal system. Geological modelling and drill target selection follow (Smith & Corbett, 1990).

e) GRAVITY/RESISTIVITY COMBINATION

A combination of gravity with resistivity surveys can be extremely rewarding. For example, this combination successfully contributed to the discovery of the Hishikari deposit in Japan. A very low resistivity anomaly, due to an argillic alteration zone, was measured directly over the Hishikari ore zone (Figure 67). An airborne electromagnetic (EM) survey was also conducted, resulting in the same resistivity anomaly. A gravity survey was then carried out in the same area and a high-gravity zone was measured above the orebody. These data led to the initiation of the first drilling survey in the Hishikari district (Nishikawa, 1992). Another example of their successful combined application is in the Hokusatsu volcanic region of Kyushu. Gravity and resistivity anomalies are important in target selection since subsurface structures, associated with resistivity lows, can be recognised by gravity survey. Uplifted basement blocks represented by gravity highs are favourable to mineralisation, probably because they focused fluid flow (Irvine & Smith, 1990).

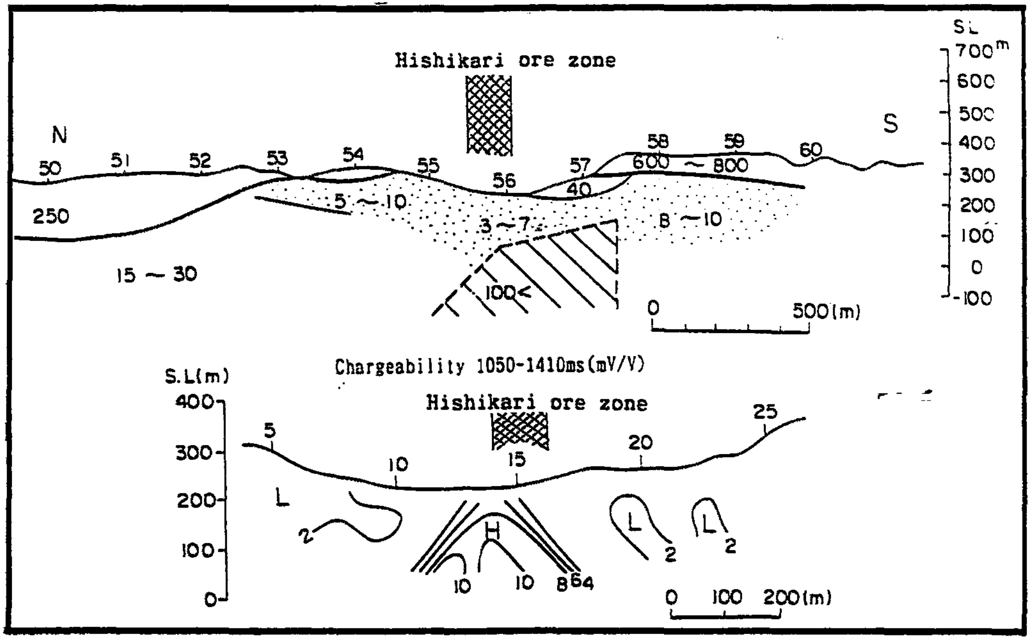


Figure 67. Resistivity and chargeability section maps of the Hishikari area (after Nishikawa, 1992).

f) AIRBORNE MAGNETIC/RADIOMETRIC COMBINATION

The most useful geophysical tool for delineating major structures and alteration systems continues to be low-level, detailed, combined magnetic and radiometric airborne surveys (Irvine & Smith, 1990). In the vicinity of Charters Towers, Northeastern Queensland for example (Table 19), high resolution airborne magnetic and radiometric data have been shown to optimise exploration for epithermal gold deposits (Webster *et al.*, 1989).

Table 19: Geophysical patterns of Gold deposits in North Queensland (after Webster *et al.*, 1989).

STYLE	TYPE EXAMPLE	SIGNATURE	AIRBORNE MAP RECOGNITION
PLUTONIC	CHARTERS TOWERS (Qld.)	- Deep seated structure - Broad pattern of intrusive	- Alignment of contours - Broad circular/ elliptical contour pattern
PORPHYRY RELATED VEIN	RAVENSWOOD FAR FANNING (Qld.)	- Deep (and shallow) structure - Broad pattern of intrusive	- Alignment of contours - Offset of trends - Sharp circular /elliptical contour pattern
PORPHYRY RELATED CALDERA (DIATREME BRECCIA)	PEAK HILL (NSW) MT. LEYSHON (Qld.)	- Intersecting structures - Porphyry pattern ... magnetic aureole ... negative anomaly - Potassium enrichment	- Offset of trends - Circular/elliptical contours with low centre - Circular/elliptical countours with "strong" "negative" core - Radiometric aureole
EPITHERMAL VEIN AND DISSEMINATED	TEMORA (NSW) PAJINGO (Qld.)	- Magnetite destruction - Shallow structure/ offsets - Potassium enrichment	- Broad "weak" magnetic low (circular pattern) - Sharp, elongated countours with offsets - Radiometric aureole

3.4.3 GEOCHEMISTRY (AND GEOLOGICAL MAPPING)

a) REGIONAL GEOCHEMISTRY

Regional geochemistry, combined with geological mapping of alteration in bedrock or float material, is of prime importance in any reconnaissance survey for epithermal mineralisation. Geochemical stream sediment and soil surveys are in general the reconnaissance methods most frequently utilised. Orientation surveys indicate the appropriate sampling parameters to be applied in conducting the survey, and will demonstrate which soil horizon in each particular area is the most effective concentrator of gold. The bulk-cyanide leach gold assay technique has developed into a qualitative and inexpensive tool for regional gold surveys. Arsenic, antimony and mercury are covariant with gold and may be used as additional pathfinders (Silberman & Berger, 1985) and base metals and copper may be useful for some deposit styles (Henley, 1991).

Two approaches may be taken in regional exploration in rugged terrains where pre-existing knowledge of geology is limited. The first approach includes an initial low-density sampling phase taking large stream samples for treatment by the bulk leach-extractable gold technique. Anomalies thus defined may then be followed up by a second phase of higher density sampling, with background areas being immediately discarded. The second approach does not include the initial low-density sampling, but rather goes directly to a high-density programme of stream sediment, pan concentrate, float and outcrop sampling. The second approach should be adopted since it is believed to be important that geologists should access all areas to map and sample rock float simultaneously with the geochemical survey to give greater geological control in the prioritisation of geochemical anomalies (Carlile *et al.*, 1990). Additionally, regional exploration using a high sampling density and including stream sediment, pan concentrate and float sample media can detect and characterise a variety of mineralisation styles. A very limited number of elements is sufficient and should include Au, Ag, Cu and As (Carlile *et al.*, 1990)

b) FRACTIONAL ANALYSES

Carlile *et al.* (1990), report that a regional exploration technique comprising fractional analyses of gold in stream sediments and pan concentrates, is able to detect mineralisation. Gold in the finer-size fractions of these media gives better discriminated anomalies and more repeatable results. Also, comparative fractional analyses for gold has shown that in both stream sediments and pan concentrates the finer fractions return higher and more repeatable results.

c) METAL RATIOING AND FLOAT MAPPING

Drainage geochemistry, from comparison with similar anomalies occurring elsewhere, may be used to predict the type of mineralisation that can be expected. In a number of cases described by Carlile *et al.*, (1990), Cu and Ag in stream sediments and soil geochemistry has enabled differentiation between mineralisation styles and thus guided ongoing exploration in areas of sparse outcrop. Metal ratioing is a useful data-smoothing technique for defining metal zones or portions of zones that are not detected by plotting raw geochemical data (compare Figures 68 and 69, Helecho system). Cu:Au (Figure 65) and Ag:Au (Figure 66) ratios are particularly useful in defining zones of relative gold enrichment (Jones, 1992). Float mapping and sampling during stream sediment surveys is an additional powerful tool giving geological control that can assist in characterising anomalies (Carlile *et al.*, 1990).

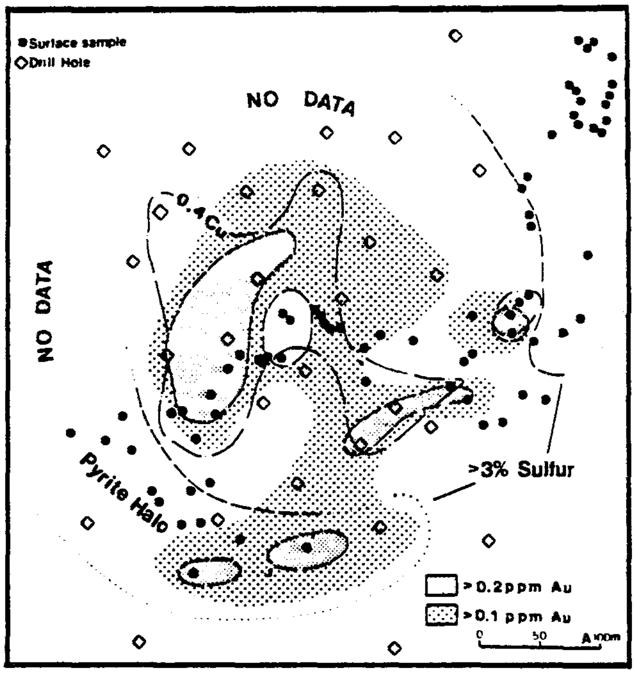


Figure 68. Distribution of gold values from drill core and surface samples at Helecho, Puerto Rico. Data from Cox (1985; after Jones, 1992). Abbreviations: gar=garnet, pyx=pyroxene, act=actinolite.

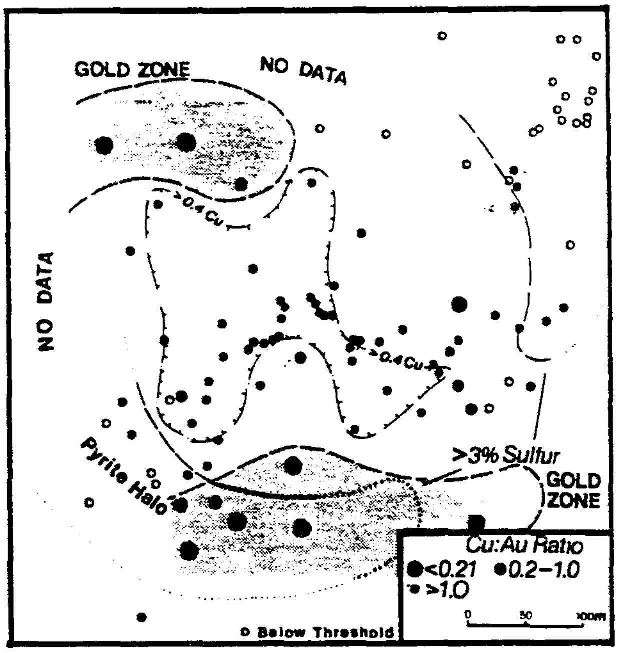


Figure 69. Distribution of copper to gold ratios for samples shown in Figure 68, Helecho, Puerto Rico. Data from Cox (1985; in Jones, 1992).

d) MANGROVE SEDIMENT RECONNAISSANCE

In the tropical weathering environment in the Western Pacific Islands, bedrock is generally not well exposed and laterite is developed over most bedrock to a depth of several tens of metres. Dense vegetation generally covers large areas of the islands, and poor access can make geochemical exploration surveys difficult to conduct. Geochemical media sampled in routine reconnaissance surveys include stream sediments, heavy-mineral concentrates from stream sediments, and mangrove sediment (Rytuba & Miller, 1990).

The low-energy environment characteristic of the intertidal zone within the mangrove and swamp domains along the coastlines of islands like Palau and Yap, for example, is conducive to trapping and retaining very fine-grained sediment. The sampling of the sediment has demonstrated that it is an effective medium for use in rapidly determining the precious-metal potential of land areas adjacent to the coast. Mangrove sediments could be sampled as a first step in an exploration programme for precious metals in these terrains (Rytuba & Miller, 1990).

e) GOLD PANNING

A relevant observation with respect to geochemistry comes from Tischler (1990), who concluded that the most effective tool in early stage exploration for gold is the panning dish, since gold will coarsen up significantly in tropical climates - even without alluvial transport (by the action of humic acids).

f) FIRST PASS EXPLORATION SUCCESS

First pass exploration can be considered successful in an area if alteration has been delineated and the presence of gold within heavy mineral concentrates, stream sediments - and above all - in rock float/outcrop has been confirmed. First pass exploration which outlines areas of alteration but fails to detect a consistent amount (even if very low) of gold, downgrades the potential of an area drastically (Tischler, 1990).

3.5 FOLLOW-UP CONSIDERATIONS REGARDING EPITHERMAL PROSPECTS

3.5.1 INTRODUCTION

Whether a potential ore fluid actually forms an ore deposit or not depends principally on two factors: focusing and deposition (White & Hedenquist, 1990).

Focusing of ore fluids occurs in zones of enhanced permeability. Palaeopermeable zones can generally be directly recognised by an increased density of mineralised fractures (typically veins), and the occurrence of hydraulic brecciation and hydrothermal eruption breccias. In other cases they may be recognised indirectly from vein mineralogy, and from the mineralogy and zoning of hydrothermal alteration products. These may also be detected using geophysical techniques (Irvine and Smith, 1990; in White & Hedenquist, 1990).

Deposition of gold in epithermal deposits can have several causes. These include boiling, fluid mixing, cooling, and wall-rock reaction, and evidence for each can be recognised from vein and alteration mineralogy. Characteristic textural and mineralogical evidence suggesting boiling can commonly be observed in the field, and consequently can provide simple practical guides in exploration (Table 20). Evidence for fluid mixing is seen in the alteration assemblage (notably mixed layer clays around the margins of vein systems, or an acid overprint near the top of the system). Both boiling and fluid mixing are likely to be enhanced in structural zones, which are at the same time effective sites of fluid focusing (White & Hedenquist, 1990).

Table 20: Interpreting obeservations in epithermal environments (after White & Hedenquist, 1990).

Observation	Inference
Vein Mineralogy/Texture	
chalcedony present	rapid cooling has occurred; may indicate boiling: deposition temperature between 190 - 100°C; can infer depth of less than 100m below water-table, assuming hydrostatic conditions.
Adularia present	boiling has occured, 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	palaeotempertures 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 steamheated waters marginal to the sytem, from acid sulphate, 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 800m.
alunite	conditions acid with high sulphate 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) temperatures 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-sulphidation deposits.

Structural studies are considered to be indispensable at the project area scale of exploration (White & Hedenquist, 1990). Typically, many structures are not mineralised, so it is necessary to distinguish the more prospective structures by examining the correspondence between structures, geochemistry, mineral occurrences and hydrothermal alteration. Favourable conjunctions of these features become the focus of prospect-scale exploration. Efficient prospect-scale exploration thus requires the careful integration of all available data, coupled with a good understanding of the processes that occur, and their likely effects. Possible topographic effects should also be considered, as these will influence the spatial distribution of conduits and hybrid fluids, and resultant mineralisation and alteration (White & Hedenquist, 1990).

3.5.2 GEOPHYSICAL APPROACH (ASSISTED BY MAPPING)

Ground geophysical techniques can play an important role in defining drill targets at the prospect scale. A number of different ground electrical and electromagnetic (EM) methods are currently applied in epithermal gold exploration, including resistivity, IP, CSAMT, VLF resistivity, and frequency and transient electromagnetics (FEM and TEM). Conscientiously acquired gravity surveys may assist in the delineation of major structures, basement highs and alteration zones (Irvine & Smith, 1990).

Hydrothermal alteration commonly lowers the electrical resistivity of most volcanic rocks by one or two orders of magnitude, from normal values of 50-250 ohm metres. This is due to the replacement of feldspar and pyroxene minerals by clay and zeolite minerals with relatively high cation exchange capacity. In contrast, portions of the alteration zone are typified by repeated silica veining or silica flooding (which generally hosts mineralisation), causing an increase in resistivity (typically 250-2000 ohm metres; Irvine & Smith, 1990). Mineralised epithermal vein systems are thus generally hosted within clay alteration zones (which are strongly conductive), enveloping or enclosing silicified zones (which are in contrast resistive).

These hydrothermal alteration effects commonly envelope mineralised veins and can provide a broad target by assisting in identifying more favourable areas (White & Hedenquist, 1990). The identification of such zoning is critical to the discovery of buried vein systems. In covered terrain high-resolution ground resistivity surveys facilitate the mapping of physical contrasts related to comparable alteration (Smith & Corbett, 1990).

More deeply buried conductive alteration systems can be defined by a combination of resistivity, electromagnetic and magnetotelluric methods. Resistive gold-bearing silicified zones and shallow quartz vein systems are commonly detectable by conventional galvanic or inductive resistivity techniques. In combination with IP surveys, they can delineate the areal distribution of the subsurface hydrothermal alteration related to mineralisation (Irvine & Smith, 1990). Gold mineralisation is also commonly accompanied by disseminated sulphides. In this case, IP too can be an effective exploration tool (Nishikawa, 1992).

Experience at Hishikari (Kawasaki *et al.*, 1986; in Smith & Corbett, 1990) and in New Zealand (Irvine & Smith, 1990) has demonstrated that the CSAMT method can locate concealed zones of clay-pyrite alteration, silicification and unaltered rock and can accurately delineate their contacts. CSAMT has the advantages of high spatial resolution, deep penetration (especially in relatively resistive areas) and good interpretability.

In the situation where gold mineralisation is accompanied by disseminated sulphides (which is commonly the case), IP can be an effective exploration tool. IP surveys are also quite effective in detecting bonanza-type deposits in the situation where the host rock is only weakly silicified (as is the case at the Hishikari deposit), (Nishikawa, 1992). As economic Au-Ag mineralisation is typically associated with silica deposition (*e.g.* Waihi, New Zealand, and Hishikari, Japan) identification of such resistors may be of prime importance in siting drill holes (Irvine & Smith, 1990). When it comes to detecting quartz veins containing gold, Sobolev *et al.* (1984; in Irvine & Smith, 1990) reported that piezoelectric surveys are routinely applied in the USSR for just that purpose. They described successful tests on large quartz lenses carrying ore-grade gold in Archaean rocks at the Giant Yellowknife Mine in Canada.

In many epithermal districts, recognition of the original geomorphology provides clues to the palaeohydrology which may then aid in developing exploration targets. Recognition of multiple alteration stages may be just as valuable. Similarly, mapping of breccia types provides a significant guide to mineralisation style and possible targets (Henley, 1991). Where outcrop is sufficient or pilot drilling has been completed, mapping of the alteration (backed by X-ray diffraction and petrography) is equally important, although the distinction of hypogene or supergene origin for kaolinite and alunite may often be impossible (Henley, 1991). This is especially so when complexities introduced by extensive lateral flow of fluids down hydraulic gradients must be factored into interpretations of alteration patterns in high-relief terrain. This is necessary since drilling in search of concealed precious-metal deposits in shallowly eroded systems should be guided by tentative reconstructions of subjacent palaeo-fluid-flow channelways (Sillitoe, 1993).

3.5.3 GEOCHEMICAL AND MINERALOGICAL APPROACH

Studies of alteration mineralogy and zoning may provide valuable insights into the hydrology of the system, and indicate possible sites of deposition; however, only geochemistry offers a direct approach to locating mineralisation (White & Hedenquist, 1990). At the prospect scale, geochemical surveys based primarily on gold (from samples recovered during pitting, trenching and drilling activities), are of great importance in locating ore zones (Henley, 1991), since the greater the dispersion of the potential ore fluid, the more widely dispersed are the geochemical effects of the fluids. Thus the geochemical response detected over an area of epithermal mineralisation depends both in extent and chemistry on the hydrology of the system, and its level of exposure. This latter point is particularly important in interpreting the anomalous level of some elements (*e.g.* Hg, Tl, As), as their concentrations can increase at least two orders of magnitude in the upper few hundreds of metres of a system (White & Hedenquist, 1990). Geochemical exploration using Hg, CO_2 and radon in soil gas is often quite effective in tracing buried fracture zones (Irvine & Smith, 1990), particularly in the situation where mineralisation has been buried below younger cover (*e.g.* regolith).

The isotopic composition of quartz veins provides the most clearly defined trends to guide future exploration. Investigation of fluid inclusion characteristics and stable isotope signatures can provide complementary data which in combination with more standard geochemical, geophysical, and geological information can provide site-specific targets for epithermal mineral concentrations (Masterson & Kyle, 1984). For example, in the Masupa Ria area, four main mineralised areas were defined in response to reconnaissance exploration. All of these were tested by drilling. Samples were selected for routine petrographic, X-ray diffraction and reconnaissance fluid inclusion studies. Petrographic and clay alteration studies were also carried out (Thompson *et al.*, 1994). Suggested additional analyses could have included those for trace-metal content of altered rock and oxygen isotopic composition of quartz veins. The results of these exploration programmes provided excellent descriptions of the geological setting and the nature of the mineralisation associated with each of the prospects (Thompson *et al.*, 1994), facilitating the generation of extremely detailed and accurate genetic models. Such models are of considerable assistance to the exploration geologist.

PART D: DISCUSSION

Recently improved understanding of epithermal deposits has been derived from studies of active geolthermal systems (*e.g.*, Henley and Ellis, 1983). In the geothermal environment, epithermal deposits form as the chemical composition of the fluid responds to changes in pressure and temperature, and to wall rock composition. In general the principles of these changes are well understood (Henley *et al.*, 1984), though we do not yet fully understand why some systems produce ore, and some, otherwise apparently identical, appear to be barren. The character and distribution of ore and wall-rock alteration are the most variable features of ore deposits, and are what makes each one unique. These aspects are dictated by the hydrology of the system, and whether the fluids are low-sulphidation (neutral pH and reduced) or high-sulphidation (acid pH and oxidised). The tectonic setting is influential in determining the hydrology of the system, and how it changes with time (White *et al.*, 1995).

The range of exploration methods available to explorationists is limited, and different methods vary in effectiveness in different situations. In most instances, unless they crop out at the surface, epithermal gold deposits of both HS- and LS-styles are difficult geophysical (and geochemical) exploration targets because they vary sharply in width and grade as no two deposits will occur in exactly the same three-dimensional array, or within the same hydrological environment. This in turn has a direct affect on the resultant alteration envelope. Successful exploration must rely on integration of a variety of exploration techniques, guided by an understanding of the characteristics of the deposits and the processes that form them. There are no simple formulae for success: what works best must be determined for each terrain and each prospect (White & Hedenquist, 1995). For example, Table 21 tabulates the exploration techniques utilised in the search for precious metals (not only of epithermal origin) in the Canadian Cordillera. Table 22, in turn, lists many of the methods available for exploration for epithermal gold deposits, and their use in different settings (see also Irvine & Smith, 1990; and Allis, 1990). The variability and subtlety of the geological, geochemical and geophysical signatures of epithermal systems usually means that no one exploration method can be relied on, and an integration of all types of data is essential (Allis, 1990).

Table 21: Exploration techniques utilised in the search for precious metals in the Canadian Cordillera (after Ericksen, 1988).

1.	GEOLOGY		- Site selection - Regional, local
2.	EXPLORATION		
	a.	Basic prospecting	- Especially along structural breaks
	b.	Air photography	- Especially for structure and alteration zones utilizing: Black and white mosaics Colour Satellite imagery (e.g. Landsat)
	c.	Geochemistry	- Mianly Au and Ag, other metals locally
			- Slits (conventional and heavy media separation)
			- Soils - including frost heaved talus (using a grid sampling)
			- Soils gas (unsuccessful due to climate)
			- On site Hg analyses
	d.	Geophysics	- EM with resistivity attachment (*e.g.* line spacing 25m with 12.5m stations)
			- IP with 800m reconnaissance type spacing - Multipole works well for large, low-grade targets
			- VLF mianly inconclusive, locally useful
			- Magnetometer for alteration zoning and skarn associations
	e.	Trenching	- Handblasting generally ineffective
			- Backhoe very effective to 5m
	f.	Drilling	- Diamond drilling with large diameter core and good recovery provides maximum geological and economic data. Percussion/reverse circulation ore cost effective.
	g.	Underground	- Ore reverse definition and bulk sampling for metallurgical testing.
3.	GEOLOGY		- Interpretation, Summary Reports
(4)	ENGINEERING/FEASIBILITY		

Table 22: Exploration methods applied to exploring different settings. Note: "Shallow" refers to a shallow level in the mineralised system, i.e. little or no erosion has occurred. "Deep" refers to a deep level in the mineralised system, i.e. it is deeply eroded (e.g. >400m) (after White *et al.,* 1995).

Setting method	Silicic depression	Andesitic stratovolcano	Cordilleran volcanism	Oceanic volcanic island
Geology	Recognition of veins and hydrothermal alteration patterns may directly locate mineralisation. Vein texture and mineralogy may indicate depth of formation. ← Geological understanding is the basis for interpreting all other data sets. →			
Geochemistry	Shallow: Low level anomalism may be widespread, unrelated to economic deposits; may not reflect most favourable sites at depth. Indicator elements important. Response commonly obscured by younger cover (e.g. ash deposits)	Shallow: Alteration commonly related to phase separated gases, so only volatile elements may be detected.	Shallow: Low level anomalism may be widespread, unrelated to economic deposits; may not reflect most favourable sites at depth. Indicator elements important.	
	← Deep: Anomalism related to mineralised structures →			
Geophysics				
1. Magnetics	Effectiveness depends on the magnetic properties of the rocks. In epithermal environments, prospect-scale surveys are typically ineffective. Regional-scale surveys may be a very effective aid in geological mapping, and may define areas of hydrothermal alteration (recognised by magnetite destruction). Structures may be traced from unaltered areas into areas of hydrothermal alteration.			
2. Induced polarisation	Shallow: Systems with abundant pyrite in hydrothermal altered cap rocks produce anomalies that are not related to gold mineralisation. May be useful on a regional scale, rather than prospect scale.			
	← Deep: Contrast in sulphide content may be a useful guide to mineralisation →			
3. Resistivity	← In active systems hot water will be detected as areas of low resistivity →			
	Shallow: High resistivity silicified caps within low resistivity clay alteration may obscure economically significant features.	Shallow: Low resistivity clay alteration blanket obscures economically significant features.	Shallow: High resistivity zones may represent economically significant veins or silicification.	Shallow: High resistivity silicified caps within low resistivity clay alteration may obscure economically significant features.
	Deep: Narrow high resistivity linear features reflect veins or silicificaion, typically surrounded by low resistivity alteration.		Deep: Narrow high resistivity linear features reflect veins or silicification.	Deep: Low resistivity represents fresh rock, veins or silicification; high resistivity may be associated with disseminated mineralisation.
4. Electromagnetics	← In active systems hot water will be detected as areas of high conductivity. →			
5. Radiometrics	Detect outcropping radiogenic sources. Variations in potassium content can be detected. Adularia and sericite in alteration zones can be recognised as positively anomalous potassium channel responses. Hydrothermally altered potassium-rich host rocks may show local potassium depletion.			
6. Gravity	Gravity surveys detect anomalies related to (1) changes in the density of host rocks due to alteration, (2) basement topography, and (3) intrusions.			
	Hydrothermal alteration causes a positive anomaly.	Hydrothermal alteration causes a negative anomaly.	Hydrothermal alteration may cause a negative or positive anomaly.	Hydrothermal alteration causes a negative anomaly.

A fairly standard approach which proves to be quite effective, is to incorporate all geophysical, geochemical and geological data into a model of the original hydrothermal system, and then to test the model by drilling. This is done in order to verify the identity of possible sets of anomalies consistent with the effects of high-temperature fluid flow, with palaeo-relief and hydraulic gradients reconstructed (Allis, 1990). Should the results of drilling substantiate the model and should circumstances warrant it, additional investigation would proceed. A word of warning however is proffered by White *et al.* (1995), who suggest that the results of exploration in the Southwest Pacific indicate that, while the principles of formation of epithermal deposits still apply, deposit models which have been developed in other tectonic/volcanic settings should not be rigidly applied in settings far removed. Explorationists should first consider the character of the geological environment at the time of mineralisation to try to predict the hydrological conditions that were prevalent during mineralisation. Once the likely volcanic environment and palaeorelief have been established, then the types, distribution and zoning of hydrothermal alteration, coupled with observations of deposit form, vein textures and mineralogy, will allow a judgement to be made on the level of system that is now exposed, and on the probable controls that localised mineralisation.

Possibly the most important conclusion regarding the geoscientists approach towards exploration is extended by White & Hedenquist (1990), who wilfully point out that exploration should be conducted so that any type of economically significant mineralisation will be identified, not only one model-type.

REFERENCES

Allen. K.D., and Hahn, G.A., 1994. Geology of the Sunbeam and Grouse Creek Gold-Silver Deposits, Yankee Fork Mining District, Eocene Challis Volcanic Field, Idaho: A Volcanic Dome- and Volcaniclastic-Hosted Epithermal System. Econ. Geol., 89 (8), 1964-1982.

Allis, R.G., 1990. Geophysical Anomalies over Epithermal Systems. J. Geochem. Expl., 36, 339-374.

Anderson, W.B., Antonio, M., Davis, B., Jones, G.F.P., Setterfield, T.N., and Tua, P., 1987. The Emperor Epithermal Gold Deposit, Vatukoula, Fiji. In: Pacific Rim Congress '87, 9-12.

Arribas, A., Jr., Hedenquist, J.W., Itaya, T., Okada, T., Concepci6n, R.A., and Garcia, J.S., Jr., 1995. Contemporaneous Formation of Adjacent Porphyry and Epithermal Cu-Au Deposits over 300 ka in Northern Luzon, Philippines. Geol., 23 (4), 337-340.

Berger, B.R., and Henley, R,W., 1988. Advances in the Understanding of Epithermal Gold-Silver Deposits, with Special Reference to the Western United States. In: The Geology of Gold Deposits - The Perspective in 1988. Econ. Geol., Mono. 6, 405-423.

Berger, B.R., and Bonham, H.F., 1990. Epithermal Gold-Silver Deposits in the Western United States: Time-Space Products of Evolving Plutonic, Volcanic and Tectonic Environments. J. Geochem. Expl., 36, 103-142.

Bodnar, R.J., Reynolds, T.J., and Kuehn, C.A., 1985. Fluid Inclusion Systematics in Epithermal Systems. In: Berger, B.R., and Bethke, P.M., eds., Geology and Geochemistry of Epithermal Systems. Soc. Econ. Geol., Reviews Econ. Geol., 2, 73-97.

Bogie, I., 1995. Geological Controls on the Formation of Gold Rich Porphyry-Type Deposits. Proceedings of the 1995 Pacrim Conference, 59-64.

Buchanan, L.J.,1981. Precious Metal Deposits associated with Volcanic Environments in the Southwest. In: Dickson, W.R.,and Payne, W.D., eds., Relations of Tectonics to Ore Deposits in the Southern Cordillera. Arizona Geol. Soc. Digest, 14, 237-262.

Camus, F., and Skewes, M.A., 1991. The Faride Epithermal Silver-Gold Deposit, Antofagasta Region, Chile. Econ. Geol., 86, 1222-1237.

Carlile, J.C., Digdowirogo, S., and Darius, K., 1990. Geological Setting, Characteristics and Regional Exploration for Gold in the Volcanic Arcs of North Sulawesi, Indonesia. J. Geochem. Expl., 35, 105-140.

Castor, S.B., and Weiss, S.I., 1992. Contrasting Styles of Epithermal Precious-Metal Mineralization in the Southwestern Nevada Volcanic Field, USA. Ore Geol. Rev., 7, 193-223.

Clarke, D.S., 1991. Contrasting Epithermal Gold Mineralization in Andesitic and Rhyolitic Terranes. In: PNG Geol., Expl. and Min. Conf., 40-48.

Clarke, D.S., and Govett, G.J.S., 1990. Southwest Pacific Epithermal Gold: A Rock-Geochemistry Perspective. In: Hedenquist, J.W., White, N.C., and Siddeley, G., eds., Epithermal Gold Mineralization of the Circum-Pacific: Geology, Geochemistry, Origin and Exploration. J. Geochem. Expl., 35, 225-240.

Cole, D.R., and Drummond, S.E., 1986. The Effect of Transport and Boiling on Ag/Au Ratios in Hydrothermal Solutions: A Preliminary Assessment and Possible Implications for the Formation of Epithermal Precious-Metal Ore Deposits. J. Geochem. Expl., 25, 45-79.

Drummond, S.E., and Ohmoto, H., 1985. Chemical Evolution and Mineral Deposition in Boiling Hydrothermal Systems. Econ. Geol., 80, 126-147.

Elston, W.E., 1994. Siliceous Volcanic Centers as Guides to Mineral Exploration: Review and Summary. Econ. Geol., 89, 1662-1686.

Ericksen, G., 1988. IUGS/UNESCO Deposit Modeling Workshop: Hydrothermal Systems in Volcanic Centers. Extended Abstracts: Revista Geologica De Chile, 15 (2), 1-20.

Field, C.W., and Fifarek, R.H., 1985. Light Stable-Isotope Systematics in the Epithermal Environment. In: Berger, B.R., and Bethke, P.M., eds., Geology and Geochemistry of Epithermal Systems. Soc. Econ. Geol., Reviews Econ. Geol., 2, 99-128.

Fournier, R.O., 1985. The Behaviour of Silica in Hydrothermal Solutions. In: Berger, B.R., and Bethke, P.M., eds., Geology and Geochemistry of Epithermal Systems. Soc. Econ. Geol., Reviews Econ. Geol., 2, 45-61.

Fournier, R.O., 1985. Carbonate Transport and Deposition in the Epithermal Environment. In: Berger, B.R., and Bethke, P.M., eds., Geology and Geochemistry of Epithermal Systems. Soc. Econ. Geol., Reviews Econ. Geol., 2, 63-72.

Giggenbach, W.F., 1992. Magma Degassing and Mineral Deposition in Hydrothermal Systems along the Convergent Plate Boundaries. Econ. Geol., 87, 1927-1944.

Glaser, M.L., and Keays, R.R., 1988. Genesis of Volcanogenic Epithermal Gold-Silver Mineralization, Budawang Rift, New South Wales. In: Bicentennial Gold '88, 268-270.

Golding, S.D., and Wilson, A.F., 1984. Stable Isotope Relationships in Epithermal Gold Deposits, Queensland. In: Gold-Mining, Metallurgy and Geology Regional Conference '84. Austr. Inst. Min. Metall., 427-435.

Gray, J.E., and Coolbaugh, M.F., 1994. Geology and Geochemistry of Summitville, Colorado: An Epithermal Acid Sulfate Deposit in a Volcanic Dome. Econ. Geol., 89, 1906-1923.

Gröpper, H., Calvo, M., Crespo, H., Bisso, C.R., Cuadra, W.A., Dunkerley, P.M., and Aguirre, E., 1991. The Epithermal Gold-Silver Deposit of Choquelimpie, Northern Chile. Econ. Geol., 86 (6), 1206-1221.

Hayba, D.O., Bethke, P.M., Heald, P., and Foley, N.K., 1985. Geologic, Mineralogic, and Geochemical Characteristics of Volcanic-Hosted Epithermal Precious-Metal Deposits. In: Berger, B.R., and Bethke, P.M., eds., Geology and Geochemistry of Epithermal Systems. Soc. Econ. Geol., Reviews Econ. Geol., 2, 129-167.

Heald, P., Foley, N.K., and Hayba, D.O., 1987. Comparative Anatomy of Volcanic-Hosted Epithermal Deposits: Acid-Sulfate and Adularia-Sericite Types. Econ. Geol., 82, 1-26.

Hedenquist, J.W., 1995. The Ascent of Magmatic Fluid: Discharge versus Mineralization. In: Thompson, J.F.H., ed., Magmas, Fluid and Ore Deposits. Mineral. Assoc. Canada, Shortcourse 23, 263-289.

Hedenquist, J.W., and Henley, R.W., 1985. The Importance of CO_2 on Freezing Point Measurements of Fluid Inclusions: Evidence from Active Geothermal Systems and Implications for Epithermal Ore Deposition. Econ. Geol., 80, 1379-1406.

Hedenquist, J.W., and Lowenstern, J.B., 1994. The Role of Magmas in the Formation of Hydrothermal Ore Deposits. Nature, 370, 519-527.

Hedenquist, J.W., Matsuhisa, Y., Izawa, E., White, N.C., Giggenbach, W.F., and Aoki, M., 1994. Geology, Geochemistry, and Origin of High-Sulfidation Cu-Au Mineralization in the Nansatsu District, Japan. Econ. Geol., 89, 1-30.

Hemley, J.J., Hostetler, P.B., Gude, A.J., and Mountjoy, W.T., 1969. Some Stability Relations of Alunite. Econ. Geol., 64, 599-612.

Henley, R.W., 1985. The Geothermal Framework for Epithermal Deposits. In: Berger, B.R., and Bethke, P.M., eds., Geology and Geochemistry of Epithermal Systems. Soc. Econ. Geol., Reviews Econ. Geol., 2, 1-24.

Henley, R.W., 1991. Epithermal Gold Deposits in Volcanic Terranes. In: Foster, R.P., ed., Gold Metallogeny and Exploration, 133-164.

Henley, R.W., and Ellis, A.J., 1983. Geothermal Systems Ancient and Modern: A Geochemical Review. Earth Sci. Reviews, 19, 1-50.

Henley, R.W., Truesdell, A.H., and Barton, P.B., Jr., 1984. Fluid-Mineral Equilibria in Hydrothermal Systems. Soc. Econ. Geol., Reviews Econ. Geol., 1, 257pp.

Henley, R.W., and Etheridge, M.A., 1995. The Structural and Hydrodynamic Framework for Epithermal Exploration. Proceedings of the 1995 Pacrim Conference, 269-277.

Hodder, A.P.W., 1987. Relationship of Palaeosubduction Regime and Prospectivity of an Epithermal Gold Field, Coromandel, New Zealand. Mineral. Deposita, 22, 247-252.

Irvine, R.J., and Smith, M.J., 1990. Geophysical Exploration for Epithermal Gold Deposits. J. Geochem. Expl., 36, 375-412.

Izawa, E., Urashima, Y., Ibaraki, K., Suzuki, R., Yokoyama, T., Kawasaki, K., Koga, A., and Taguchi, S., 1990. The Hishikari Gold Deposit: High-Grade Epithermal Veins in Quaternary Volcanics of Southern Kyushu, Japan. J. Geochem. Expl., 36, 1-56.

Jones, B.K., 1992. Application of Metal Zoning to Gold Exploration in Porphyry Copper Systems. J. Geochem. Expl., 43, 127-155.

Kavalieris, I., and Gonzalez, J.M., 1990. Geology and Mineralization of the Pao Prospect: A High Sulphidation Epithermal Gold System Hosted in Alkaline Volcanics, North Luzon, Philippines. In: Pacific Rim Congress '90, Austr. Insti. Min. Metall., 413-418.

Kontis, E., Kelepertsis, A.E., and Skounakis, S., 1994. Geochemistry and Alteration Facies Associated with Epithermal Precious Metal Mineralization in an Active Geothermal System, Northern Lesbos, Greece. Mineral. Deposita, 29, 430-433.

Kruse, F.A., Hummer-Miller, S., and Watson, K., 1988. Thermal Infrared Remote Sensing of the Carlin Disseminated Gold Deposit, Eureka County, Nevada. In: Schafer, R.W., Cooper, J.J., and Vikre, P.G., eds., Bulk Mineable Precious Metal Deposits of the Western United States, 734-735.

Larson, L.T., and Erler, Y.A., 1993. The Epithermal Lithogeochemical Signature - A Persistent Characterization of Precious Metal Mineralization at Kursunlu and Örencik, Two Prospects of Very Different Geology in Western Turkey. J. Geochem. Expl., 47, 321-331.

Licence, P.S., Terrill, J.E., and Fergusson, L.J., 1987. Epithermal Gold Mineralization, Ambitle Island, Papua New Guinea. In: Pacific Rim Congress '87, Austr. Insti. Min. Metall., 273-278.

Marcoux, E., and Milési, J.P., 1994. Epithermal Gold Deposits in West Java, Indonesia: Geology, Age and Crustal Source. J. Geochem. Expl., 50, 393-408.

Masterson, W.D., and Kyle, J.R., 1984. Geological, Geochemical, Stable Isotope and Fluid Inclusion Characteristics of Epithermal Gold Mineralization, Velvet District, Nevada. J. Geochem. Expl., 20, 55-74.

Mitchell, A.H.G., 1992. Andesitic Arcs, Epithermal Gold and Porphyry-Type Mineralization in the Western Pacific and Eastern Europe. Trans. Insti. Min. Metall., (Sect. B: Appl. Earth Sci.), 101, B125-B138.

Mitchell, A.H.G., and Balce, G.R., 1990. Geological Features of Some Epithermal Gold Systems, Philippines. J. Geochem. Expl., 35, 241-296.

Mitchell, A.H.G., and Leach, T.M., eds., 1991. Epithermal Gold in the Philippines: Island Arc Metallogenesis, Geothermal Systems and Geology, 457pp.

Nishikawa, N., 1992. The Use of Electrical Methods in Recent Exploration for Epithermal Gold Deposits in Japan. Expl. Geophys., 23, 249-254.

Readdy, L.A., 1988. Application of Photogeology and Remote Sensing to Geochemical Exploration for Epithermal Sediment-Hosted Gold Deposits. In: Schafer, R.W., Cooper, J. J., and Vikre, P.G., eds., Bulk Mineable Precious Metal Deposits of the Western United States, 733-734.

Reed, M.H., and Spycher, N., 1985. Boiling, Cooling, and Oxidation in Epithermal Systems: A Numerical Modeling Approach. In: Berger, B.R., and Bethke, P.M., eds., Geology and Geochemistry of Epithermal Systems. Soc. Econ. Geol., Reviews Econ. Geol., 2, 249-272.

Reyes, A.G., 1990. Petrology of Philippine Geothermal Systems and the Application of Alteration Minerolgy to their Assessment. J. Volcanol. and Geotherm. Res., 43, 279-309.

Rye, R.O., 1993. The Evolution of Magmatic Fluids in the Epithermal Environment: The Stable Isotope Perspective. Econ. Geol., 88, 733-753.

Rytuba, J.J., 1994. Evolution of Volcanic and Tectonic Features in Caldera Settings and Their Importance in the Localization of Ore Deposits. Econ. Geol., 89, 1687-1696.

Rytuba, J.J., and Miller, W.R., 1990. Geology and Geochemistry of Epithermal Precious Metal Vein Systems in the Intra-Oceanic Arcs of Palau and Yap, Western Pacific. J. Geochem. Expl., 35, 413-447.

Sänger-von Oepen, P., Friedrich, G., and Vogt, J.H., 1989. Fluid Evolution, Wallrock Alteration, and Ore Mineralization Associated with the Rodalquilar Epithermal Gold-Deposit in South-East Spain. Mineral. Deposita, 24 (4), 235-243.

Saunders, J.A., Utterback, W.C., Day, W.C., and Christian, R., 1988. Characteristics of Bonanza Epithermal Gold Mineralization at the Sleeper Deposit, Nevada, U.S.A. In: Bicentennial Gold '88, 386-388.

Schoen, R., White, D.E., and Hemley, J.J., 1974. Argillization by Descending Acid at Steamboat Springs, Nevada. Clays and Clay Minerals, 22, 1-22.

Seward, T.M. 1982. The Transport and Deposition of Gold in Hydrothermal Systems. In: Foster, R.P., ed., Gold '82: The Geology, Geochemistry and Genesis of Gold Deposits: Balkema, Rotterdam, Geol. Soc. Zimbabwe, Spec. Publ. No. 1, 165-181.

Seward, T.M., 1988. The Hydrothermal Chemistry of Gold and Its Implications for Ore Formation: Boiling and Conductive Cooling as Examples. In: Keays, R.R., Ramsay, W.R.H., Groves, D.I., eds., The Geology of Gold Deposits: The Perspective in 1988. Econ. Geol., Mono. 6, 398-404.

Silberman, M.L., and Berger, B.R., 1985. Relationship of Trace-Element Patterns to Alteration and Morphology in Epithermal Precious-Metal Deposits. In: Berger, B.R., and Bethke, P.M., eds., Geology and Geochemistry of Epithermal Systems. Soc. Econ. Geol., Reviews Econ. Geol., 2, 203-232.

Sillitoe, R.H., 1977. Metallic Mineralization Affiliated to Subaerial Volcanism: A Review. In: Volcanic Processes in Ore Genesis. Geol. Soc. London, Spec. Publi. 7, 99-116.

Sillitoe, R.H., 1981. Ore Deposits in Cordilleran and Island-Arc Settings. In: Dickson, W.R., and Payne, W.D., eds., Relations of Tectonics to Ore Deposits in the Southern Cordillera. Arizona Geol. Soc. Digest, 14, 237-262.

Sillitoe, R.H., 1988. Gold Deposits in Western Pacific Island Arcs: The Magmatic Connection. In: Keays, R.R., Ramsay, W.R.H., Groves, D.I., eds., The Geology of Gold Deposits: The Perspective in 1988. Econ. Geol., Mono. 6, 274-291.

Sillitoe, R.H., 1993. Epithermal Models: Genetic Types, Geometrical Controls and Shallow Features. In: Kirkham, R.V., Sinclair, W.D., Thorpe, R.I., and Duke, J.M., eds., Mineral Deposit Modelling. Geol. Assoc. Canada, Spec. Paper 40, 403-417, 798pp.

Sillitoe, R.H., 1994. Erosion and Collapse of Volcanoes: Causes of Telescoping in Intrusion-Centered Ore Deposits. Geol., 22, 945-948.

Simmons, S.F., and Browne, P.R.L., 1990. Mineralogic, Alteration and Fluid-Inclusion Studies of Epithermal Gold-Bearing Veins at the Mt. Muro Prospect, Central Kalimantan (Borneo), Indonesia. J. Geochem. Expl., 35, 63-103.

Simmons, S.F.,and Christenson, B.W., 1994. Origins of Calcite in a Boiling Geothermal System. American J. Sci., 294, 361-400.

Smith, M.J., and Corbett, G.J., 1990. Epithermal Gold Exploration in Covered Terrain at Togi, Japan. In: Pacific Rim Congress '90, Austr. Insti. Min. Metall., 785-787.

So, C-S., Dunchenko, V.Y., Yun, S-T., Park, M-E., Choi, S-G., and Shelton, K.L., 1995. Te- and Se-Bearing Epithermal Au-Ag Mineralization, Prasolovskoye, Kunashir Island, Kuril Island Arc. Econ. Geol., 90, 105-117.

Stoffregen, R.E., 1987. Genesis of Acid-Sulfate Alteration and Au-Cu-Ag Mineralization at Summitville, Colorado. Econ. Geol., 82, 1575-1591.

Swindell, E.P.W., 1986. Epithermal Gold Deposits. Expl. Geol. Dept., Internat. Div., Gencor Internal Report, 43pp.

Syka, M.M., and Bloom, M.S., 1990. The Geology and Paragenesis of Epithermal Gold Mineralization at the Kerimenge Prospect, Papua New Guinea. In: Pacific Rim Congress '90, Austr. Insti. Min. Metall., 293-299.

Thompson, J.F.H., Abidin, H.Z., Both, R.A., Martosuroyo, S., Rafferty, W.J., and Thompson, A.J.B., 1994. Alteration and Epithermal Mineralization in the Masupa Ria Volcanic Centre, Central Kalimantan, Indonesia. J. Geochem. Expl., 50, 429-456.

Tischler, S.E., 1990. Geotectonic Framework of Epithermal/Mesothermal Gold Mineralization Styles: Its Use in the Design of Exploration Methodology. In: Pacific Rim Congress '90, Austr. Insti. Min. Metall., 545-548.

Vikre, P.G., 1985. Precious Metal Vein Systems in the National District, Humboldt County, Nevada. Econ. Geol., 80, 360-393.

Wake, B.A., and Taylor, G.R., 1988. Major's Creek, N.S.W., Australia - A Devonian Epithermal Gold Deposit. Mineral. Deposita, 23, 239-246.

Webster, S.S., Henley, R., and Harris, L., 1989. High Resolution Airborne Geophysics for Epithermal Gold Exploration in North-Eastern Queensland. North Queensland Gold '89 Conference, Austr. Insti. Min. Metall., 41-46.

White, N.C., and Hedenquist, J.W., 1990. Epithermal Environments and Styles of Mineralization: Variations and Their Causes, and Guidelines for Exploration. J. Geochem. Expl., 36, 445-474.

White, N.C., and Hedenquist, J.W., 1995. Epithermal Gold Deposits: Styles, Characteristics and Exploration. Econ. Geol., 23, 8-13.

White, N.C., Leake, M.J., McCaughey, S.N., and Parris, B.W., 1995. Epithermal Gold Deposits of the Southwest Pacific. J. Geochem. Expl., 54, 87-136.

Worthington, J.E., 1981. Bulk Tonnage Gold Deposits in Volcanic Environments. In: Dickson, W.R., and Payne, W.D., eds., Relations of Tectonics to Ore Deposits in the Southern Cordillera: Arizona Geol. Soc. Digest, 14, 263-270.