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Epithermal gold deposits: Styles, characteristics, and exploration

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Epithermal gold deposits: Styles, characteristics, and exploration

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Epithermal gold deposits: Styles, characteristics, and exploration

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1) Introduction

Epithermal ore deposits form at shallow depth. Lindgren [ref. 35] based this conclusion on ore mineralogy and related textures, plus geologic reconstructions; more recent studies indicate that most epithermal ores form over the temperature range of <150 to ~300°C, from the surface to as deep as 1 to 2 km[6]. Based on our observations and those from the literature, we summarize the principal characteristics of epithermal mineralization in which Au is the dominant economic metal. We highlight this brief summary with case studies of eight deposits that represent two end-member styles of epithermal Au deposits, high sulfidation (HS) and low sulfidation (LS)(Table 1.1); these examples represent the range of depth that epithermal deposits form, from just below the surface to more than 1 km depth.

The two deposit styles form from fluids of distinctly different chemical composition in contrasting volcanic environments (Fig. 1.1). The ore of HS deposits is hosted by leached silicic rock (Fig. 1.2) associated with acidic fluids generated in the volcanic-hydrothermal environment (Figs. 1.1 and 1.3). The presence of high-sulfidation state sulfide minerals (section 4) indicates high-oxidation states typical of acidic hypogene fluids [17]. In contrast, the fluid responsible for formation of LS ore veins (Fig. 1.4) is similar to waters tapped by drilling beneath hot springs (Fig. 1.5) into geothermal systems [25]; low-sulfidation state minerals form from these reduced, neutral-pH waters.

This poster has two purposes: 1) to introduce students of economic geology to one of the principal environments of formation of hydrothermal ore deposits, and 2) to initiate discussion among exploration geologists regarding the essential nature of epithermal deposits. We stress the basic characteristics of epithermal systems, especially those that can be recognized easily in the field, as these characteristics reflect ore-forming processes. Interpretations of such processes can then be used to guide efficiently mineral exploration [60].



High-sulfidation deposits = volcanic-hydrothermal environment
Fig. 1.2 (top) Kasuga high-sulfidation deposit, Nansatsu district, Japan. Products of hypogene acidic alteration include Au-mineralized residual silica (>95% SiO₂) in sharp contact with alunite-kaolinite halo (downwards and to right)[21].
Fig. 1.3 (bottom) Volcanic-hydrothermal system, Satsuma Iwojima, Japan. 870°C fumaroles vent from summit of rhyolite dome; acidic (pH 1.5) hot springs, rich in Fe and Al leached from host rock, discharge from volcano flanks to the sea [22].

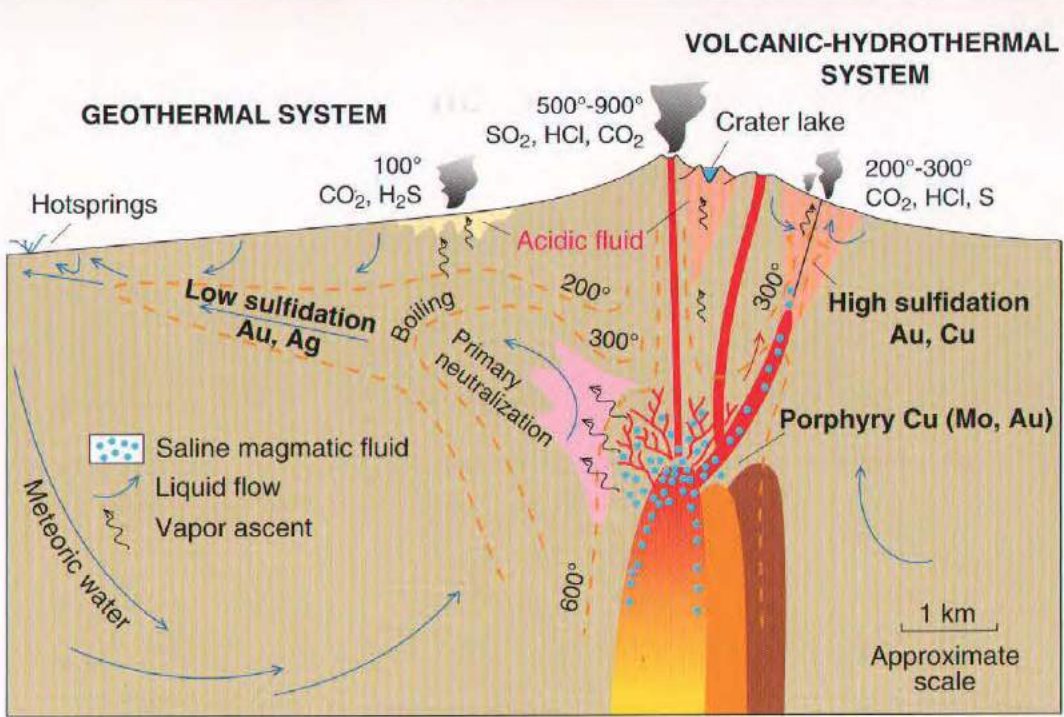
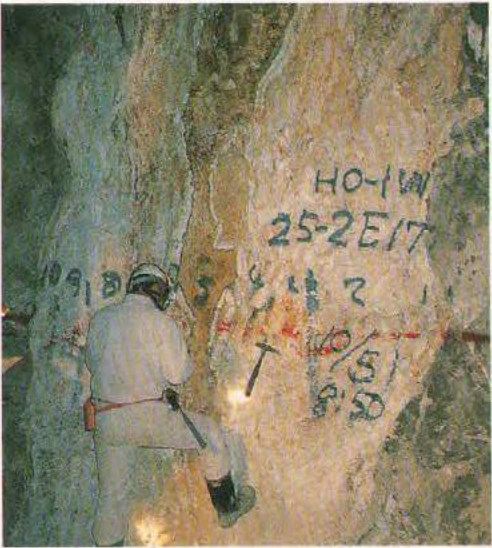


Fig. 1.1 Schematic cross-section showing shallow sub-volcanic intrusions and associated stratovolcano, and environments deduced for formation of porphyry Cu, and high- and low-sulfidation epithermal ore deposits [20,25]. Active volcanic-hydrothermal systems extend from degassing magma to fumaroles and acidic springs, and incorporate porphyry and/or high-sulfidation ore environments, whereas low-sulfidation ore deposits form from geothermal systems characterized by neutral-pH waters that may discharge as hot springs.

Table 1.1 General characteristics of epithermal gold deposits associated with subaerial volcanic rocks [6,16,49,60,62]

	Low sulfidation (LS) (Adularia-sericite)[16, 60]	High sulfidation (HS) (Acid-sulfate)[16, 60]
Deposit form	Open-space veins dominant, stockwork ore common Disseminated and replacement ore minor	Disseminated ore dominant, replacement ore common Stockwork ore minor, veins commonly subordinate
Textures	Veins, cavity filling (bands, colloforms, druses), breccias	Wallrock replacement, breccias, veins
Ore minerals	Pyrite, electrum, gold, sphalerite, galena (arsenopyrite)	Pyrite, enargite, chalcopyrite, tennantite, covellite, gold, tellurides
Gangue	Quartz, chalcedony, calcite, adularia, illite, carbonates	Quartz, alunite, barite, kaolinite, pyrophyllite
Metals	Au, Ag, Zn, Pb (Cu, Sb, As, Hg, Se)	Cu, Au, Ag, As (Pb, Hg, Sb, Te, Sn, Mo, Bi)



Low-sulfidation deposits = geothermal environment
Fig. 1.4 (left) High-grade, low-sulfidation quartz-adularia veins, Hishikari deposit, Japan [28] (photo by K.Ibaraki).



Fig. 1.5 (right) Champagne Pool, Waiotapu geothermal system, New Zealand. Neutral-pH hot spring rimmed by sinter with orange As-Sb-Au-Ag-Hg-Tl-rich precipitates. Steam-heated acid-sulfate water forms kaolinite alteration in craters adjacent to hot spring [18].

2) Tectonic Setting

Epithermal Au deposits occur largely in volcano-plutonic arcs (island arcs as well as continental arcs) associated with subduction zones (Fig. 2.1). Epithermal deposits occur commonly on the back-arc side, within 100 km of the active volcanic front. Subduction-induced magmatism and associated hydrothermal activity has a tendency to shift trenchward with time in several arcs, for example in Kyushu and New Zealand, and hydrothermal activity and Au mineralization occurred about 0.5 Ma after the initiation of volcanism in each volcanic center. Epithermal mineralization is associated with porphyry Cu (Au) type mineralization [48] in several places such as the Baguio district and Lepanto, Philippines, and in Chile and Papua New Guinea (e.g., Ladolam on Lihir island).

Gold deposits are hosted mainly by volcanic rocks and contemporaneous volcanogenic sedimentary rocks, and sometimes in basement units (e.g., Baguio district, Hishikari, Porgera). Although associated igneous rocks are commonly subaerial volcanic rocks of magnetite series calc-alkalic andesite-dacite-rhyolite composition, shoshonitic or alkalic igneous rocks may occur in mature (crustal thickness >20 km) intraoceanic arcs (e.g., Ladolam and Emperor) and in continental regions (e.g., Cripple Creek). In Kyushu, New Zealand and Indonesia, Au deposits occur in volcano-tectonic depressions (grabens), which represent the pull-apart structures resulting from strike-slip fault movement.

Most principal epithermal Au deposits in the western Pacific formed during the late Miocene, Pliocene and Quaternary. The ages for deposits in the western Americas and Caribbean are early Cretaceous to late Miocene (110-9 Ma [51]). This correlation reflects the difference in erosion rates between island and continental arcs (Fig. 2.3). Erosion rates in island arcs are high owing to rapid uplift and high rainfall, and vary between 0.1 mm/a in Kyushu and almost 1 mm/a in the Philippines and Papua New Guinea [51]. In contrast, erosion rates in continental settings are low (e.g., a rate of 0.01 mm/a is common), except during sector collapse of the volcanic edifice.

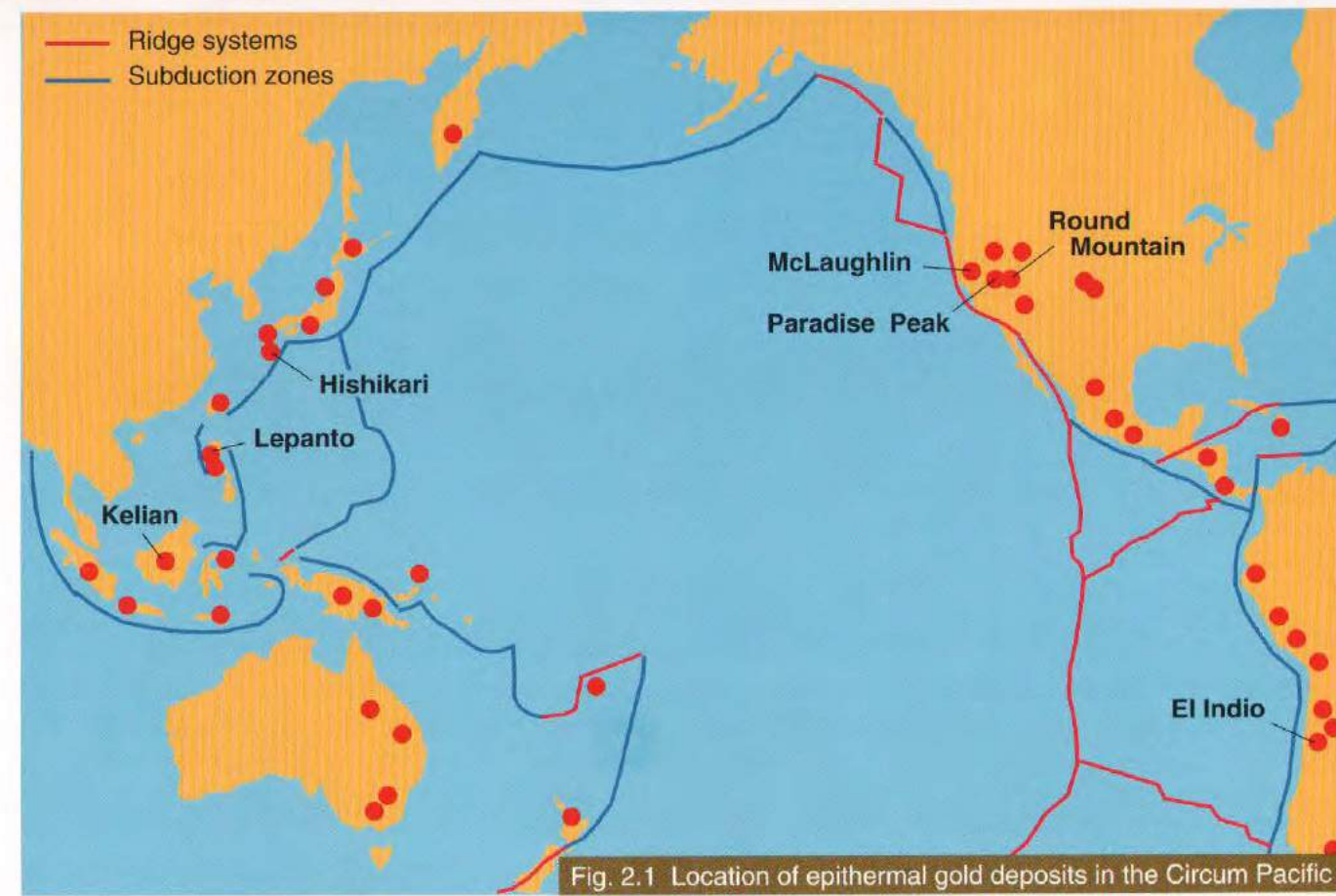


Fig. 2.1 Location of epithermal gold deposits in the Circum Pacific

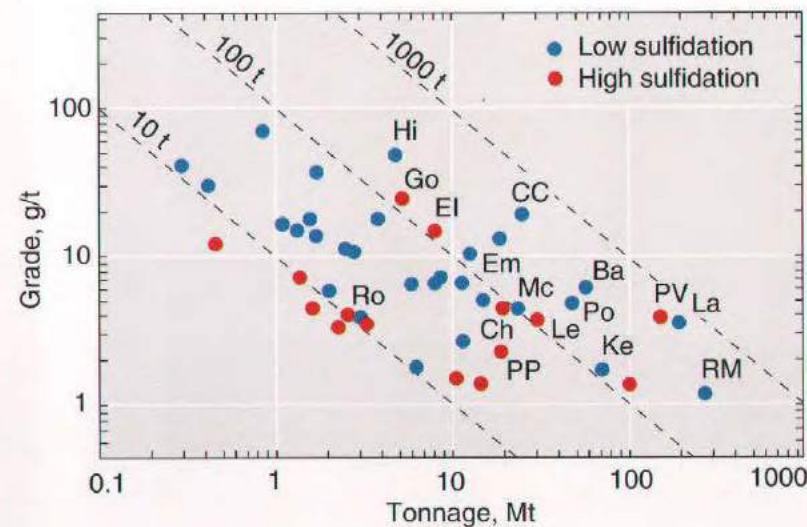


Fig. 2.2 Grade-tonnage plot for epithermal Au deposits. LS: Ba, Baguio (Philippines); CC, Cripple Creek; Mc, McLaughlin and RM, Round Mountain (USA); Em, Emperor (Fiji); Hi, Hishikari (Japan); Ke, Kelian (Indonesia); La, Ladolam and Po, Porgera Zone VII (Papua New Guinea). HS: Ch, Chinkuashih (Taiwan); El, El Indio (Chile); Go, Goldfield and PP, Paradise Peak (USA); Le, Lepanto (Philippines); PV, Pueblo Viejo (Dominican Republic); Ro, Rodalquilar (Spain) [31,48,50].

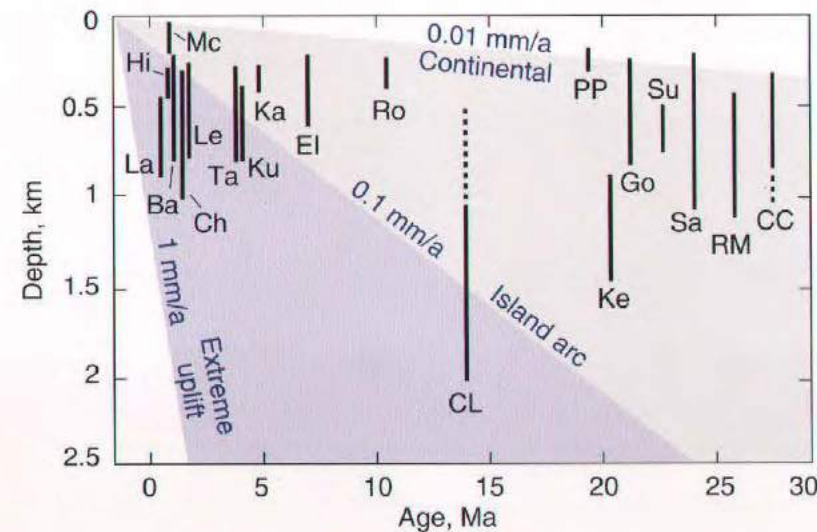


Fig. 2.3 Age of epithermal Au deposits and vertical extent of mineralized zones. Typical erosion rates of volcano-plutonic arcs are shown for reference: 0.01 mm/a, at continental margins; 0.1 mm/a, in island-arcs; 1 mm/a, in mountainous regions, e.g., Papua New Guinea and the Philippines. Abbreviations from Fig. 2.2, plus Ka, Kasuga; Ku, Kushikino; Sa, Sado and Ta, Taio (Japan); CL, Comstock Lode and Su, Summitville (USA) [31].

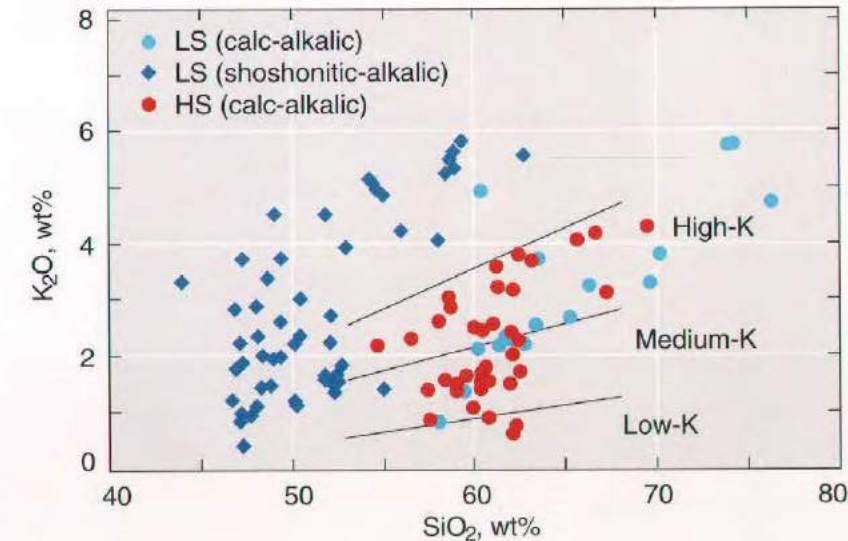


Fig. 2.4 Plot of K_2O versus SiO_2 for volcanic rocks [15] thought to be genetically related to Au mineralization [1,31,50]. HS deposits appear to be associated with a narrow range of igneous rock composition, dominated by dacite and andesite. LS deposits are largely associated with andesite and rhyolite, but major deposits such as Ladolam, Porgera and Cripple Creek are associated with shoshonitic and alkalic rocks [31].

3) Alteration

Many hydrothermal minerals are stable over limited temperature and/or pH ranges (Fig. 3.1)[8,25,41]. Thus, mapping the distribution of alteration minerals allows the thermal and geochemical zonation to be reconstructed, leading to a model of the hydrology of the extinct hydrothermal system. Alteration minerals are also crucial in distinguishing the style of deposit, LS or HS (Fig. 3.2).

The ore-associated alteration in LS deposits is produced by near-neutral pH thermal waters; temperature decreases with decreasing depth and increasing distance from the fluid conduits. Paleoisotherms can be drawn by mapping alteration mineralogy (particularly the clays, e.g., Hishikari), to identify the conduits of paleoflow. The level of erosion can also be determined; prospects with indications of low paleotemperatures are encouraging, whereas indications of high paleotemperatures suggest that the depth interval with potential for epithermal ore (Fig. 3.1) may be limited.

In contrast to the neutral-pH alteration characteristic of LS deposits, HS deposits contain acid-stable minerals [23,42] such as alunite, kaolinite, dickite, pyrophyllite, diaspore and zunyite, some of which are also temperature sensitive (Fig. 3.1). These minerals comprise the advanced argillic alteration assemblage formed during initial leaching in the HS environment. The most acid-altered rock is a silica residue, termed vuggy quartz; this leached rock is commonly the ore host, with a halo of advanced argillic alteration (Fig. 3.2) zoned outward to illite, illite/smectite or smectite, depending on the paleotemperature.

Some lower temperature acid-stable minerals, such as cristobalite, kaolinite and alunite, are also typical of steam-heated alteration that forms in the vadose zone, above the groundwater table, because of H₂S oxidation (Fig. 3.2)[45]. These moderately acidic waters (pH 2-3) generally overlie ore in LS deposits, but they may flow downwards along fractures and be heated [41], thus possibly overprinting ore and resulting in the apparent paradox of acidic alteration in association with LS ore [49,60]. Significant erosion may occur in volcanic arcs during hydrothermal activity [41,51], resulting in collapse of the system and the formation of a steam-heated alteration overprint to shallow ore; this late-stage alteration must be distinguished from that associated with ore.

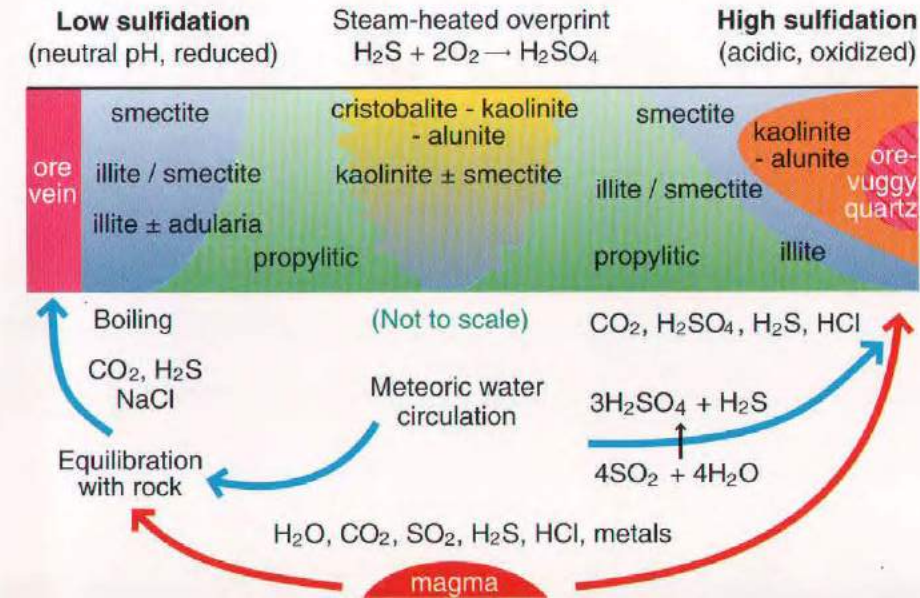


Fig. 3.2 (1.) Schematic distribution of hydrothermal alteration associated with LS [9] and HS [54] epithermal deposits [60]; ore is located in paleoconduits (Fig. 4.1). Quartz is stable throughout. Propylitic alteration occurs outside conduit zones (i.e. low water-rock ratio). Steam-heated alteration can occur in LS or HS systems, but is most apparent in LS deposits, where mineralogy is distinct from that associated with ore (Table 4.1). In LS systems, silica sinter may form where hot springs discharge at surface; in high-relief areas hot water may flow laterally and be diluted before discharge (Fig. 1.1), precluding sinter formation. Thus, veins may terminate at paleosurface in sinter (e.g., McLaughlin), or may pinch out upwards into argillic-altered rock (e.g., Hishikari). In HS deposits, ore occupies parts of intensely acid-altered cores of alteration zones, which typically pinch out above and below orebodies.

Beneath the epithermal environment, a degassing magma body contributes heat, water, acidic gases, and ore metals. In LS systems, the magmatic component equilibrates with wall rocks at depth during meteoric-water dominated convection, before ascent to epithermal depths [14]. In HS systems, magmatic volatiles ascend to the epithermal environment with little modification, then are absorbed by meteoric water to form an acidic hypogene fluid that leaches wallrock outward from conduits. Ore metals are commonly introduced by late-stage magmatic-meteoric water mixtures [20].

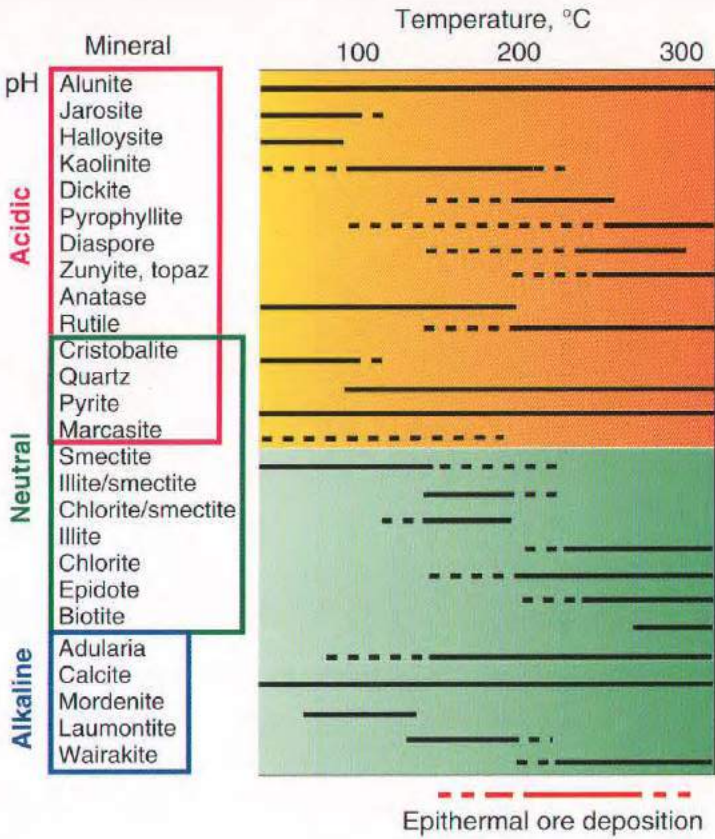


Fig. 3.1 Temperature stability of hydrothermal minerals (deduced from geothermal studies [25,41]) common in the epithermal environment, arranged by their stability with respect to pH. Zones of mineral assemblages may be more meaningful than individual minerals in mapping paleoisotherms [31,41].

Legend for sections 3 to 13

Geologic units :

- Post-mineral cover
- Syn-hydrothermal breccia
- Basalt (lava and scoria)
- Dacite / rhyolite (lava dome)
- Dacite / rhyolite (pyroclastic flow deposits)
- Andesite / dacite (lava and pyroclastic deposits)
- Intrusion
- Basement rock
- Geologic contact
- Fault

Alteration:

- Advanced argillic (low T) (cristobalite, kaolinite, alunite)
- Advanced argillic (high T) (quartz, dickite, pyrophyllite, alunite)
- Quartz-sericite (quartz, illite, pyrite)
- Silicification and residual silica (quartz, rutile or anatase)
- Argillic (low T) - smectite (cristobalite, smectite, kaolinite)
- Argillic (high T) - illite (quartz, illite, illite/smectite)
- Propylitic (low T) - smectite (smectite, igneous plagioclase)
- Propylitic (high T) - chlorite (albite, chlorite, epidote, calcite)
- Ore

Note: Rock and alteration [40] types are necessarily simplified; see original references for details.

4) Mineralization

The most basic characteristics of any ore deposit are the form of the orebody (Fig. 4.1) and its mineralogy (Table 4.1), the textures of ore and gangue minerals, and alteration zoning (section 3). Comparisons of observations [6,16,49,60,62] of LS and HS deposits highlight the considerable overlap in characteristics. There are, however, many distinctive features, such as the difference in the typical deposit form; both styles usually are structurally controlled, although in HS disseminated deposits the structure may be concealed. Epithermal deposits are extremely variable in form (Fig. 4.1), because of the low-pressure, hydrostatic conditions under which they form [49]. Much of the geometrical variability results from the effect of permeability differences (structural, hydrothermal, lithologic; Fig. 4.1) in the host rock.

Although there are many minerals common in both LS and HS deposits, several clear differences exist in ore mineralogy; some of them reflect the distinct redox conditions of the hydrothermal fluids. Arsenopyrite and high-Fe sphalerite, both low-sulfidation state minerals [4], are common in LS deposits, whereas they are scarce to rare in HS deposits (Table 4.1). By contrast, HS deposits commonly contain Cu-As minerals, especially the high-sulfidation state sulfosalts enargite and luzonite. Such sulfides, including the relatively high-sulfidation state mineral tennantite [4], are comparatively rare or absent in LS deposits. The total abundance of sulfide minerals (mainly pyrite) is variable, and can be higher in HS than LS deposits [49].

The gangue minerals associated with the two styles show considerable overlap (Table 4.1), but there are clear differences reflecting the pH (reactivity) of the ore fluid. Quartz is ubiquitous in both styles. Adularia and calcite, both indicating near-neutral pH conditions (section 3), are common in LS deposits (most common after quartz [9]), but are absent in HS deposits. Minerals formed under relatively acidic conditions, such as kaolinite and alunite, are common but minor as gangue in HS deposits. In LS deposits, alunite does not occur with ore, but is common in areas affected by steam-heated (and supergene) overprinting by surficial waters (section 3).

The ore and gangue mineral textures of the two styles of deposit are distinct (Table 1.1). LS deposits show a variety of textures including banding (Fig. 4.2), cavities filled by crustiform quartz and mineral druses, and multiple generations of breccia. Lattice-textured bladed calcite (Fig. 4.2), a common consequence of boiling [53], may be pseudomorphed by quartz (Fig. 4.2). Silica sinters (Fig. 4.3) may be present in areas of little erosion [61], indicating paleosurface. In contrast, the typical textures of HS deposits show little variation; the most characteristic feature is massive bodies of residual (vuggy) quartz (Fig. 4.4) formed as a result of leaching by acidic (pH <2) 200-250°C fluid. Massive to banded sulfide veins of pyrite and enargite may cut the vuggy quartz bodies. Silica sinters do not form in this acidic environment [60].

Table 4.1 Frequency and abundance of ore (red) and gangue (blue) minerals in Au-rich epithermal deposits

Low sulfidation	High sulfidation
Ubiquitous	
Pyrite (a)	Pyrite (a)
Quartz (a)	Enargite-luzonite (±)
	Quartz (a)
Common	
Electrum (±)	Native gold (vm)
Native gold (vm)	Tellurides (vm)
Chalcopyrite (vm)	Covellite (m)
Sphalerite (±)	Tennantite (±)
Galena (±)	Tetrahedrite (±)
Tetrahedrite (vm)	Chalcopyrite (m)
Arsenopyrite (m)	Sphalerite (±)
Tellurides (vm)	Galena (±)
Pyrargyrite (vm)	Barite (m)
Chalcedony (±)	Alunite (±)
Adularia (±)	Kaolinite (m)
Illite (a)	Pyrophyllite (±)
Calcite (±)	Diaspore (vm)
Smectite (m)	Illite (m)
Uncommon or rare	
Selenides (vm)	Electrum (vm)
Stibnite (vm)	Selenides (vm)
Cinnabar (vm)	Pyrargyrite (vm)
Enargite-luzonite (vm)	Arsenopyrite (vm)
Tennantite (vm)	Cinnabar (vm)
Covellite (vm)	Stibnite (vm)
Barite (vm)	Chalcedony (m)
Kaolinite (vm)	Smectite (m)
Absent except as overprint	
Pyrophyllite	Calcite
Diaspore	Adularia
Alunite	

Based on >200 deposits in the western Pacific and North American regions [9,62]. Abundant, a; minor, m; very minor, vm; variable, ±.

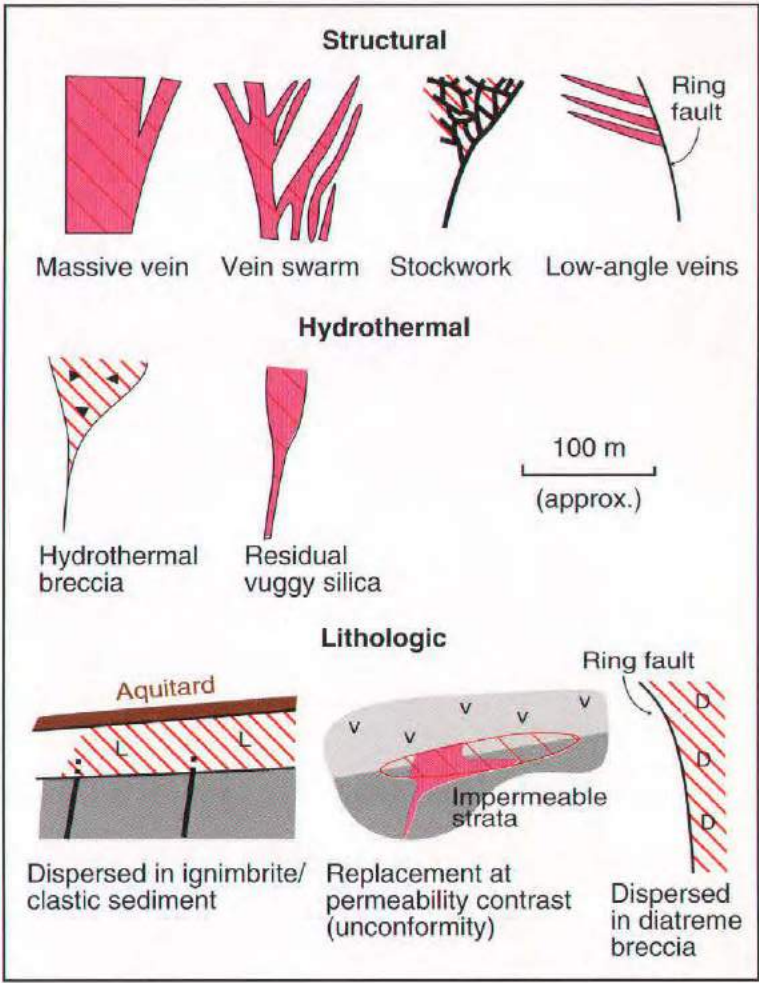


Fig. 4.1 Selected styles and geometries of epithermal deposits to illustrate schematically the influence of structural, hydrothermal and lithologic controls on permeability, i.e., fluid conduits; from Sillitoe [49].



Fig. 4.2 (L.) Banded botryoidal quartz vein from the Sleeper LS deposit, Nevada [44]; note electrum bands and quartz after bladed calcite (photo by J. Saunders). (R.) Coarse bladed calcite from the Martha Hill LS deposit, New Zealand (photo by S. Simmons).



Fig. 4.3 Diagnostic textures of silica sinter [61]. (L.) Columnar growth structures perpendicular to sinter laminations, Devonian Durah Creek sinter, Queensland (coin, 19 mm). (R.) Columnar structure formed by bacterial stromatolite, Broadlands geothermal system, New Zealand (photo 6 cm wide).

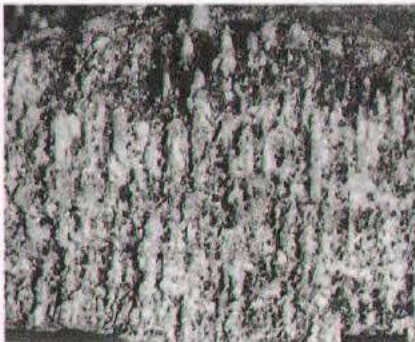


Fig. 4.4 Diagnostic textures of HS epithermal deposits. (L.) Vuggy quartz caused by leaching of sanidine phenocrysts (≤4 cm long) by acidic fluid (pH <2 [55]). (R.) Same rock (quartz latite porphyry) altered to alunite-kaolinite in the alteration halo to vuggy quartz ore zone of Summitville deposit.



Low-sulfidation deposits

5) McLaughlin

The McLaughlin deposit was discovered in 1978 at the site of the Manhattan Hg mine in northern California [47,57]. Ore veins occur along the contact between ophiolitic footwall rocks and overlying mudstones (Fig. 5.1). The deposit formed just beneath the paleosurface, as evidenced by the preservation of silica sinter over parts of the ore body (Fig. 5.4). The size of the deposit, including production and reserves as of 1993, was 24.3 Mt at an average Au grade of 4.49 g/t. Gold is present in veins as electrum, and is commonly associated with brown hydrocarbon-bearing opal or chalcedony. Cinnabar is present sporadically throughout the deposit, and stibnite is common in late-stage veins. Multiple generations of pyrite and marcasite do not correlate with Au content. Barite and carbonate minerals (calcite, dolomite, siderite, magnesite) occur as gangue, with calcite marginal to the ore body and magnesite more common at depth. The deposit is anomalous in the typical LS metal suite, Au-Ag-Sb-As-Hg-Tl, whereas base metals, Mn, Se, Te, F and Mo are not anomalous.

The alteration of the hanging-wall mudstone and silicification of the footwall serpentized ophiolite were critical in leading to fracturing and vein formation. Alteration and vein formation began prior to shallow intrusion of basaltic andesite, although veins also cut these intrusions. Phreatomagmatic breccias localized along the Stony Creek fault can be traced to these syn-hydrothermal intrusions (Fig. 5.1).

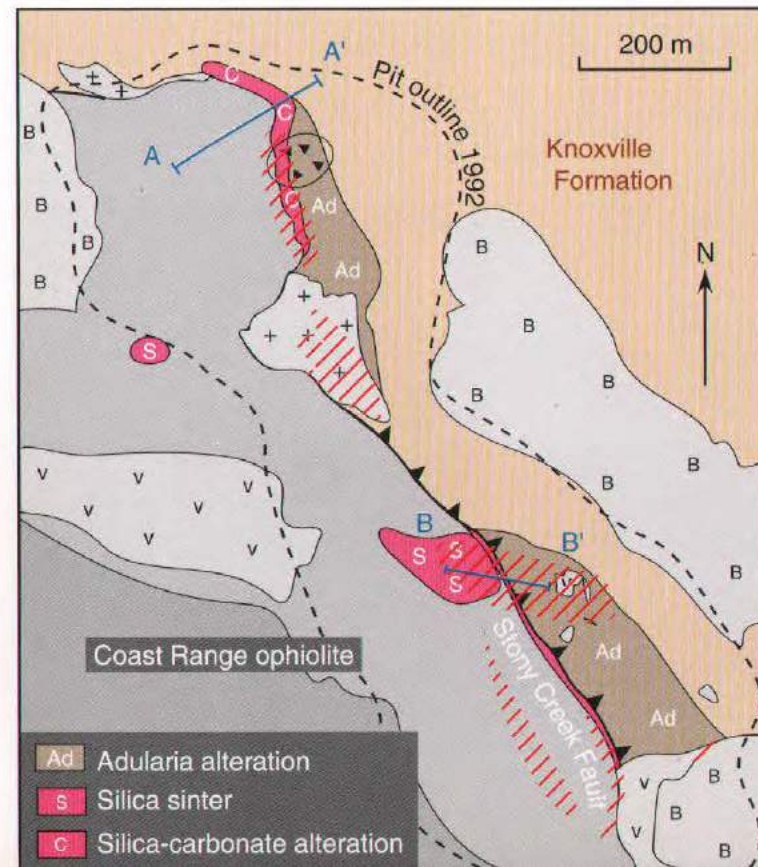


Fig. 5.1 The McLaughlin deposit is localized along contact between serpentized ultramafic and mafic rocks of the Middle Jurassic Coast Range ophiolite (footwall) and mudstone of the Upper Jurassic Knoxville Formation (hanging wall) (basal formation of Great Valley sequence). The Stony Creek fault separates these units, and dips moderately to NE. Pliocene basaltic andesite and volcanoclastic rocks unconformably overlie sedimentary and ophiolitic rocks in area. North and South ore bodies are connected by a narrow zone of mineralization [57].

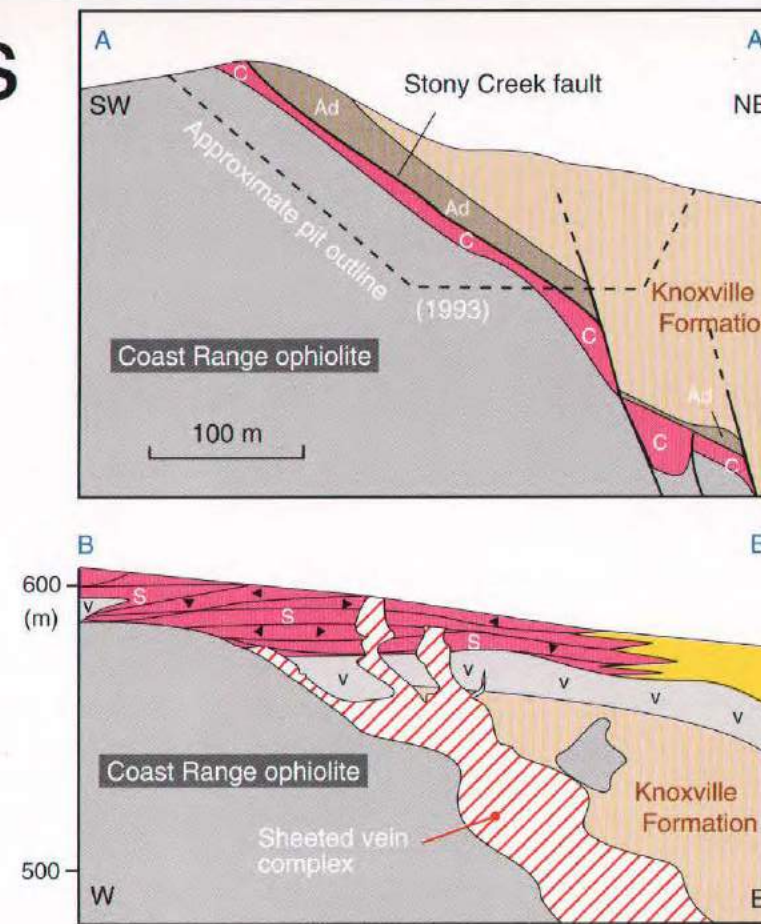


Fig. 5.2 (t.) In the North ore body, hanging-wall mudstone is altered to adularia, and footwall serpentinites are silicified, forming a discontinuous tabular body along Stony Creek fault, site of paleoconduit. Overlying basaltic andesite shows propylitic and locally argillic alteration [57].

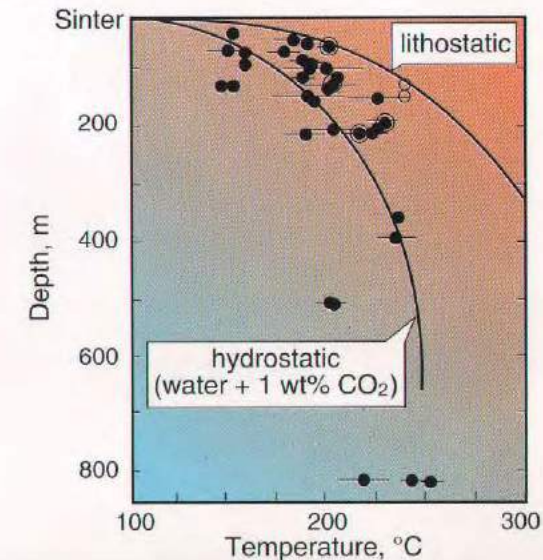
Fig. 5.3 (b.) South ore body is dominated by a sheeted vein complex that extended upward into silica sinter (Fig. 5.4) and interbedded silicified hydrothermal eruption breccia. Minor advanced argillic alteration, consisting of alunite and kaolinite of steam-heated origin, occurred adjacent to the sinter. The sinter was enriched in Hg, hence the old mining activity; Au concentrations in sinter were low, typically <0.3 g/t in the laminated sinter, except where the sinter was cut by mineralized veins at its base [47]. Ore-grade Au and Ag mineralization is restricted to the uppermost 250 m of veins.

Fig. 5.6 (r.) Fluid inclusion data indicate that temperatures in the McLaughlin deposit were controlled by boiling of a low-salinity, gas-rich fluid. However, at shallow depth intermittent sealing by mineral deposition may have led locally to pressures exceeding hydrostatic; tectonic rupture [18] followed by hydraulic fracturing and brecciation would thus account for the multiple episodes of fracturing and formation of hydrothermal breccia deposits [17].



Fig. 5.4 (t.) Original outcrop of San Quentin sinter, adjacent to Manhattan Hg mine, showing typical laminated textures (photo by Etheridge Henley Williams).

Fig. 5.5 (b.) Gold mineralization at McLaughlin occurs in opal, chalcedony, and quartz veins in multiply fractured host rock; South ore body.



6) Hishikari

The bonanza-type Au deposit at Hishikari is located in the Hokusatsu district, southern Kyushu, Japan. A regional exploration program by the Metal Mining Agency of Japan led to its discovery in 1981. Production from July, 1985 to March, 1996 totalled 74 metric tons of Au, making it one of the major Au deposits in the western Pacific [50]. Reserves including ore produced are estimated to be 3.2 Mt at an average grade of 63 g/t Au in the Honko-Sanjin ore zone and approximately 2 Mt at about 25 g/t Au in the Yamada ore zone (Fig. 6.1); a total of at least 250 t of Au was precipitated in the 750-m wide vein system, which extends over 2 km along strike [28].

The geology of the mine area is composed of basement metasedimentary rocks of the Cretaceous Shimanto Supergroup, and volcanic rocks of Quaternary age (Figs. 6.1 and 6.2). The Hishikari veins consist mainly of quartz and adularia with minor smectite, Au-Ag minerals such as electrum, naumannite-aguilare and pyrrargyrite, and very minor sulfide minerals such as chalcocopyrite, tetrahedrite, sphalerite, galena and pyrite; the average Ag/Au value is 0.7. K-Ar dating of adularia in ore indicates an early Pleistocene age of 1.25 to

0.6 Ma for Au mineralization [30].

Low-temperature argillic alteration (cristobalite-smectite and quartz-kaolinite) is widespread at the present surface, and quartz-alunite and cristobalite-alunite occur at higher elevations (350 to 400 m) in the eastern part of the area [28]. A hydrothermal eruption crater filled with mudstone and sinter is present above the Yamada deposit [29], beneath young ash-flow deposits (Fig. 6.3). Fluid inclusion studies indicate that temperatures of Au deposition were 210°C in basement and <200°C in volcanic rocks of the Honko area, suggesting about 200 m of erosion since vein formation [50].

The vertical and horizontal distribution of wallrock alteration assemblages indicate a simple zonation (Figs. 6.1 and 6.2). Chlorite-illite alteration is associated directly with high-grade Au mineralization at the Honko-Sanjin ore zone. Mixed-layer clays and smectite envelop the mineralized center and form a near-horizontal layer of intense argillization, located 50 to 100 m above the Honko-Sanjin veins. These alteration types are surrounded by a zone of cristobalite-smectite transitional to unaltered rocks. There is a slight

downward drape to the alteration zones, reflecting distance from the veins (Fig. 6.2). In contrast, the ore-grade veins at Yamada occur in the mixed-layer clay and chlorite-illite zones. The extremely high-grade Au mineralization occurs near the unconformity between basement and overlying volcanic rocks in the Honko-Sanjin area, whereas volcanic rocks host the Yamada veins (Figs. 6.3 and 6.4).

Studies of chemical variation related to Au mineralization show that most of the unaltered volcanic rocks at Hishikari contain 2 to 5 ppm As, 10 to 40 ppb Hg and <1 to 1 ppb Au. Argillized surface rocks directly above the Honko veins contain up to 15 ppm As, 1500 ppb Hg and 31 ppb Au [26].

Subsurface structures were recognized by a detailed gravity survey, with uplifted basement blocks represented by gravity highs. Schlumberger vertical soundings were useful in estimating the depth to the basement and to identify the approximate resistivity structure, while resistivity and IP surveys defined the areal distribution of the hydrothermal alteration related to ore.

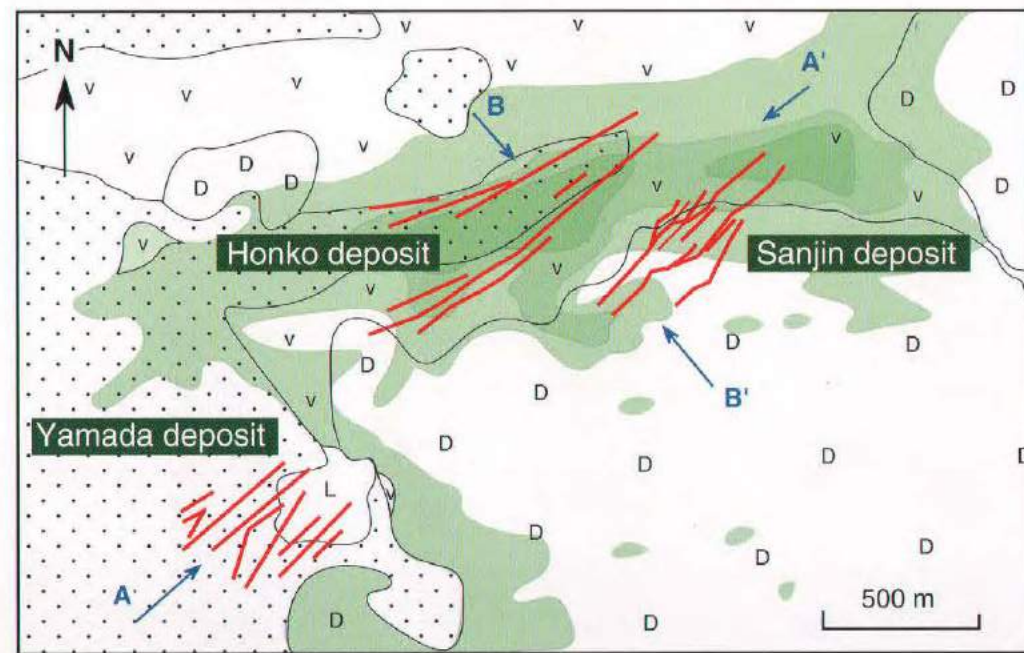


Fig. 6.1 Geology and alteration of Hishikari deposit. Western part of area covered by young pyroclastic flow deposits. Red lines show surface projections of veins. Hydrothermally altered volcanic rocks are grouped into four zones based on mineral assemblages [28]. The deepest and innermost chlorite-illite zone (Fig. 6.2) is defined by presence of chlorite and local illite. The typical mineral assemblage is quartz-chlorite-adularia-calcite-albite. The mixed-layer clay mineral zone is characterized by presence of mixed-layer chlorite/smectite and/or illite/smectite. Quartz-adularia-calcite-smectite-laumontite are commonly present. Presence of quartz with smectite and/or kaolinite defines quartz-smectite zone. Cristobalite and tridymite occur with smectite in cristobalite-smectite zone.

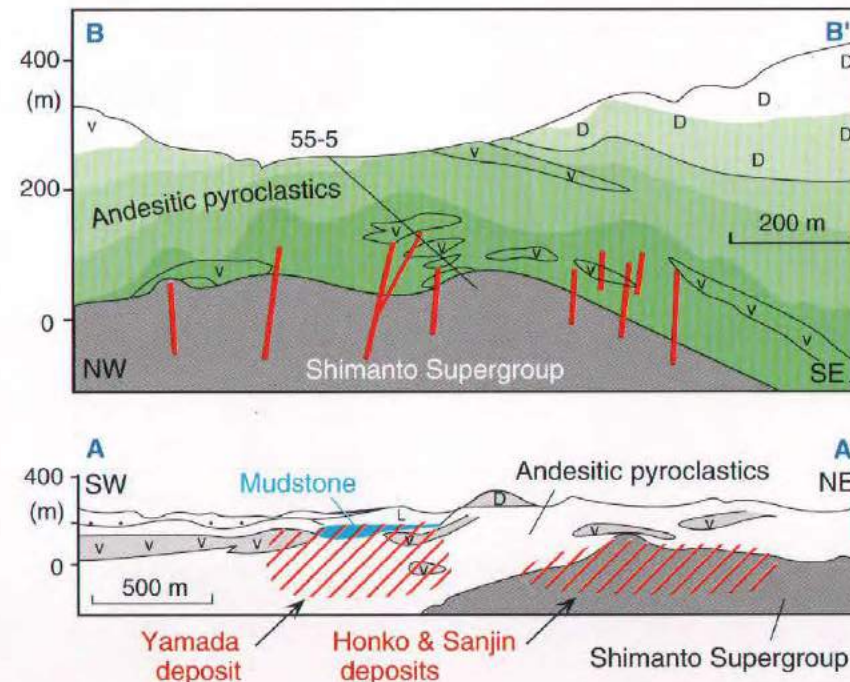


Fig. 6.2 (L) Vertical alteration zoning in section [28]. Hydrothermal alteration of basement sedimentary rocks is developed only as narrow halos to veins. Alteration is represented by chlorite-illite assemblage, although this is difficult to distinguish from metamorphic chlorite-sericite in graywacke. 55-5: discovery drillhole, 1981. High-T and Low-T propylitic alteration here subdivided into four zones.

Fig. 6.3 (b.) Schematic longitudinal section of Hishikari deposit [26]. Honko and Sanjin veins are largely developed in Shimanto Supergroup basement, whereas lower grade veins of Yamada ore body are hosted by Pleistocene volcanic units.



Fig. 6.4 Vein formation at Hishikari was influenced by unconformity between basement sedimentary rocks and overlying volcanic rocks. A clear example is shown here where the Zuisen 1 vein in the Honko system splays and feathers out on passing from the Shimanto graywacke basement to the overlying volcanic rocks (photo by K. Ibaraki).

7) Round Mountain

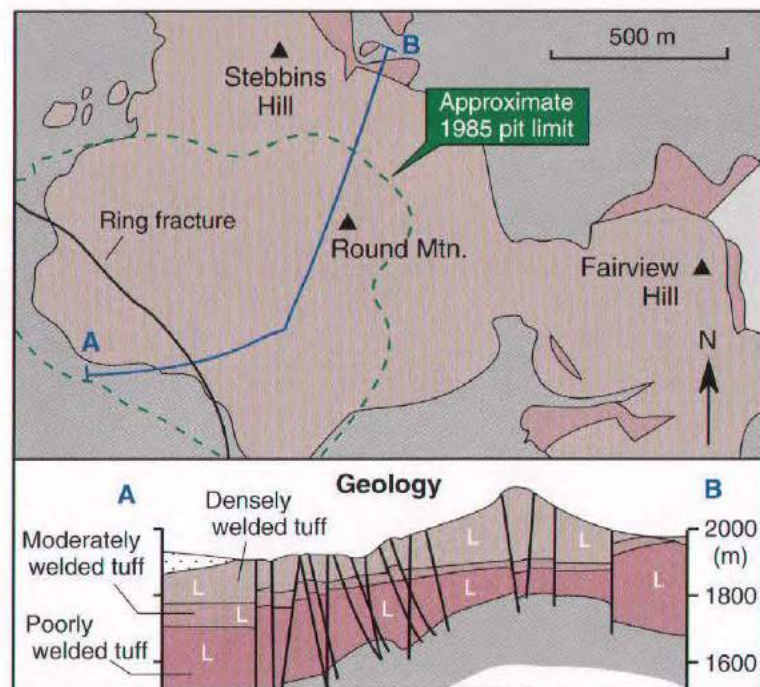


Fig. 7.1 Geologic map and section of the Round Mountain mine area [43,56]. The basement units consist of Paleozoic metasedimentary rocks, Cretaceous granites, and Tertiary intrusions and ash-flow tuffs. Unconformably overlying the basement are the host ash-flow tuffs. NW-striking faults associated with a ring fracture offset the tuffs, and along with joints, provide permeable conduits through the middle and top units of welded tuffs.

The Round Mountain deposit is a large disseminated ore body (277 Mt at 1.2 g/t Au of open-pit reserves plus past open-pit production) located in southwest Nevada [43]. Most of the ore occurs in a poorly welded unit of a poorly to densely welded ash-flow tuff (Fig. 7.1) erupted at 26.7 Ma from an adjacent caldera. The ore body occurs near a caldera ring fracture [56], and both are essentially contemporaneous. Part of the ore (10 Mt) occurs locally as veins or fissure filling in cooling joints developed in welded tuff; these veins pass downwards into thick (up to 150 m) zones of lower grade stratabound ore hosted by poorly welded tuff (Fig. 7.1).

Early hydrothermal alteration resulted in district-wide pervasive propylitic alteration (Fig. 7.2). This was overprinted by quartz-sericite-adularia alteration closely related to fractures. The abundance of fractures filled by quartz and adularia within the densely welded tuff correlate with ore grade, though the bulk of the ore occurs at low grades, disseminated within the poorly welded units. Later in the evolution of the system, silicic and argillic alteration occurred at shallow depths (Fig. 7.2), accompanied locally by breccia and high- to low-angle veins of cockscomb quartz (chalcedony)-pyrite containing high-grade Au ore (though this mineralization accounts for <10% of the total Au in the deposit). Pyrite is the dominant sulfide, and electrum occurs as free grains and as inclusions in pyrite. There are also trace quantities of Au-Ag-tellurides, sphalerite, galena, chalcopryrite, pyrrhotite, arsenopyrite, tetrahedrite, pyrargyrite, marcasite and realgar; Au:Ag ranges from 1:1 to 1:20 [43].

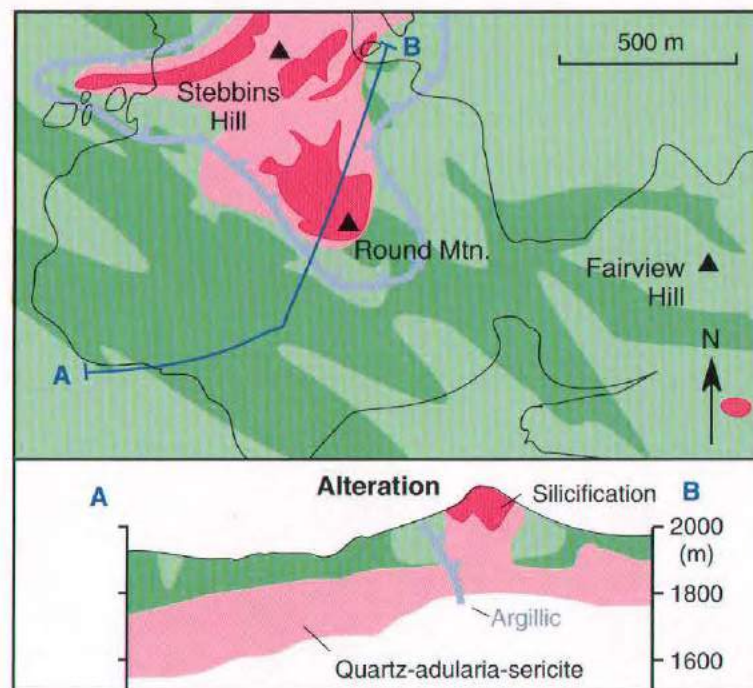


Fig. 7.2 Distribution of quartz-adularia-albite-chlorite-pyrite-calcite-epidote (propylitic) alteration, overprinted by an assemblage of quartz-adularia-sericite-calcite-pyrite (shown as quartz-sericite pink), primarily within lower poorly welded tuff. Mixed-layered illite-smectite and smectite alteration forms a large halo around surface outcrops of silicic alteration; both extend downwards as a late overprint (within hatched argillic-blue line)[43].

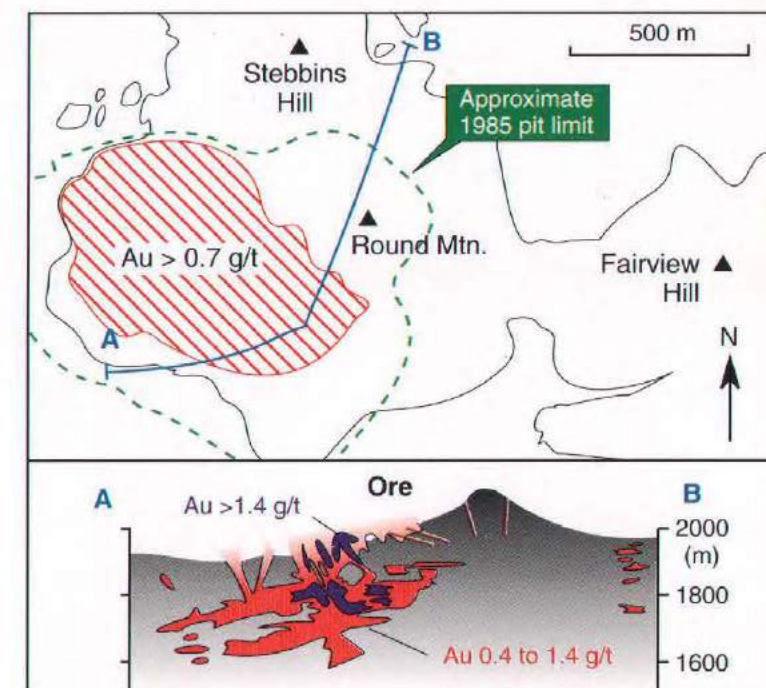


Fig. 7.3 Distribution of Au grades at the surface and in cross section at the Round Mountain mine [43]. Note the large low-grade ore body formed within the lower unit of poorly welded tuff, beneath the cap of the moderately and densely welded tuffs (Fig. 7.1). Higher grade ore occurs where fluid ascent was focused along faults and joints. The distribution of argillic and silicic alteration types is not related to ore-grade mineralization.



Fig. 7.4 Round Mountain, looking to the east (photo by Etheridge Henley Williams). Topographic high caused by silicified rocks resistant to erosion, though the zone of strong silicification is not related to Au mineralization (Figs. 7.2 and 7.3).

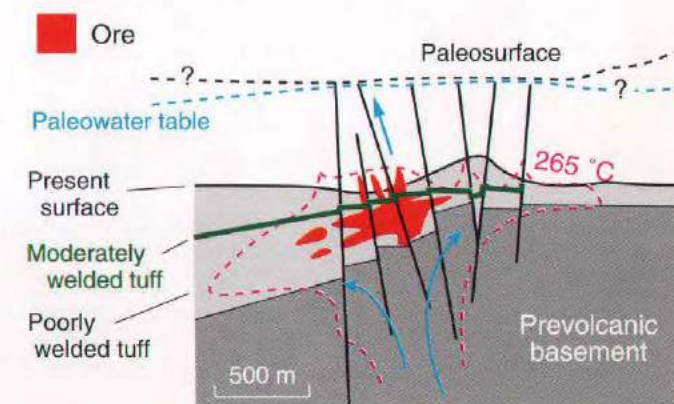


Fig. 7.5 Schematic model of fluid flow [43], with paleoisotherms mushrooming in poorly welded tuff beneath welded tuffs; subsequent fluid ascent was focused by faults and joints. The presence of adularia in fractures correlates with higher Au grades, although 90% of ore is located within poorly welded tuff which acted as an aquifer for lateral flow (blue arrows).

8) Kelian

The Kelian LS disseminated deposit is one of a number of recently discovered epithermal prospects associated with Tertiary volcanic rocks in a 400-km long belt extending in a northeast trend across central Kalimantan, Indonesia. The local Dayak people discovered alluvial Au in 1947, and mined until 1963. The discovery outcrop was located in 1976 by a multinational exploration group during follow-up of alluvial Au occurrences, using traditional stream-sediment, panned concentrate, rock float and outcrop sampling. During the next two years, soil sampling and trenching identified areas of >3 g/t Au, though scout drilling yielded only 2 g/t Au. Exploration halted until 1979, when higher Au prices and a new geologic interpretation provided new impetus for further drilling. A further exploration license was granted in 1986 that allowed detailed investigations including soil sampling, trenching and augering, ground magnetics and IP, and $>60,000$ m of drilling. This work identified a resource of >75 Mt at 1.8 g/t Au, though with poor Au recovery. Economic viability was not demonstrated until 1987-88, when higher grade mineralization with good Au recovery was discovered at depth.

Mining began in 1992 by open-pit methods; pre-construction costs totalled US\$ 27 million. The first year of mining produced 14.5 t of Au; the enlarged resource at that time was 97 Mt at 1.85 g/t Au, with 53.5 Mt mineable at 1.97 g/t, making Kelian the largest known epithermal Au deposit in Indonesia [58]. This exploration and assessment history of Kelian [58,59] illustrates how some deposits are "made", rather than simply discovered.

The alteration zone at Kelian is about 1 km in diameter, with the vertical extent of mineralization being at least 600 m, based on results from deep drilling. The deposit is associated spatially with Early Miocene diatreme and subvolcanic andesite bodies intruded into an Eocene sequence of ash-flow tuffs and sedimentary rocks (Fig. 8.1). The mineralization is genetically related to post-andesite rhyolitic magmatism [58], which gave rise to the phreatomagmatic diatreme. Hydrothermal brecciation is common in the ore zone (Fig. 8.2).

Quartz veining and silicification are poorly developed, probably because of the high paleotemperatures (about 300°C , indicating as much as 900 m of erosion above the present surface)[59]. The sequence of hydrothermal events began with propylitic alteration (stage I) followed by brecciation and veining associated with illite-adularia alteration (stage II); Au mineralization was associated subsequently with carbonate-base metal introduction (stage III). A late alteration event of kaolinite, Fe-carbonate, and cinnabar overprinted the three previous stages during a downward collapse of shallow (steam-heated?) water [59]. Pyrite (1-5%) is ubiquitous, and there is a complex vertical zonation of carbonate minerals. The fluid inclusion, isotopic and geological evidence is consistent with a magmatic component to the principal Au mineralizing fluid. This evidence, and the relatively great depth of mineralization (to >1500 m below the paleosurface) suggests that Kelian formed at the lower limits of the LS epithermal environment.

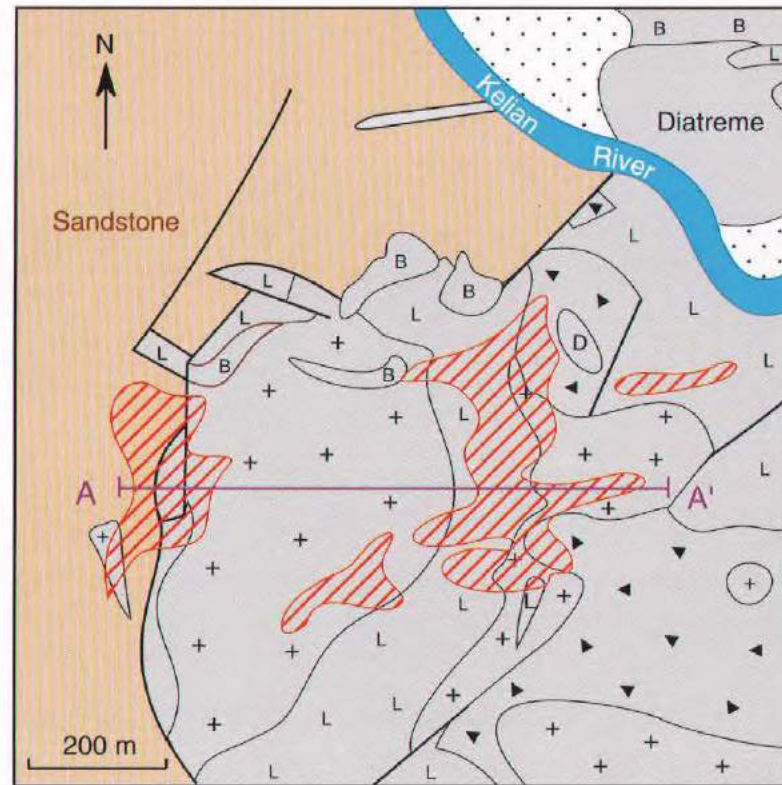


Fig. 8.1 Simplified geology of Kelian deposit [59]. The oldest units exposed are Eocene carbonaceous sedimentary rocks and various types of pyroclastic rocks, which were overlain by Miocene sandstone. These rocks were intruded by stocks and dikes of andesite, followed by rhyolite. The andesite intrusions are strongly altered, and commonly are mineralized, particularly at their contact with the host pyroclastic rocks.

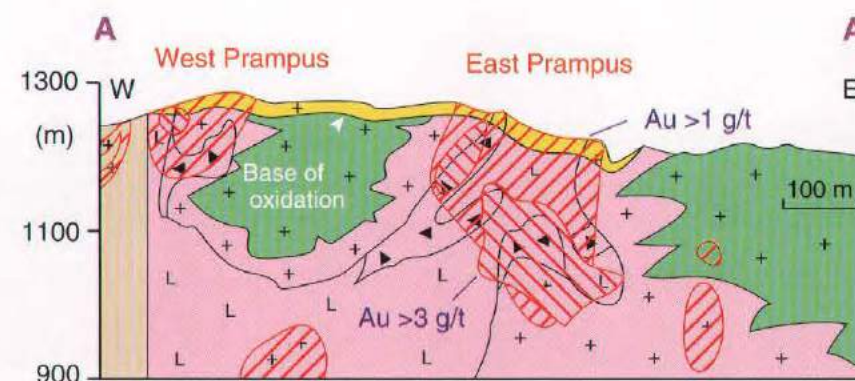


Fig. 8.2 Schematic section through the East and West Prampus ore bodies, showing geology, alteration and ore distribution [59]. Higher grade ore (>3 g/t Au, right-slanting red lines) is associated closely with abundant carbonate in veins and vugs related to stage III mineralization.



Fig. 8.3 Panning by local people in stream crossing alluvial plain of Kelian River. Such small-scale operations, commonly employing a pump to sluice water, are estimated to have produced a total of about 20 t Au since 1955, exceeding 1 t/yr at times [59]. The Kelian discovery outcrop was identified by following upstream Au anomalies determined by panning; indeed, this is still one of the best prospecting methods for Au in non-arid terrain.



Fig. 8.4 Travel by dugout canoe with outboard motor (ketingting) up Kelian River. During prospecting and assessment stages most supplies were transported by this method until 48-km long road was cut through jungle when decision was made to develop the deposit. This illustrates one of many non-geological factors that affect the logistics of mineral exploration in less-developed regions; other factors anywhere in the world include relief, vegetation, climate, etc.

High-sulfidation deposits

9) Rodalquilar

The small but well-exposed Rodalquilar HS Au deposit (10 t Au produced) is located on the Mediterranean coast of southeastern Spain and is the first example in Europe of epithermal mineralization related to a caldera [2]. Ore deposits within the Rodalquilar caldera complex occur within intracaldera ash-flow and ash-fall units and rhyolite domes (Figs. 9.1 and 9.2). Mineralization consists of LS base-metal veins and the economically more important HS Au-(Cu-Te-Sn) deposit. The latter is related temporally to andesite porphyry intrusion and is localized preferentially along ring and radial faults in the eastern wall of the nested Lomilla caldera (Fig. 9.4). Advanced argillic alteration with alunite-kaolinite-diaspore-pyrophyllite-zunyite envelopes the ore zone and extends to 300 m depth, about 200 m below the extent of Au mineralization (Fig. 9.2) Deep drilling shows a gradation to a quartz-sericite-pyrite assemblage which changes laterally to propylitic alteration. K/Ar dating of alunite and sericite indicates an average age of about 10.4 Ma for the mineralization, less than 1.0 M.y. younger than the age of the Rodalquilar caldera (age of precaldra dome = 11.1 Ma).

Zones of vuggy quartz are structurally controlled and related spatially to Au mineralization. Hypogene Au ore is hosted by hydrothermal breccias and horizontally banded black (pyrite-rich) chacedony which occurs in veins and as open-space fillings (average grade was 2 to 8 g/t). Calaverite, native Te, and pyrite are common hypogene minerals, which have been weathered at 3-4 Ma to produce native Au, Fe hydroxides and secondary Te minerals (Fig. 9.5), in addition to a widespread supergene alunite-kaolinite overprint (Fig. 9.2). Geochemical studies [2] indicate two hypogene alteration-mineralization stages: acidic wallrock alteration and leaching followed by Au mineralization under less acidic and more reduced conditions. A significant magmatic component was present in the fluid during both stages [2](Fig. 14.1), with salinities attaining 40 wt% NaCl equiv. at the depth of sericitic alteration (Fig. 9.3).

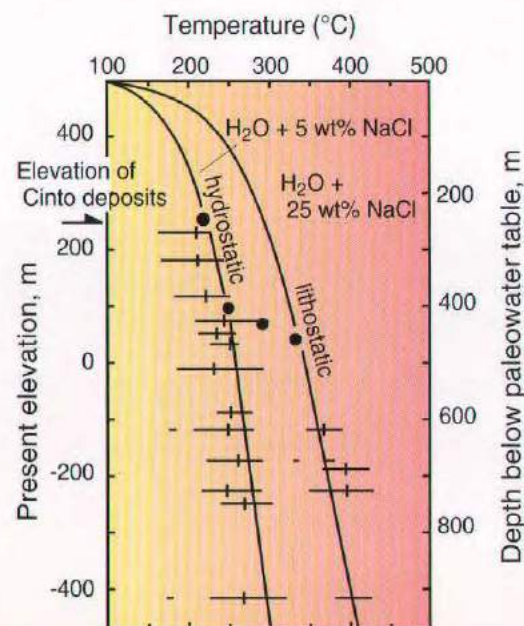


Fig. 9.3 (l.) Elevation vs. temperature diagram showing range and average of fluid inclusion homogenization temperatures (T_h) from Rodalquilar (Cinto) HS deposit and the S-isotopic fractionation temperatures (circles) for coexisting pyrite-alunite [2]. Hydrostatic boiling-point curve (for 5 wt% NaCl solution) and lithostatic pressure curve (for 25 wt% NaCl) are fitted to the data, suggesting low-salinity fluids (associated with inclusions of lower T_h values) were at hydrostatic pressures, whereas the higher temperature, higher salinity fluids (at a given depth) associated with quartz-sericite alteration were at lithostatic pressures, perhaps at paleodepths of only 600 m. The present surface is 200-300 m beneath the paleosurface.

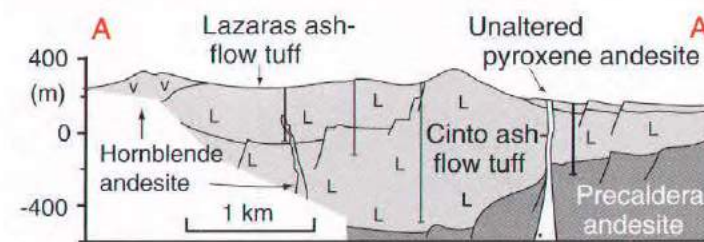
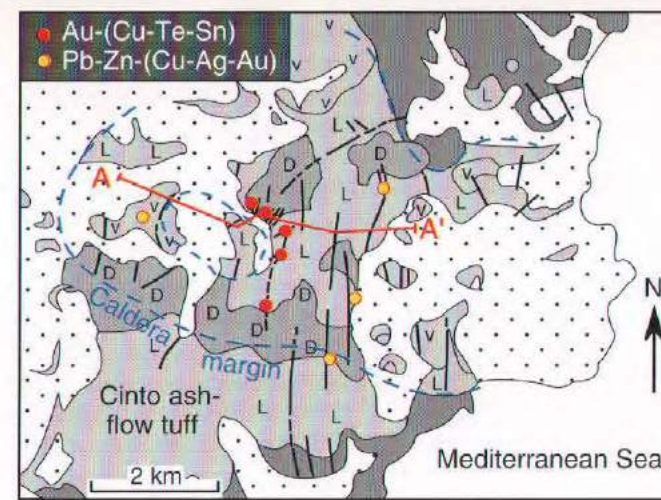


Fig. 9.1 a (t.) Generalized geologic map of the Rodalquilar caldera complex in the Cabo de Gata volcanic field, SE Spain, showing margins of the Rodalquilar and Lomilla calderas, major faults, and location of HS Au and LS base metal-Ag-Au ore bodies [2]. **b (b.)** Section through the Rodalquilar district (A-A', Fig. 9.1a), showing downfaulted basement of pre-caldra andesitic volcanic rocks, caldera fill sequences of the Rodalquilar and Lomilla calderas and intrusions of unaltered post-ore pyroxene andesite [2].

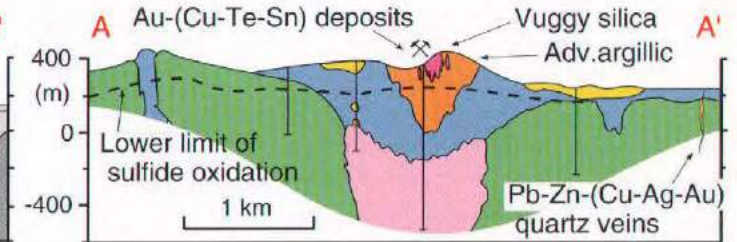
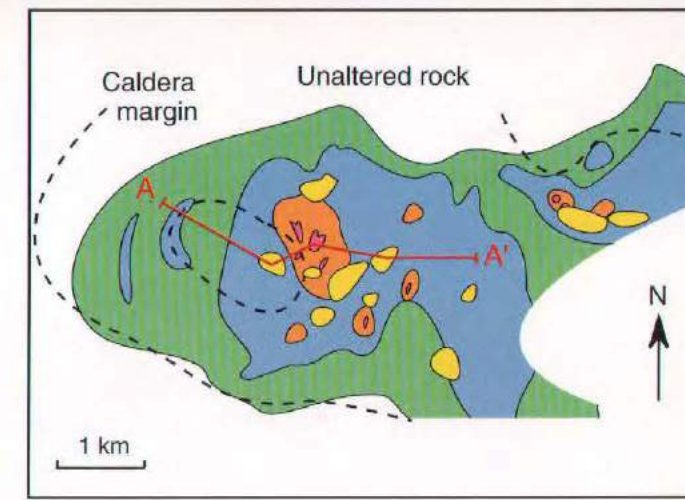


Fig. 9.2 a (t.) Schematic alteration map of the Rodalquilar caldera; only areas of most intense supergene overprint are shown. **b (b.)** Section (A-A', Fig. 9.2a) showing generalized alteration pattern, based on drill hole samples (location of holes shown). The vuggy quartz core that hosts ore has outward and downward halos of advanced argillic and argillic alteration; the argillic alteration pinches out downward at depths >350 m and gives way to quartz-sericite alteration typical of porphyry systems.

Fig. 9.4 (r.) View of eastern topographic wall of the Lomilla caldera, which is defined by arcuate scarp in middle distance. Gold was extracted from open pits and underground workings in caldera wall. The hill to left is a rhyolite dome emplaced along northern structural margin of the Rodalquilar caldera (Fig. 9.1).



Fig. 9.5 (far r.) Sample of hydrothermal vein breccia consisting of altered (vuggy quartz and advanced argillic) rhyolite (D) and earlier vein clasts (C_1) in a chalcedony matrix (C_2). Hypogene Au occurs as dark calaverite-bearing chalcedony [2].



10) El Indio-Tambo

The El Indio HS Au-Ag-Cu vein deposit is located in the Andes of north-central Chile, at elevations >4000 m. Outcropping veins were worked for a decade by pirquineros, small-scale miners, until exploration began in 1978. Since 1980, mines at El Indio and Tambo (Fig. 10.1) have produced 126 t Au, 630 t Ag and 0.32 Mt Cu, with much of the Au coming from high-grade ore (direct shipping ore of >100 g/t). Reserves at the end of 1993 were 76 t Au, 1000 t Ag and 0.61 Mt Cu at El Indio, and 45 t Au and 75 t Ag at Tambo (Cu mineralization at Tambo is rare to the depths investigated, <300 m)[32].

The local basement consists of Triassic calc-alkaline granodiorite intrusions and volcanic rocks, bounded on the east by the Baños del Toro reverse fault [32]. Further east, a series of reverse faults bound a graben filled by Tertiary andesitic lava and dacitic pyroclastic flows that range in age from 27-22, 19-11, and 6-5 Ma (Fig. 10.1). An intrusion of monzonite was emplaced at 13.3 Ma, 3 km south of El Indio. The age of mineralization is close to 7 Ma [32].

The El Indio district is associated with extensive alteration over a belt 8 km long and 3-4 km wide, elongate in the direction of the graben. Dacite to rhyolite ash-flow and ash-fall tuffs are intruded by dacite stocks; these rocks host the El Indio vein system, contained within a pervasively altered area 2 km² (Fig. 10.3). Early Cu mineralization, dominated by massive enargite veins with advanced argillic and quartz-

sericite alteration halos, was followed by quartz-Au veins with alteration halos of quartz-illite (Fig. 10.4). Tennantite, associated with Au mineralization, replaces enargite, indicating an evolution to relatively lower sulfidation states (section 14). Supergene weathering extends locally to 150 m depth, and although Cu is leached near surface, Ag is enriched while Au remains constant. Fluid inclusion data [32] indicate that both mineralization stages formed at about 250°C, with salinities <5 wt% NaCl equiv. during the Cu stage and <3 wt% NaCl during the Au stage.

A different style of mineralization at Tambo, 6 km south of El Indio, is hosted by tectonic and hydrothermal breccias (Fig. 10. 2) contained within a 5 km² area of advanced argillic alteration, consisting of barite-alunite-quartz [32]. Silver is lower in grade than at El Indio, and Cu minerals are rare. Fluid inclusion data indicate a temperature of about 230°C during Au precipitation from fluid with a salinity of <3.5 wt% NaCl.

A blanket of powdery, friable fine-grained alunite-kaolinite-cristobalite-native sulfur alteration caps higher parts of the Tambo area, and likely formed by steam-heated acid-sulfate waters [32]. Locally, at Tambo, this late-stage alteration is associated with native Au, suggesting remobilization.

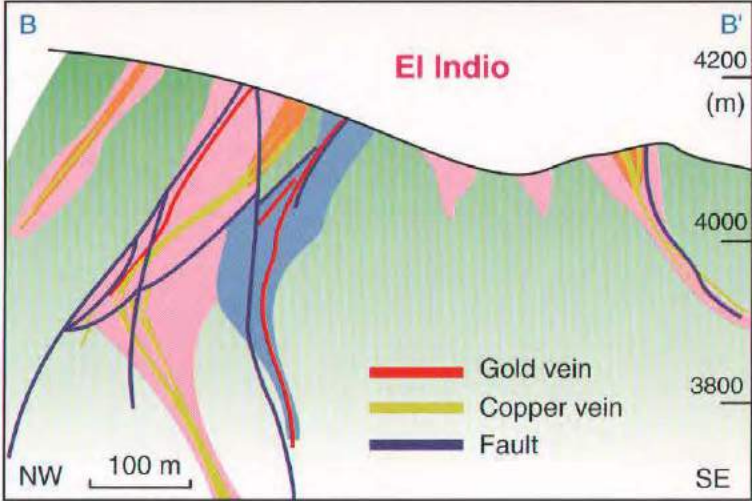


Fig. 10.3 Section of El Indio mine looking NE, showing some of the veins and associated alteration assemblages. The early enargite-pyrite veins are overlain by advanced argillic alteration (kaolinite-alunite-quartz) and have halos of quartz-sericite (kaolinite-sericite-quartz-pyrophyllite); Au content of these veins is relatively low, only a few g/t. Later cross-cutting quartz-Au veins have halos of quartz-illite alteration (illite-quartz-pyrite-pyrophyllite). Propylitic alteration envelops vein system [32].

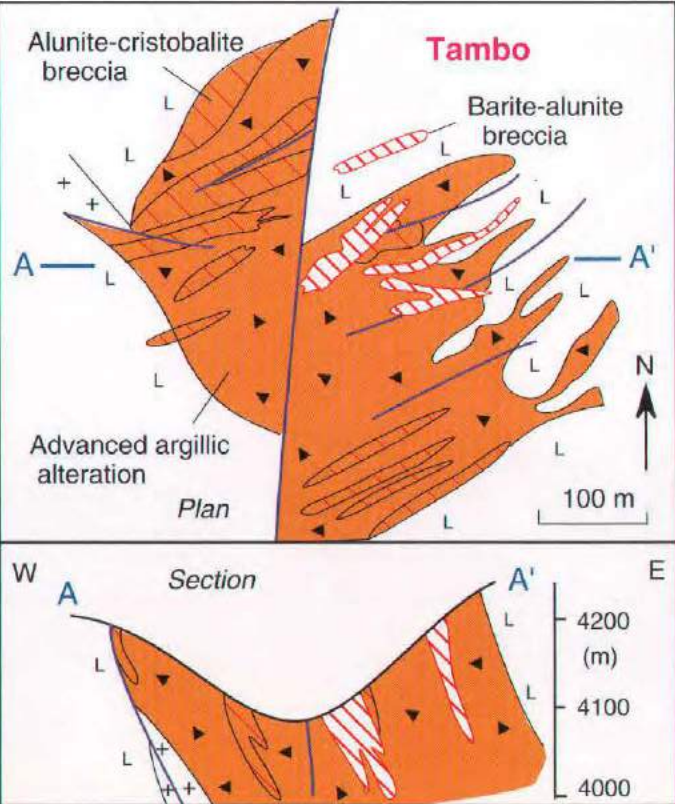
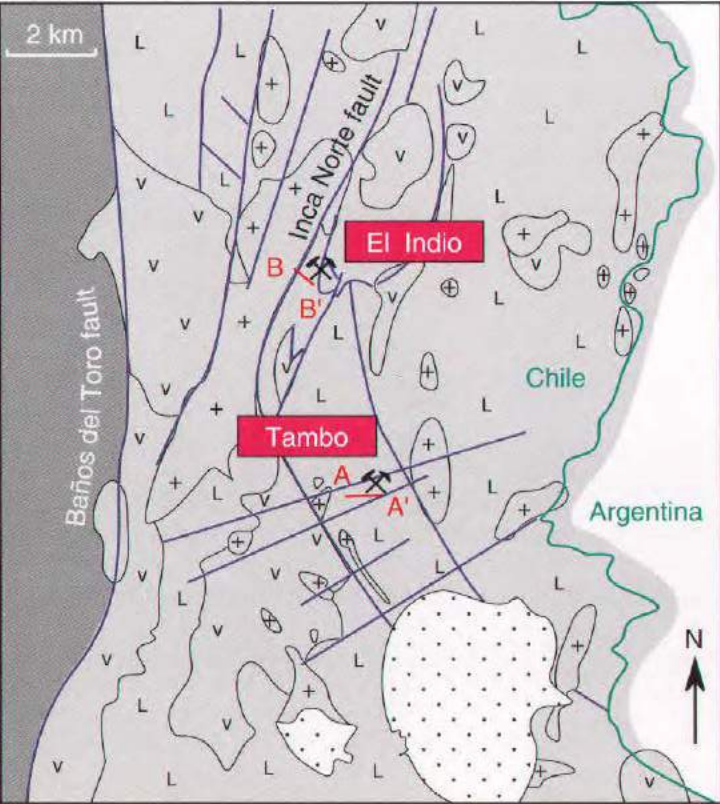


Fig. 10.1 (far L) Simplified geologic map of El Indio district showing major structures and location of El Indio and Tambo HS deposits [32,33]. A ridge underlain by a 700-m thick sequence of ash-flow tuff separates the two deposits.

Fig. 10.2 (L) Generalized surface geology and section (A-A') of the Kimberly breccia, Tambo mine [32]. Map shows location of mineralized breccia bodies; high-grade Au is hosted by barite-alunite breccia, and low-grade ore by alunite-cristobalite breccia. Kimberly breccia is up to 400 m in vertical extent and has an elongated shape that follows an ENE structural trend, parallel to district faults (Fig. 10.1); the breccia is advanced argillically altered and alteration grades outward to argillic and propylitic halos. Section (below), looking north, shows steeply dipping mineralized breccia bodies. Gold mineralization formed 200-500 m below paleosurface [32].

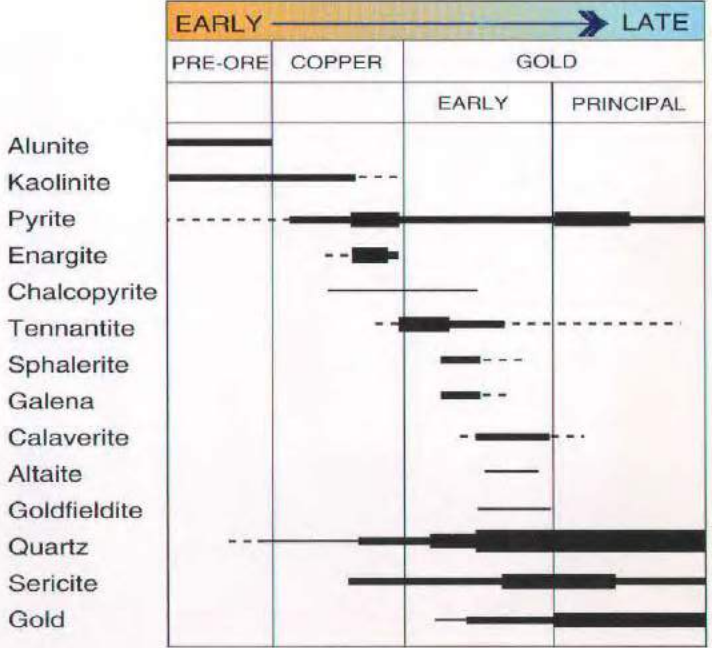


Fig. 10.4 Paragenetic sequence of hypogene vein-filling and alteration minerals at El Indio [32]. Early advanced argillic alteration was followed by deposition of enargite and, later, tennantite-Au mineralization, with formation of quartz-sericite alteration indicating less acidic conditions with time. This paragenesis is typical of HS ore deposits (e.g., Lepanto).

11) Paradise Peak

The Paradise Peak Au-Ag-Hg district is located in western Nevada. The Paradise Peak HS ore deposit was discovered in 1983 [34], and satellite HS ore bodies were identified subsequently, plus a porphyry Au stockwork to the west (Fig. 11.1). Mining ceased in 1993 after reserves were exhausted, with the production of 47 t of Au, 1255 t Ag, and 457 t Hg, from a total of 20.6 Mt of ore (averaging 3.8 g/t Au in Paradise Peak; other ore bodies had lower concentrations)[52]. The ore bodies constitute one of several major Miocene epithermal deposits in the Walker Lane belt, a zone of extension related to northwest trending strike-slip faulting along the western margin of the Great Basin [52].

The HS ore bodies are hosted by pervasively silicified (leached), welded ash-flow tuff. The stratabound ore-bearing tuff is bounded on all sides, both stratigraphically and structurally, by aquitards, mainly andesite lava (Fig. 11.2). The highest Au values occur in hydrothermal breccia which cuts the silicified tuff (Fig. 11.3), although the Paradise Peak ore extends into overlying andesite lava altered to a quartz-alunite assemblage. Where unoxidized, locally laminated sulfide ore shows a breccia texture and contains pyrite and marcasite plus minor stibnite, bismuthinite and traces of sphalerite, galena, chalcopryite, tetrahedrite, argentite, arsenopyrite and native Au, along with barite and anatase [34,52]. A lower andesite sequence hosts low-grade Au mineralization associated with sericitic alteration (Fig. 11.2) interpreted to be related to a porphyry system; quartz veinlets contain Au and Au-tellurides, including krennerite, calaverite and petzite [52]. Significant Hg mineralization, not typical in HS deposits, is attributed to late-stage collapse of steam-heated water over the HS ore body as the paleowater table descended; cristobalite alteration around the HS bodies was produced by this overprinting [52]. Unlike other HS ore bodies with similar alteration assemblages, the ore at Paradise Peak is notably low in Cu.

Detachment faulting affected the district prior to 10 Ma, after steep, normal faulting and Au mineralization were complete at 18-19 Ma [34]. All the HS deposits occur in structural duplexes, bounded by fault gouge (Fig. 11.2); as a result, some of the ore bodies were dissected and detached from their feeder zones [52]. Supergene alteration began after detachment faulting, leading to sulfide oxidation and production of oxide ore. The district-wide base of supergene oxidation coincides with the low permeability zones created by detachment faults. Where unoxidized, sulfides comprise 10-90 vol% of siliceous refractory rock; after oxidation, this rock became friable, powdery ore.

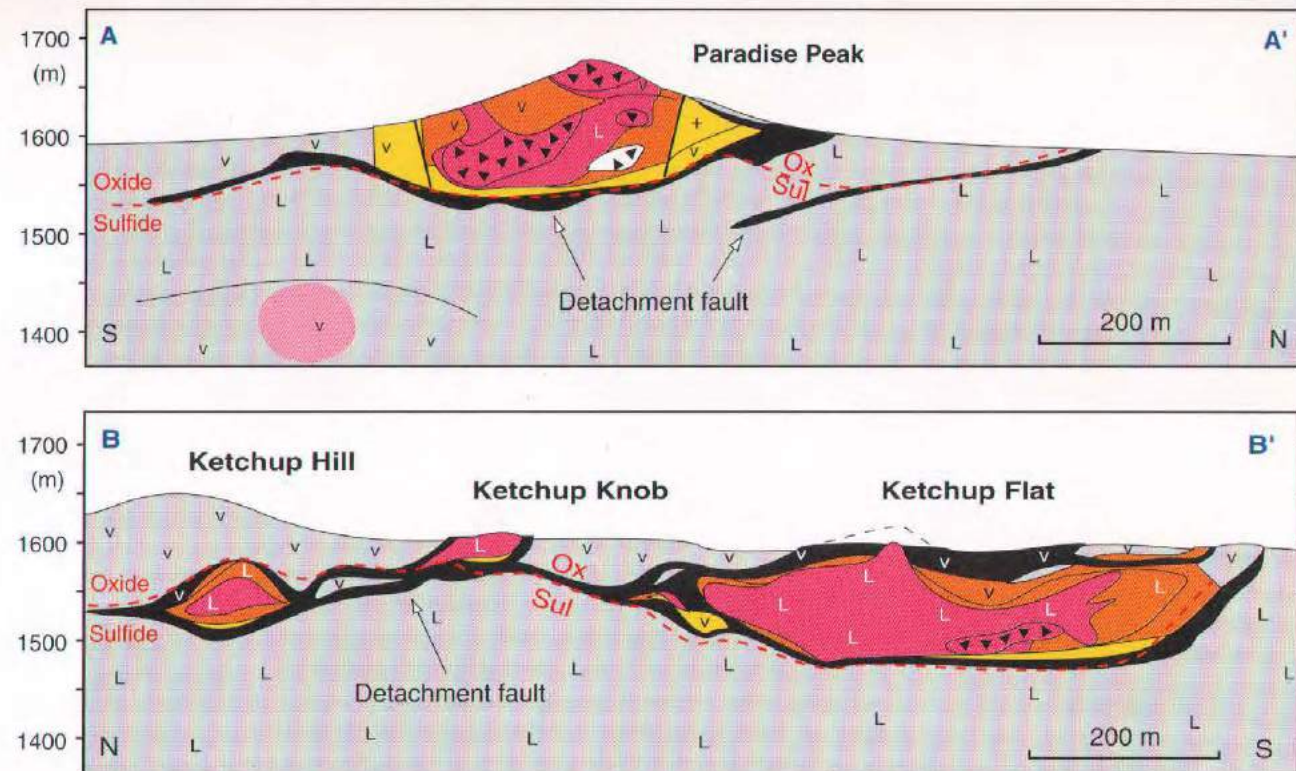
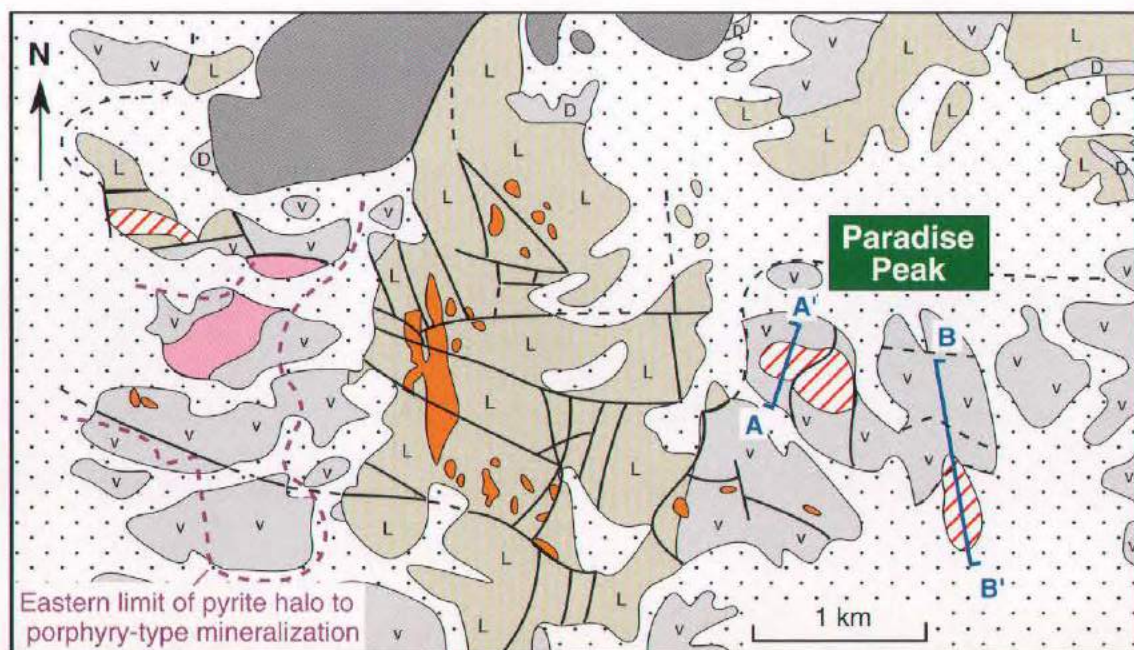
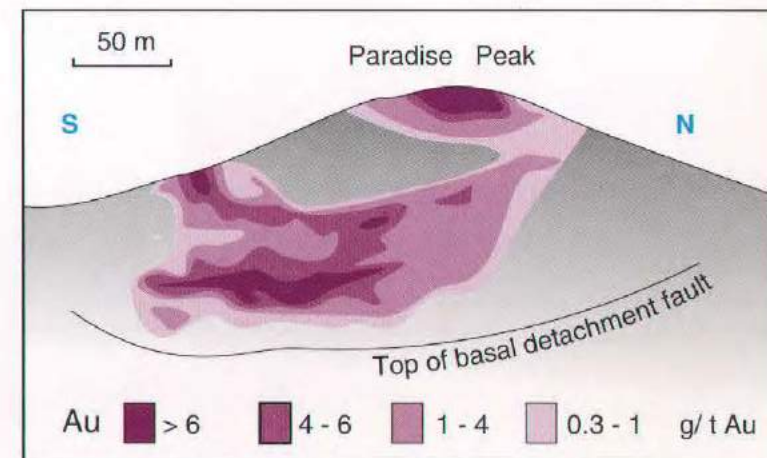


Fig. 11.2a (t.) Geologic section, looking west, through the Paradise Peak deposit (line A-A', Fig. 11.1). The ore body was dissected by a detachment fault at its base, and on its northern margin. The core of mineralization is dominantly silicified residual quartz (90-95% SiO₂) of welded tuff, and the highest grade part of the ore body consists of a silicified breccia with a black porous matrix containing Au and Ag. Outward from the ore body a selvage of quartz-alunite alteration of andesite lava grades outward to kaolinite and smectite-chlorite alteration zones. Much of the kaolinite is supergene in origin, associated with pyrite oxidation. A deep drillhole intersected a sericitic assemblage attributed to porphyry alteration. **b (b.)** Geologic section through the Ketchup Hill, Ketchup Knob and Ketchup Flat deposits (line B-B'), showing detachment-fault control on the extent of mineralization in the three ore bodies; breccia is less common than at Paradise Peak.

Fig. 11.1 (l.) Geologic map of Paradise Peak district [52]. Surface projection of HS Au mineralization is shown, as well as surface extent of quartz-alunite (advanced argillic) alteration (orange) and eastern limit of pyrite halo and quartz-sericite alteration (light pink) related to porphyry-style mineralization.

Fig. 11.3 (r.) Section, looking west, through the Paradise Peak deposit (Fig. 11.2a) showing distribution of Au grade. A close correlation exists between high Au grades and black matrix breccia; high Ag is offset to the north [52].



12) Lepanto-Far Southeast

The Mankayan district in the Central Cordillera of northern Luzon is one of the richest and most diverse mining districts in the Philippine archipelago. The 25-km² district contains several porphyry Cu-Au and epithermal precious and base-metal deposits, both HS and LS types (Fig. 12.1). The Lepanto HS enargite-Au deposit and the subjacent Far Southeast (FSE) porphyry Cu-Au deposit are closely related, both spatially and temporally. The FSE porphyry Cu-Au deposit, discovered in 1980, occurs about 600 m below the surface, below and to the southeast of the Lepanto ore body, which extends 3 km northwest at elevations between 700-1200 m (Fig. 12.2). The two deposits have combined metal contents of >550 t Au and >3.6 Mt Cu, based on a cutoff grade for FSE of Cu and Au equivalent in value to 1 wt% Cu, and 2.5 wt% Cu for Lepanto (the latter ore body constituting ~25% of the total) [11].

The district is underlain by volcanic and volcanoclastic rocks intruded to the west by Miocene tonalite. Porphyritic dacite and pyroclastic rocks of Pliocene (2.2 to 1.8 Ma) and Pleistocene (1.2 to 0.9 Ma) age predate and postdate, respectively, porphyry and HS epithermal Cu-Au mineralization. Near FSE, numerous dikes and two volcanic vents filled with Pliocene dacite have been intersected by deep drill holes. FSE ore is centered on altered and mineralized dikes and irregular intrusive bodies of melanocratic quartz diorite porphyry (Fig. 12.2). The Lepanto ore body occurs largely northwest of the porphyry deposit along the Lepanto fault and shallow flared volcanic-vent contact. The porphyry and epithermal mineralization are contemporaneous at 1.5 to 1.2 Ma (Fig. 12.4)[3].

Lepanto ore is similar mineralogically and paragenetically to many HS deposits (e.g., Rodalquilar), and includes an association with vuggy quartz enveloped by advanced argillic alteration (alunite-pyrophyllite-kaolinite). Ore is dominated by early pyrite and enargite-luzonite (type locality for luzonite) in strongly silicified breccia, and is followed by tennantite-tetrahedrite and chalcopyrite plus Au (similar to El Indio). Electrum, selenides, tellurides, and Sn-minerals are present. Fluid inclusion study of enargite [37] indicates that the temperature of formation decreased regularly from about 280° to 180°C with increasing distance from the FSE ore body, accompanied by a decrease in salinity from 4 to 2 wt% NaCl equiv., indicating progressive entrainment of groundwater as the mineralizing fluid flowed northwest, away from the FSE intrusion along the Lepanto fault.

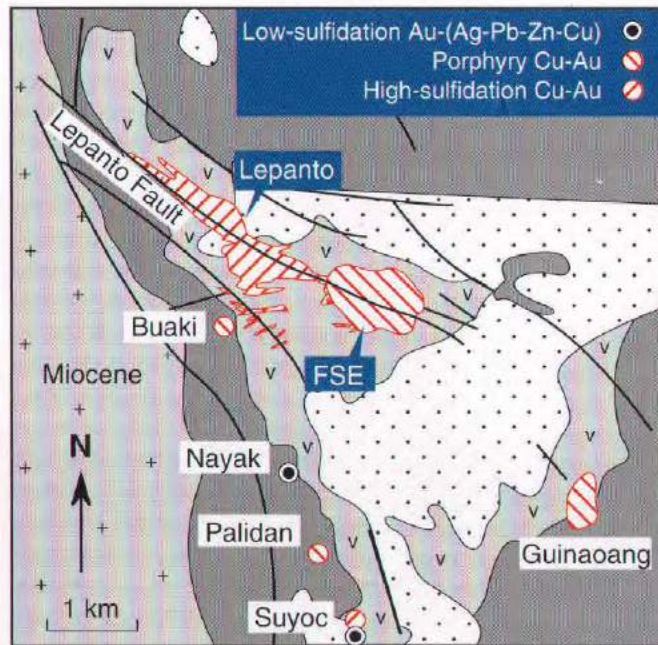


Fig. 12.1 Simplified geology and location of deposits in Mankayan district. Outlines of economically most important deposits, FSE (Far Southeast), Guinaoang and Lepanto are shown as surface projections [13]. Lepanto ore body is controlled by NW-trending Lepanto fault and subsidiary E-W branch faults in foot- and hanging walls.

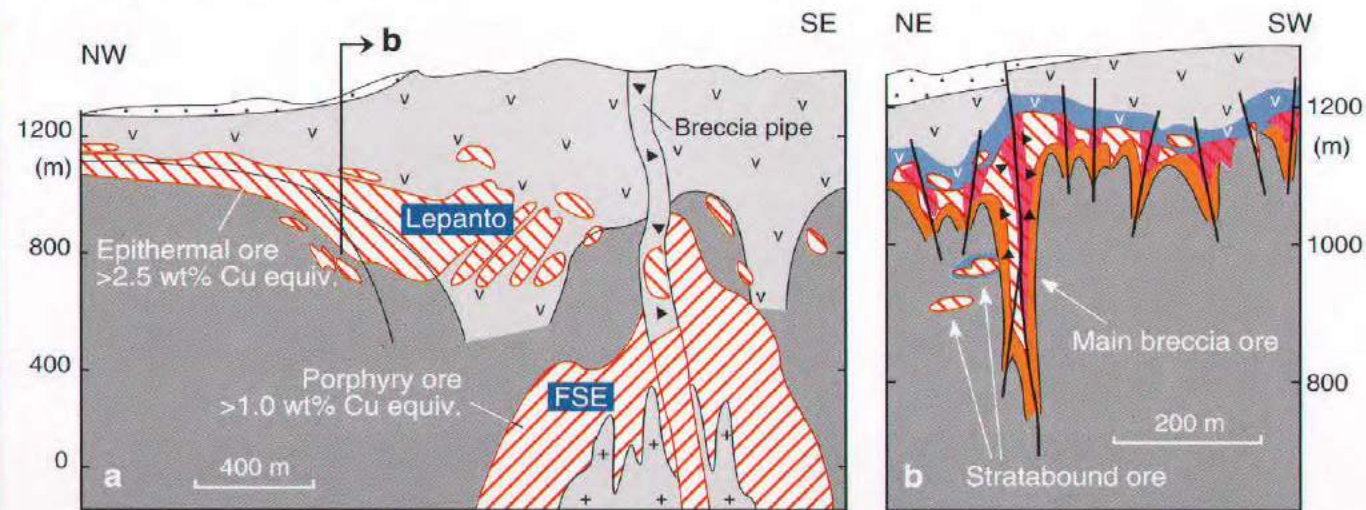


Fig. 12.2a Schematic section of Lepanto and FSE deposits in plane of Lepanto fault [13], showing major lithologic units and outline of epithermal enargite-Au and porphyry Cu-Au mineralization. Quartz diorite intrusion has concentric halo of potassic alteration overprinted by illite-chlorite assemblage associated with quartz-illite and anhydrite-pyrite (chalcopyrite-bornite) veins. Outward, particularly beyond upper part of FSE deposit, alteration grades into argillic (kaolinite) and advanced argillic (alunite-pyrophyllite-diaspore) assemblages, subeconomic in grade, that mark transition to Lepanto deposit [3]. Enargite veins locally crosscut FSE porphyry-style alteration and mineralization. A breccia pipe containing both styles of mineralization cuts through FSE deposit and overlying dacite [11]. **b** Schematic section through Lepanto deposit (see 12.2a) showing vuggy quartz and silicified fault-breccia as hosts to ore, with halo of advanced argillic alteration grading out to argillic and propylitic assemblages. Ore body and alteration halo flare where Lepanto fault intersects unconformity at base of dacite pyroclastic rocks; stringer ore associated with splay faults and stratabound ore occur locally in basement and overlying pyroclastic rocks.



Fig. 12.3 View north from Palidan (1200 m elevation; Fig. 12.1); FSE and Lepanto ore bodies lie beneath flat hills (elevation 1100 m) in distance; total relief to valley in far distance is nearly 1000 m. It is likely that similar topographic relations existed during volcanic and hydrothermal activity about 1.5 M.y. ago, although at somewhat higher elevation, accounting for NW hydraulic gradient (Fig. 12.2a)(photo by A. Arribas, Sr.).

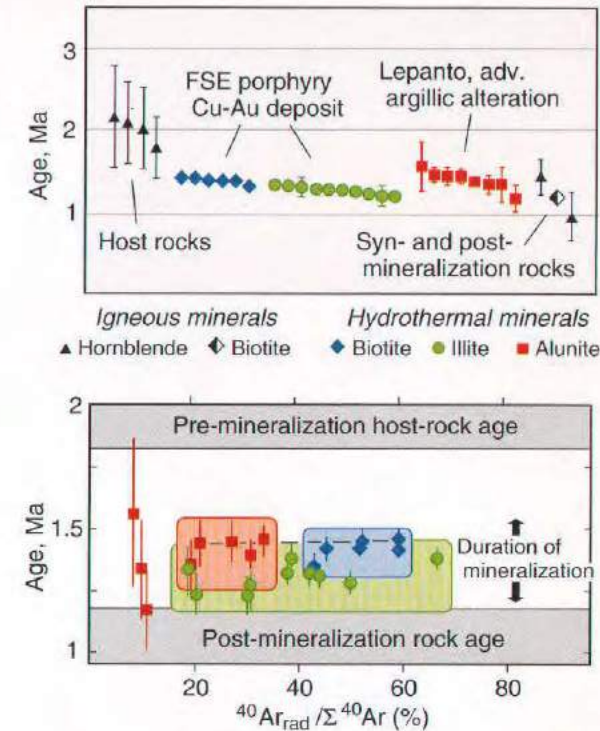


Fig. 12.4 a (t.) Radiometric (K/Ar) ages (Ma) for mineral separates from fresh and hydrothermally altered igneous rocks associated with Lepanto and FSE deposits (error bars show 95% confidence interval)[3]. Ages of host rock and post-mineral volcanic rocks are 2.2 to 1.8 and 1.2 to 0.9 Ma, respectively, and bracket all hydrothermal mineral ages. **b (b.)** K/Ar age versus percent of radiogenic ⁴⁰Ar in total ⁴⁰Ar released from hydrothermal minerals. The two deposits are contemporaneous within <0.3 M.y. period. Excluding alunite samples with low radiogenic Ar (impure mineral separates), average ages of biotite from FSE potassic alteration and alunite from advanced argillic alteration (samples from FSE cap through to the distal extent of Lepanto alteration, 4 km NW of FSE) agree at 1.40 Ma. In contrast, average age of illite samples associated with later quartz-anhydrite-sulfide veins overprinting FSE potassic alteration is 1.30 Ma, though there is some overlap when error bars are considered. Thus, advanced argillic host to vuggy quartz and contained HS epithermal ore body formed synchronously with porphyry intrusion, potassic alteration and magmatic degassing [3]. Enargite and Au deposition may have been contemporaneous with later illite overprinting of porphyry system.

13) Exploration

Epithermal deposits have been discovered in regions previously thought to be unprospective for this type of ore deposit, showing that favorable geologic environments (section 2) of any age may be prospective. The character and distribution of ore and wall rock alteration are the most varied features of epithermal ore deposits, and contribute to making each unique. The variations are controlled mainly by the hydrology of the system, and the nature of the hydrothermal fluid: neutral pH and reduced (LS) or acidic and oxidized (HS). Tectonic and geologic setting influence the relief and hence hydrology of the system, and its changes with time.

For Au exploration it is necessary first to establish the character of the tectonic and geologic environment at the time of hydrothermal activity, including paleorelief. In many cases this should be possible by careful documentation of the volcanic and sedimentary rock types present in the region, and interpretation of volcanic facies [10,36]. The likely forms of mineralization then may be assessed on the basis of the preliminary hydrologic model that is developed [49,62]. Once the volcanic environment has been identified, the types, distribution and zoning of hydrothermal alteration, coupled with observations of vein textures and mineralogy, allow a judgement to be made on several important aspects, including: 1) the chemical nature of the system, HS or LS [60], 2) the level at which the system is exposed (i.e., the depth of erosion), 3) the nature (Fig. 4.1) and location of paleoconduits, and 4) the probable controls on ore distribution.

Simple but nevertheless important field observations include: 1) recognition of sinter (which is distinct from, e.g., silicified laminated sediments), as it is evidence for paleosurface and identifies the prospect as LS; 2) identification of temperature-sensitive minerals, which indicate the approximate level of erosion and paleoisotherms (i.e., direction to paleoconduits); 3) recognition of bladed calcite (commonly infilled or replaced by quartz) in LS veins, which resulted from boiling; and 4) identification of the style of acid-sulfate alteration, e.g., steam-heated alteration occurring as a blanket or overprint in LS (and HS) systems, versus the advanced argillic alteration that forms halos to vuggy quartz-hosted ore in HS systems. Interpretation of geochemical and geophysical results commonly rest on a correct understanding of the environment of mineralization. For example, geochemical anomalies of Au, Ag, As, Sb, etc., will not be associated with steam-heated alteration (Fig. 13.2), as this forms over the water table (though Hg, being volatile, may be anomalous). By contrast, there may be Au anomalies associated with HS advanced argillic alteration.

The range of exploration methods is limited (Table 13.1), and their relative importance at different stages of exploration varies (Table 13.2; Fig. 8.4). Once a prospect is identified, most commonly on the basis of geologic observations, a combination of geologic, geochemical and geophysical methods is used to construct a model for the location of a possible ore body, which can then be tested by drilling. Any single method is simply a tool; none is infallible, and applicability must be determined on a case-by-case basis.

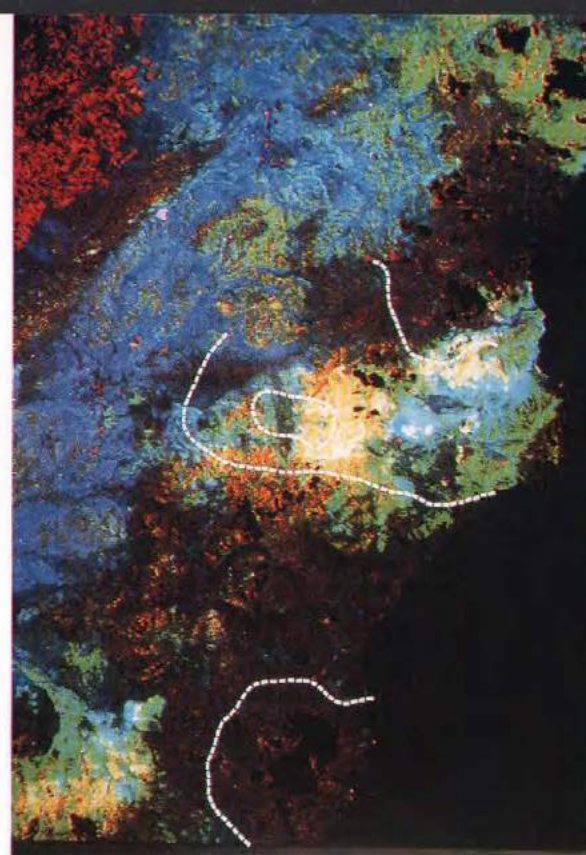


Fig. 13.1 (l.) Landsat Thematic Mapper color-ratio-composite image of Cabo de Gata volcanic field [2]; outlines shown of Rodalquilar and Lomilla calderas (Fig. 9.1) and Los Frailes caldera (south). Combination of band ratios results in white-yellow colors in areas of argillic or advanced argillic outcrop (Fig. 9.2), due to hydroxyl and sulfate minerals; alteration zoning is clearly distinguished and prospective areas readily identified. Such remote sensing is generally not applicable in areas of extreme relief or in tropical, heavily vegetated regions.

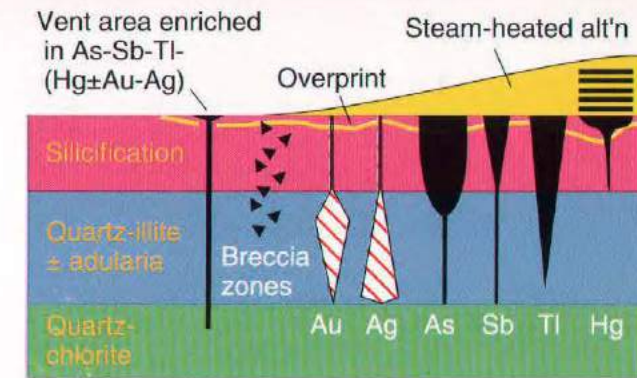
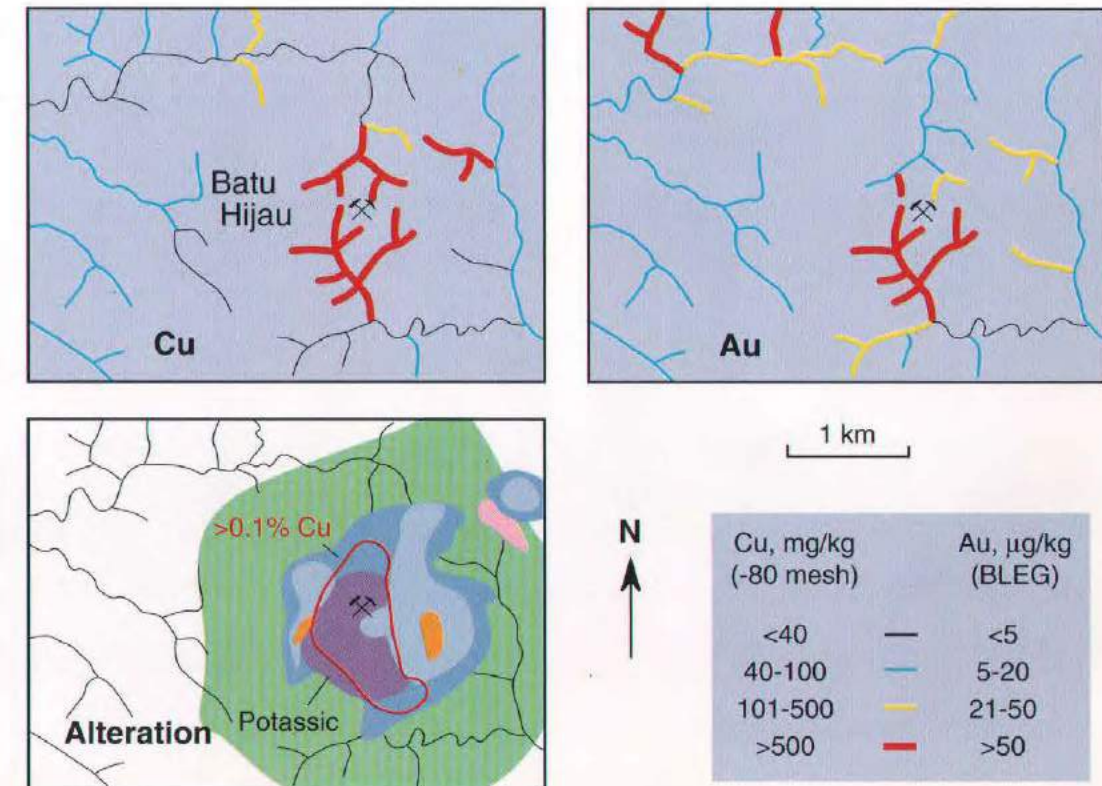


Fig. 13.2 Schematic section showing typical variation of LS epithermal indicator elements as function of depth below paleosurface, and relative association of alteration zoning with depth [6]. Volatile elements (Hg, Tl) show largest anomaly where paleotemperature was lowest, i.e., at paleosurface. Sinters may be anomalous in As, Sb, Tl, and locally Au-Ag, in addition to Hg, whereas areas of steam-heated alteration commonly show enrichment only of Hg, unless previous mineralization is overprinted. Thus interpretation of rock geochemical anomalies must include knowledge of alteration style (distinguishing LS from HS, and recognition of overprinting) and erosion level.

Fig. 13.3 (r.) Stream sediment survey results for Batu Hijau porphyry Cu-Au deposit, Sumbawa, Indonesia [39]. Initial search in region, not recognized as prospective for porphyry deposits, was for Au. Survey shown here was designed to follow-up initial observations of Cu sulfides in float, and Au in BLEG (Bulk-Leach Extractable Au)[63] samples and Cu in -80 mesh silts; two watersheds were identified as having anomalous values. Subsequent reconnaissance work upstream along tributaries revealed weak Cu mineralization in outcrop and highly anomalous Cu and Au in float. Alteration mapping and bedrock geochemistry (> 0.1 wt% Cu contour shown) then outlined the extent of surface mineralization; drilling has proven a 1994 resource of 330 Mt at 0.8% Cu and 0.7 g.t Au. Advanced argillic alteration outcrops along ridges, and is underlain by argillic alteration.



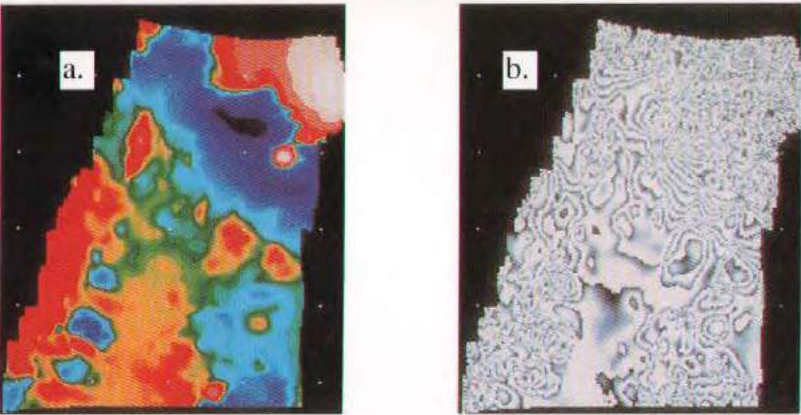


Fig. 13.4 Aeromagnetic survey at the Ohui goldfield, Coromandel Peninsula, New Zealand. Gold occurrences are centered on 3.5x1.5-km area of altered Miocene to Pliocene andesite and rhyolite lava and pyroclastic rocks, covered by recent sediments and tephra. **a** Magnetic intensity image; strongly magnetized rocks surround NNE trending corridor of low magnetic response (100-m line spacing and 60-m mean terrain clearance [27]). **b** “Alteration” filter image, highlighting areas of magnetite destruction. **c** Simplified interpretation of magnetic image, in relation to known Au prospects, alteration, and an area of high resistivity (Box resistor; defined by controlled-source audiomagneto-tellurics). Structures are clearly distinguishable; altered rocks cover a much larger area than that indicated by surface mapping.

Table 13.1 Relative importance of different methods at different stages in exploration (generalized)

Method	Regional scale	Project scale	Prospect scale
Geology	★★★★	★★★★	★★★★
Geochemistry			
stream sediment	★★★★	★★	
soil		★★★★	★★★★
alteration mapping	★	★★	★★★★
Geophysics			
remote sensing	★★	★	
magnetics	★★	★	
induced polarization			★★
resistivity		★	★★
electromagnetics		★	★★
radiometrics	★★	★	
gravity		★	

- Strongly magnetic volcanic rocks
- Reversely magnetised volcanic rocks
- Mapped alteration zone
- Alteration zone interpreted from air magnetics

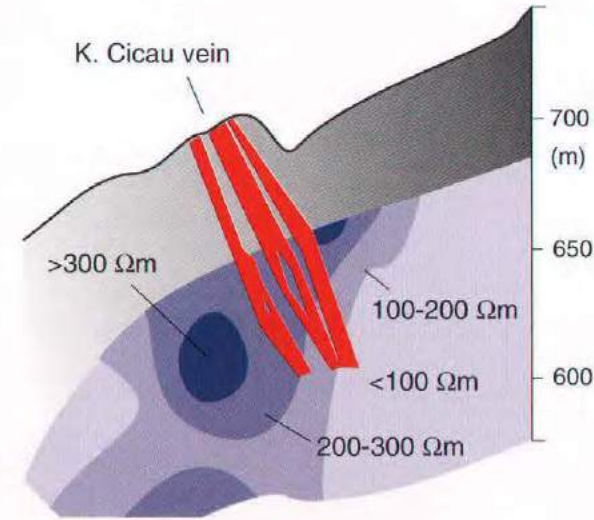


Fig. 13.5 Apparent resistivity pseudosection (dipole-dipole, $a=40$ m) across the Kubang Cicau vein (red), Gunung Ponkor LS Au-Ag deposit, west Java [5]. Prospect was identified initially by a variety of methods, including historical base-metal anomalies; Au assays of stream sediment, float and outcrop samples; mapping to define vein outcrops and extent and zoning of alteration. Induced polarization (IP)/resistivity surveys then helped define strike extent of steeply dipping, resistive quartz veins; this was used along with trenching to site scout drill holes.

Table 13.2 Exploration methods for epithermal gold deposits

Method	Response
Geology	Recognition of veins and hydrothermal alteration patterns may directly locate mineralization. Vein texture and mineralogy may indicate erosion level. Geologic understanding is basis for interpreting all other data.
Geochemistry	SHALLOW: Alteration commonly related to vapor transport, so only volatile elements may be present. Anomalies of low concentration may be widespread, unrelated to economic deposits. Indicator elements important (Fig. 13.2). Anomalies commonly obscured by younger cover. DEEP: Anomalies related to mineralized structures.
Geophysics	
1) Magnetics	Effectiveness depends on magnetic properties of rocks. Regional-scale surveys may aid geological mapping, and define areas of hydrothermal alteration (magnetite destruction). Structures may be traced from unaltered areas into areas of alteration. Prospect-scale surveys typically ineffective.
2) Induced Polarization	SHALLOW: Prospects with abundant pyrite in hydrothermally altered cap rocks have anomalies unrelated to ore mineralization. May be more useful on regional scale than on prospect scale. DEEP: Contrasts in sulfide content may be a useful guide to mineralization.
3) Resistivity	SHALLOW: High-resistivity silicified caps may occur over low-resistivity clay alteration, obscuring economically significant features on prospect scale. DEEP: Narrow high-resistivity bodies or linear features reflect silicification or veins, typically surrounded by low-resistivity clay alteration. In active systems, hot water will be detected as areas of low resistivity.
4) Electro-magnetics	SHALLOW: Conductive clay-altered zones may be detected; may detect contacts with low-conductivity zones, e.g., silicified areas or veins. Can test below unaltered cover. DEEP: Conductivity contrasts between veins and wall rocks may be detected. In active systems hot water will be detected as areas of high conductivity.
5) Radiometrics	Will detect outcropping radiogenic sources. Variations in K content detected (adularia and illite alteration recognized as positive anomalies). Altered K-rich host rocks may show K depletion.
6) Gravity	Surveys detect anomalies related to a) changes in density of host rocks due to alteration, b) basement topography, and c) intrusions (if of different density than host rock). Depending on host rocks, alteration may cause positive or negative anomaly.

Note: SHALLOW implies limited erosion, DEEP erosion of, say, >400 m.

14) Origin

The study of epithermal Au deposits and their active analogues (Fig. 1.1) has resulted in the identification of critical parameters of hydrothermal systems, including ore deposition processes [20,25]. LS ore deposits form from neutral-pH, reduced waters close to equilibrium with host rocks [14]. The waters are low-salinity (<1 wt% NaCl equiv.) but relatively gas-rich (1-2 wt% CO₂, subordinate H₂S)(Fig. 14.1), and are meteoric-water dominant; however, there is evidence [14,20,38] for an early magmatic component that may also introduce ore metals to the system. Boiling is the critical process to deposit high concentrations of Au (Fig. 14.2). In contrast, HS deposits have a significant magmatic component throughout their history (Fig. 14.4), particularly in the early, leaching stage when the fluid is most acidic and oxidized (Fig. 14.5). The process responsible for Au deposition is not understood since the ore fluid has not been characterized fully (Fig. 14.3), though ore minerals (Fig. 10.4) and hence ore fluid (Fig. 14.3) evolve with time (Fig. 14.5), to less acidic and more reduced conditions.

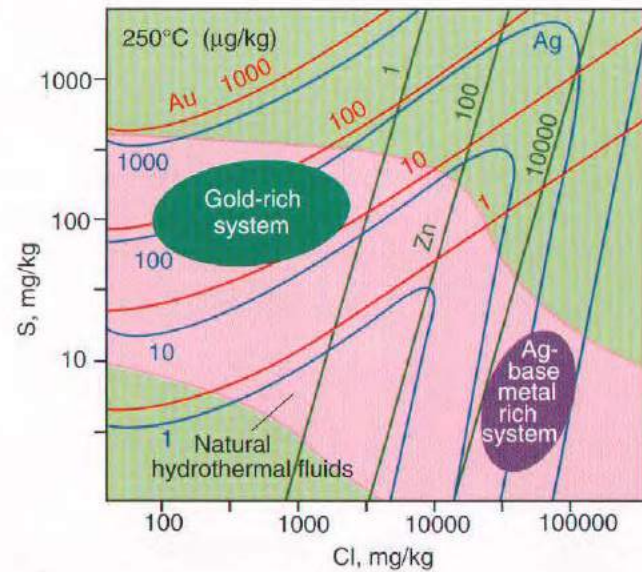


Fig. 14.1 (l.) Solubility of Au, Ag, and Zn as a function of S and Cl concentration at pH and redox of LS mineral assemblages [24]. Cl-poor solutions typical of Au-rich LS ore deposits [19] transport Au as bisulfide complexes, but cannot transport much chloride-complexed base metals [24].

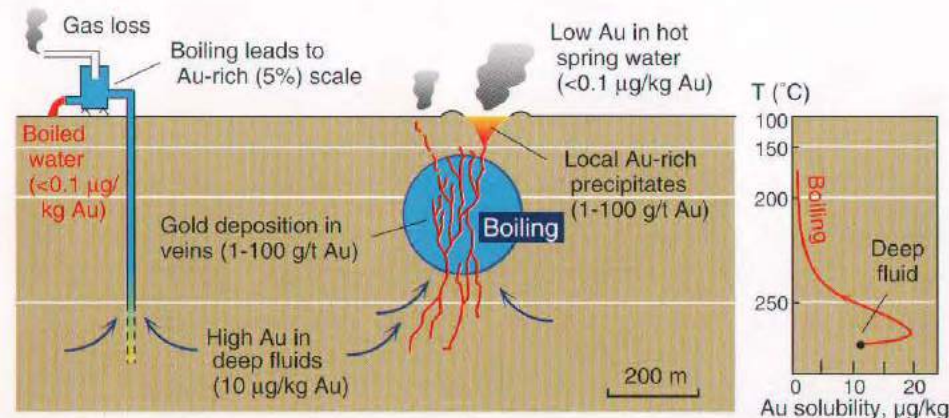


Fig. 14.2 Gold-rich scale in geothermal pipes [7] indicates that Au concentration in deep Broadlands waters (prior to boiling and gas loss) is about 10 µg/kg. In contrast, hot spring waters contain <0.1 µg/kg Au; thus, Au precipitates during ascent and boiling. This is consistent with calculations indicating decrease of Au-bisulfide solubility as H₂S and CO₂ are lost from liquid to vapor [46].

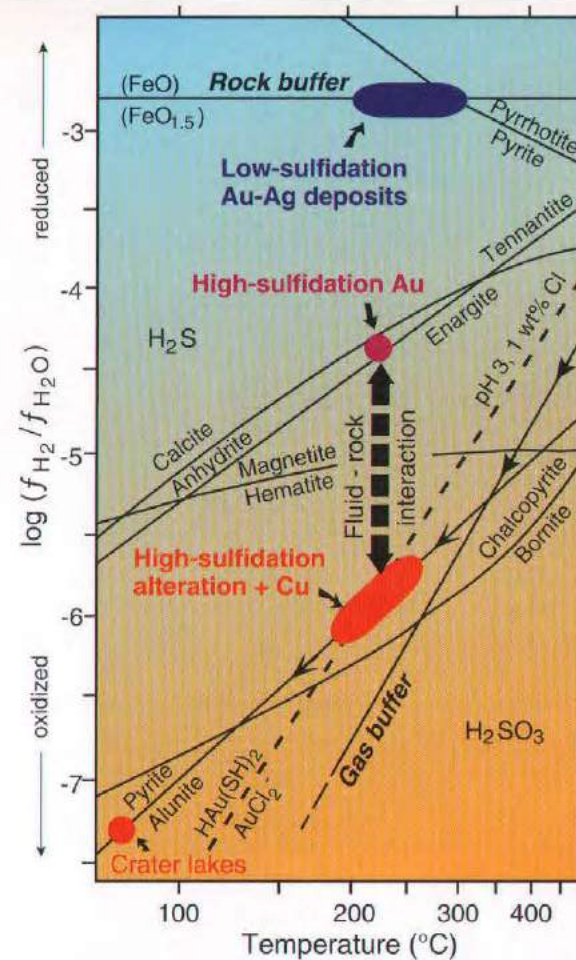


Fig. 14.3 (t.) Redox potential versus temperature for fluids limited by the oxidized volcanic gas buffer and reduced rock buffer [14,21]. HS leaching and alunite-pyrite alteration occur under oxidized conditions; enargite deposition also occurs at relatively oxidized states. Gold deposition is typically later in many HS deposits, associated with relatively reduced conditions (near tennantite stability), possibly caused by waning of oxidized and acidic magmatic component, and/or result of water-rock interaction. In contrast, LS fluids are reduced; oxidized magmatic components (Fig. 3.2) react with the host rock, and are neutralized beneath epithermal environment [14]. Gold is transported as a bisulfide complex under relatively reduced conditions (above dashed line); under relatively oxidized conditions (or at a low pH or high salinity), Au-chloride complexing dominates.

Fig. 14.5 (r.) Model for evolution of HS ore systems [1]. **a** In early stage, degassing magma releases vapor rich in SO₂, H₂S, HCl, which either discharges from high-temperature fumaroles (Fig. 1.3), or is absorbed by shallow meteoric water (Fig. 14.4); the latter forms hot, acidic solution that leaches host rock and leaves a permeable vuggy quartz conduit for later ore fluids. **b** Stable isotopic and fluid inclusion data are insufficient to determine if ore fluid is low-salinity, metal-rich magmatic vapor absorbed by meteoric water (feasible based on metal concentrations in fluid discharging from erupting volcanoes [20]), or if ore fluid is hypersaline and diluted by meteoric water (as indicated by results for Julcani [12]). Both scenarios may occur in different situations, accounting for differences among HS deposits.

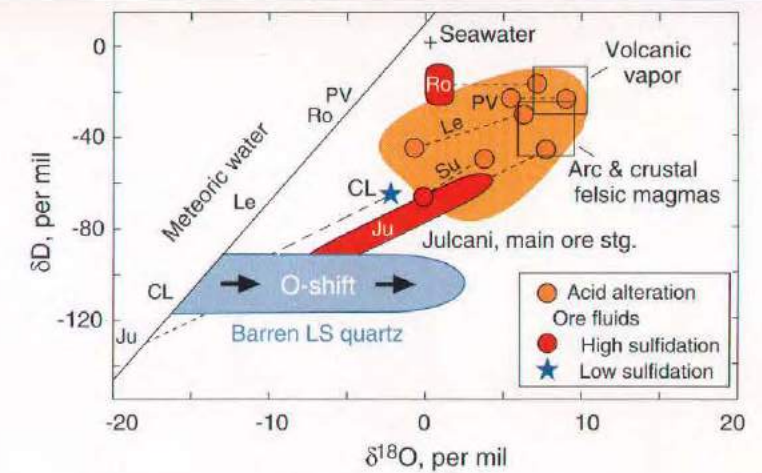
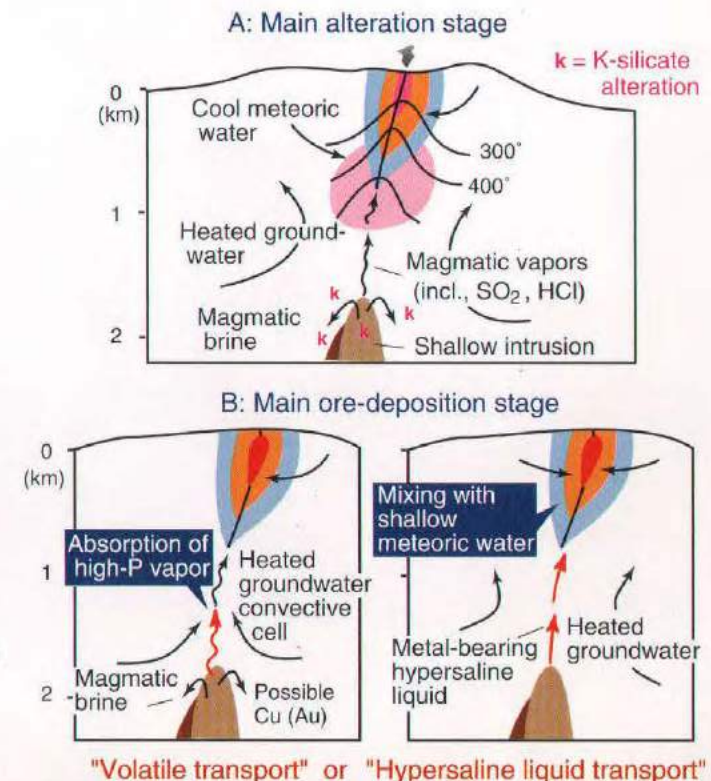


Fig. 14.4 Variation in isotopic composition of hydrothermal water at different stages in HS ore deposits [1]. Early acidic fluid dominated by magmatic water (volcanic vapor absorbed by shallow meteoric water; Fig. 14.5) leaches host rock. In systems where later ore fluid is characterized (Julcani, Ju; Summitville, Su; Rodalquilar, Ro), there is evidence for increased meteoric water component. At Julcani, ore fluid is saline (to 20 wt% NaCl equiv.; i.e. the magmatic endmember is hypersaline), but at Lepanto (Le) the enargite ore fluid contains 4 wt% NaCl equiv. In LS deposits, gangue quartz indicates hydrothermal fluids are O-shifted meteoric water in origin, though in some deposits (Hishikari, Comstock Lode; CL), there is evidence for a large magmatic water component associated with the highest grade Au ore [38], perhaps caused by episodic pulses of magmatic fluid [20].



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Typical texture of LS epithermal gold vein, McLaughlin mine, California (natural size) (photo by Y. Matsuhisa).

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