

Chapter 7

Exploration for Epithermal Gold Deposits

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Abstract

The successful exploration geologist uses knowledge of geologic relationships and ore-deposit styles, tempered by experience, to interpret all information available from a given prospect in order to develop an understanding of its mineral potential. In the case of exploration for epithermal gold deposits, this understanding can be augmented by familiarity with active hydrothermal systems, their present-day analogues. Just as geological skills and exploration experience are the defining elements of a philosophy of exploration, the needs of a company determine, as much as the funding and skills available, which level of exploration it pursues and where: grassroots, early-stage or advanced targets. Epithermal gold deposits have size, geometry, and grade variations that can be broadly organized around some genetic classes and, therefore, influence the exploration approach or philosophy.

Nearly 80 years ago, Waldemar Lindgren defined the epithermal environment as being shallow in depth, typically hosting deposits of Au, Ag, and base metals plus Hg, Sb, S, kaolinite, alunite, and silica. Even before this, Ransome recognized two distinct styles of such precious-metal deposits, leading to the conclusion that the two end-member deposits form in environments analogous to geothermal springs and volcanic fumaroles, which are dominated by reduced, neutral-pH versus oxidized, acidic fluids, respectively. The terms we use are low- and high-sulfidation to refer to deposits formed in these respective environments. The terms are based on the sulfidation state of the sulfide assemblage. End-member low-sulfidation deposits contain pyrite-pyrrhotite-arsenopyrite and high Fe sphalerite, in contrast to pyrite-enargite-luzonite-covellite typifying high-sulfidation deposits. A subset of the low-sulfidation style has an intermediate sulfidation-state assemblage of pyrite-tetrahedrite/tennantite-chalcopyrite and low Fe sphalerite. Intermediate sulfidation-state deposits are Ag and base metal-rich compared to the Au-rich end-member low-sulfidation deposits, most likely reflecting salinity variations.

There are characteristic mineral textures and assemblages associated with epithermal deposits and, coupled with fluid inclusion data, they indicate that most low-sulfidation and high-sulfidation deposits form in a temperature range of about 160° to 270°C. This temperature interval corresponds to a depth below the paleowater table of about 50 to 700 m, respectively, given the common evidence for boiling within epithermal ore zones. Boiling is the process that most favors precipitation of bisulfide-complexed metals such as gold. This process and the concomitant rapid cooling also result in many related features, such as gangue-mineral deposition of quartz with a colloform texture, adularia and bladed calcite in low-sulfidation deposits, and the formation of steam-heated waters that create advanced argillic alteration blankets in both low-sulfidation and high-sulfidation deposits.

Epithermal deposits are extremely variable in form, and much of this variability is caused by strong permeability differences in the near-surface environment, resulting from lithologic, structural, and hydrothermal controls. Low-sulfidation deposits typically vary from vein through stockwork to disseminated forms. Gold ore in low-sulfidation deposits is commonly associated with quartz and adularia, plus calcite or sericite, as the major gangue minerals. The alteration halos to the zone of ore, particularly in vein deposits, include a variety of temperature-sensitive clay minerals that can help to indicate locations of paleofluid flow. The areal extent of such clay alteration may be two orders of magnitude larger than the actual ore deposit. In contrast, a silicic core of leached, residual silica is the principal host of high-sulfidation ore. Outward from this commonly vuggy quartz core is a typically upward-flaring advanced argillic zone consisting of hypogene quartz-alunite and kaolin minerals, in places with pyrophyllite, diaspore, or zunyite. The deposit form varies from disseminations or replacements to veins, stockworks, and hydrothermal breccia.

During initial assessment of a prospect, the first goal is to determine if it is epithermal, and if so, its style, low-sulfidation or high-sulfidation. Other essential determinations are: (1) the origin of advanced argillic

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alteration, (i.e., hypogene, steam-heated, or supergene), (2) the origin of silicic alteration (e.g., residual silica or silicification), and (3) the likely controls on grade (i.e., the potential form of the orebody), because this is one of the most basic characteristics of any deposit. These determinations will define in part the questions to be asked, such as the relationship between alteration zoning and the potential ore zone, and will guide further exploration and eventual drilling, if warranted. Observations in the field must focus on the geologic setting and structural controls, alteration mineralogy and textures, geochemical anomalies, etc. Erosion and weathering must also be considered, the latter masking ore in places but potentially improving the ore quality through oxidation. As information is compiled, reconstruction of the topography and, hence, hydraulic gradient during hydrothermal activity, combined with identification of the zones of paleofluid flow, will help to identify ore targets. Geophysical data, when interpreted carefully in the appropriate geological and geochemical context, may provide valuable information to aid drilling by identifying, for example, resistive and/or chargeable areas.

The potential for a variety of related deposits in epithermal districts has exploration implications. For example, there is clear evidence for a spatial, and in some cases genetic relationship between high-sulfidation epithermal deposits and underlying or adjacent porphyry deposits. Similarly, there is increasing recognition of the potential for economic intermediate sulfidation-state base metal \pm Au-Ag veins adjacent to high-sulfidation deposits. By contrast, end-member low-sulfidation deposits appear to form in a geologic environment incompatible with porphyry or high-sulfidation deposits of any economic significance. The explanation for these empirical metallogenic relationships may be found in the characteristics of the magma (e.g., oxidation potential) and of the magmatic fluid genetically associated with the epithermal deposit.

For effective exploration it is essential to maximize the time in the field of well-trained and experienced geologists using tried and tested methods. Understanding the characteristics of the deposit style being sought facilitates the construction of multiple working hypotheses for a given prospect, which leads to efficiently testing each model generated for the prospect, using the tools appropriate for the situation. Geologists who understand ore-forming processes and are creative thinkers, and who spend much of their time working in the field within a supportive corporate structure, will be best prepared to find the epithermal deposits that remain hidden.

Introduction

MINERAL EXPLORATION, or more appropriately, mineral discovery, starts with exploration geologists walking over rocks and making observations. The successful explorationist uses knowledge of geologic relationships and ore-deposit styles, tempered by experience, to interpret all information available from a given prospect; these interpretations lead to action. Insight into how the characteristics of an ore style may be affected by various geologic settings can be bolstered by an understanding of how the processes of ore formation are manifested as hydrothermal products such as patterns of alteration and ore minerals. In the case of exploration for epithermal gold deposits, this understanding can be augmented by familiarity with active hydrothermal systems, their present-day analogues.

Although finding an attractive prospect in the first place is hard work, a majority of exploration budgets these days is spent on assessment. Therefore, we focus on the exploration of epithermal deposits from the perspective of an explorationist assessing a prospect or submission. The job of the explorationist includes building the information base that is necessary to develop multiple working hypotheses for a prospect, and establishing valid tests of each hypothesis, whether this is a mental test or a drill hole.

During assessment of a prospect, it is essential to know the geologic and economic objectives of a company, and to keep these in mind during each step. Regardless of how good a prospect looks, if the geologic controls indicate a tonnage that is insufficient to meet the company's economic requirements, or if the mineralized zone is likely to be either low grade or refractory, the answer should be to stop

further expenditure. Although no one wishes to walk away from an orebody waiting to be found, careful consideration of economic factors at each step of the assessment of a project is essential to prevent excessive unwarranted expenditure. Although the adage that mines are made, not discovered, is commonly true, no amount of investment can make a mine from an orebody that is not there.

We begin our discussion with some comments on exploration philosophy and its connection with genetic and exploration models. Following this, we define the epithermal environment and related terminology, discuss the hydrothermal processes that are important in the epithermal environment, and review the characteristics of epithermal deposits, including their tops, bottoms, and sides. In this context, the first goal during the assessment of a prospect is to determine if it is epithermal, and if so, its style, as this will determine in part the questions to be asked. The field geologist can then focus on the geological setting and the various features of the hydrothermal system. These features provide the framework to identify the location of the paleofluid flow channels, and to determine whether or not there is ore potential. There are a variety of tools that can assist these efforts, and these are reviewed below. We finish with a consideration of the causes of regional variations among giant epithermal deposits.

Exploration Philosophy

An exploration philosophy is more a set of beliefs than an organized system of thought, although it requires some of the latter to be useful. It resides usually with those responsible for leading a company's exploration efforts, and at its best, is understood and supported by the corporate leaders. These beliefs, coupled with the corporate needs and

resources, largely determine the most fundamental choices made in constructing an exploration plan, strategic or otherwise, and the organization that will implement it.

The elements of this philosophy can range from experience-driven abstract concepts such as odds-of-discovery, in general or as is known for a given deposit type, to the confidence (or belief) in specific scientific theories (genetic models). It also includes pragmatic business notions such as relative finding costs versus time to discovery, or discovery costs per ounce versus stage of exploration.

Most exploration geologists are arguably driven more by the desire to succeed, to discover an economic deposit, than by a wish to prove a particular theory. Consequently, the tendency is to rely on exploration philosophies that are a mix of pragmatic and scientific beliefs. An example of an end-member philosophy that has nonetheless formed the basis of large, long-lived corporate exploration efforts is that of using the best metallogenic science and the best geologists, and being consistent and persistent in the most favorable metallogenic provinces. There is a strongly contrasting philosophy, also often applied, based on the argument that the discovery of a significant orebody is statistically highly improbable by any single company. This leads to the policy that the more effective approach is to examine the best prospects developed by others and acquire them at an optimum moment in their exploration process.

In evaluating prospects, epithermal gold occurrences in particular—given their relative abundance and similarity, either of the end-member philosophies outlined above, or any combination of the two, require a solid basis of diagnostic characteristics. These are derived from genetic and exploration (descriptive) models, and must be applicable to prospects ranging from unexplored to those in the advanced, grid-drilling stage. Accurate evaluation, however, when based on a complex and extensive model, requires time and resources.

It has taken almost a century to develop a systematic, albeit still imperfect, understanding of epithermal ore deposits. To expect such an understanding to be translatable during a brief exploration review in the field, intended to accurately define whether the prospect is epithermal, and its style, size potential, and grade, is unrealistic. Furthermore, the acquisition, compilation, and analysis of sufficient data to make informed decisions on whether or not to pursue a certain prospect, assuming there are few exploration data available to start with, is not possible for all prospects. Thus, it becomes necessary to reduce the number of prospects to the few that can be addressed systematically. To accomplish such a reduction, some basic philosophical framework must first exist, which uses economic, social, logistical, and in-company resource parameters to select prospects irrespective of their genetic affiliation.

A fundamental first step is recognition and application of company assets. If these include a strong geological, field-oriented exploration group, even though funding may be comparatively low, the search for virgin or early-stage prospects has a sound basis. A good understanding at the corporate level of a company's exploration capabilities, and

of the exploration process, is essential to continued support and eventual success. For this reason, buying good exploration capabilities in the absence of an in-house geological tradition rarely lasts and seldom succeeds.

Just as geological skills and exploration experience are the defining elements of a philosophy of exploration, the needs of the company determine, as much as the funding and skills available, which level of exploration it pursues and where: grassroots, early stage, or advanced targets, and targets near company-owned ore bodies or in far-flung locations. These needs include consideration of existing reserves and the production goals of a company, and the timing for additions to or replacement of reserves. Given the impracticality of examining every prospect, the need to reduce the number of prospects will benefit from an analysis of the economic and logistic characteristics of different types of deposits of the particular metal being sought. This assumes that the corporate time horizon allows the search to begin with virgin districts or superficially explored prospects.

Epithermal gold deposits have size, geometry, ore-type, and grade variations that can be broadly organized around some genetic classes and, therefore, influence the exploration approach or philosophy. For example, the economic-related characteristics of some disseminated high-sulfidation deposits (e.g., Yanacocha, Perú and Pascua-Lama, Chile-Argentina) and low-sulfidation deposits (Round Mountain, Nevada) suit large, relatively low capital-cost open-pit operations. Conversely, structurally controlled low-sulfidation vein-type deposits (e.g., Hishikari, Japan and Midas, Nevada) are more likely to be high-grade underground operations, with contrasting capital requirements, development schedules, and environmental impact. In addition, the medium-grade high-sulfidation deposits (e.g., Pueblo Viejo, Dominican Republic, and Chelopech, Bulgaria), unless oxidized, tend not to be economically viable because of their complex metallurgy and high arsenic content. Keeping in mind such varied economic considerations and their effect on exploration philosophy, we next examine the geologic features of these epithermal gold deposits.

Nature of the Epithermal Environment

Why is there an epithermal environment of ore deposition?

A rigorous definition of the much-used term epithermal is difficult (Henley, 1991). Lindgren (1922) first defined the epithermal environment as being shallow in depth, typically hosting deposits of Au, Ag, and base metals (Table 1). Epithermal systems have also been exploited for a wide range of metals and minerals, including Hg, Sb, S, kaolinite, alunite, and silica. Lindgren's estimate for maximum depth of formation was 3,000 feet (about 1,000 m), based on geologic reconstruction. The estimate of the upper pressure limit was 100 atmospheres, which corresponds to that of low-salinity, low-gas water that is boiling at ~1,000-m depth. Lindgren deduced a temperature range of 50° to 200°C from the perceived stability limits of various minerals, and similarities of vein textures with those of hot spring deposits, the latter formed at less than 100°C.

TABLE 1. Alternative Nomenclature Used for the Two Main Epithermal Environments and Correspondence to Understanding of Geothermal Activity

Geothermal (dominated by neutral pH and reduced hypogene fluid)	Volcanic-hydrothermal (dominated by early acidic and oxidized hypogene fluid)	Reference
	Goldfield type	Ransome, 1909; Bethke, 1984
Au-qtz veins in andesite and rhyolite Ag-Au, Ag, Au-Te, and Au-Se veins Base metal veins with Au, Ag Cinnabar, stibnite veins	Au-alunite	Lindgren, 1922, 1933
	Enargite Au	Ashley, 1982
Hot spring		Giles and Nelson, 1982
Adularia-sericite	Acid sulfate	Heald et al., 1987
Low sulfur	High sulfur	Bonham, 1986
Low sulfidation	High sulfidation	Hedenquist, 1987
Adularia-sericite	Alunite-kaolinite	Berger and Henley, 1989
	Intermediate sulfidation	This study
	Barren quartz-alunite lithocap	Sillitoe, 1995a; this study

We now know that deposits with textures and mineral assemblages characteristic of the epithermal environment have minerals and fluid inclusions that record a maximum temperature of about 300°C, although most deposits form in a temperature range of about 160° to 270°C. The maximum temperature at a given depth under hydrostatic pressure is limited by the vapor pressure of boiling water. As there is abundant evidence that boiling is common within epithermal ore zones, this temperature interval corresponds to a depth range below the paleowater table of about 50 to 700 m, respectively. Few deposits with epithermal characteristics have formed below 1,000 m depth (Hedenquist et al., 1996; Sillitoe, 1999).

Lindgren (1933, p. 452) concluded that ore deposition occurs because focused, rapidly ascending fluids quickly change composition within a kilometer or so of the surface. We now know that this change is caused by boiling, the process that most favors precipitation of bisulfide-complexed metals such as gold. Boiling and the concomitant rapid cooling also result in many related features, such as gangue-mineral deposition of quartz with a colloform texture, adularia and bladed calcite, and the formation of steam-heated waters that create advanced argillic and argillic (Table 2) alteration blankets and halos. In addition, sharp depressurization follows hydraulic fracturing, and this also focuses the flow of vigorously boiling fluid. For these reasons there is an epithermal environment of ore deposition.

End-member styles of hydrothermal system

Two contrasting styles of hydrothermal systems exist within the epithermal environment, and both are well known from the study of active examples (e.g., Henley and Ellis, 1983). The two epithermal deposit styles of contrasting alteration and ore assemblages form within these distinctly different systems in somewhat contrasting volcanic settings (Fig. 1).

At one extreme are geothermal systems with a near-neutral

pH and reduced deep fluid that is essentially in equilibrium with the altered host rocks owing to its relatively slow ascent, resulting in a rock-dominated system (Giggenbach, 1992a). The liquid typically is low salinity, less than 1 to 2 wt percent NaCl equivalent, and may be gas-rich, with CO₂ and H₂S the dominant gases. Where this liquid discharges at the surface, boiling, neutral-pH springs deposit silica sinter. Steam-heated waters also occur in this environment, formed by condensation of vapor into ground water. Vapor that condenses within the vadose zone, above the water table, forms blankets of sulfate-rich waters. Where the vapor condenses on the margins of the system below the water table, the waters are CO₂-rich. Surface features associated with the steam-heated zone include steaming ground, mud volcanoes, and collapse craters in clay-altered ground.

Geothermal systems typically occur some distance from a volcanic edifice, although they can also occur in areas without contemporaneous volcanic activity or volcanic rocks. In most cases the geothermal systems are driven by intrusions located as much as 5 to 6 km below the surface. Although systems with relatively saline waters also occur, the high density of the deep liquid prevents discharge at the surface, and these systems are not well known. An exception is the well-studied but atypical Salton Sea system, with the brine being amagmatic in origin (McKibben and Hardie, 1997).

At the opposite extreme, volcanic-hydrothermal systems occur in a location proximal to volcanic vents that focus the discharge of magmatic vapors to the surface (Fig. 1). Their principal surface expressions are high-temperature fumaroles and related condensates of extremely acid water. The acidic, oxidized fluid is far from equilibrium with the host rocks, reflecting its magmatic affiliation (Giggenbach, 1992b). The strong structural control causes the rapid fluid ascent that is responsible for its reactive nature and fluid-dominated character (Giggenbach, 1992a). The parent intrusions may be very shallow, even erupting to the surface.

TABLE 2. Alteration Assemblages Relevant to the Epithermal Environment with Comments on Characteristic Lithologic or Mineralogical Features

Assemblage	Low sulfidation	High sulfidation
Silicic (see Table 4)	Quartz veins and veinlets, silicified breccia and/or stockwork; shallow silicification, including chalcedony and/or opaline blankets; silica sinter	Residual quartz bodies, often field termed "uggy silica"; partial to massive silicification; quartz veins and silicified breccias; shallow silicification, including chalcedony and/or opaline blankets; no sinter
Advanced Argillic	Kaolinite-alunite-(illite/smectite-native sulfur) \pm opaline blankets of steam-heated origin; commonly underlain by chalcedony blankets Kaolinite/halloysite-alunite-jarosite blankets or zones of supergene origin	Alunite-kaolinite/dickite-pyrophyllite-diaspore of hypogene origin, typically surrounding silicic cores; also sericite-pyrophyllite roots Kaolinite-alunite blankets of steam-heated or supergene origin
Argillic or Intermediate Argillic	Illite/smectite halo to veins; illite \pm smectite halo to deeper sericitic zones	Illite and illite/smectite halo to advanced argillic core
Propylitic	Broad host to ore system, in some cases deuteric in origin and of questionable direct genetic relation to epithermal ore forming system; typically chloritic (not epidote), except at deeper levels	
Sericitic/Phyllic	Sericite-dominated rock, typically as a halo to deep quartz veins	Sericitic roots to advanced argillic areas; may be transitional to advanced argillic

Note: These alteration assemblages may accompany ore, or may be barren; in the case of hypogene advanced argillic assemblages (quartz-alunite halos to residual quartz), the term lithocap is used when there is thought to be a spatial relationship to an underlying porphyry system (Sillitoe, 1995a)

¹Sericite is a field term used to designate fine-grained white mica, and may constitute illite, 2M muscovite, and intermediate K-dioctahedral micas (Meyer and Hemley, 1967)

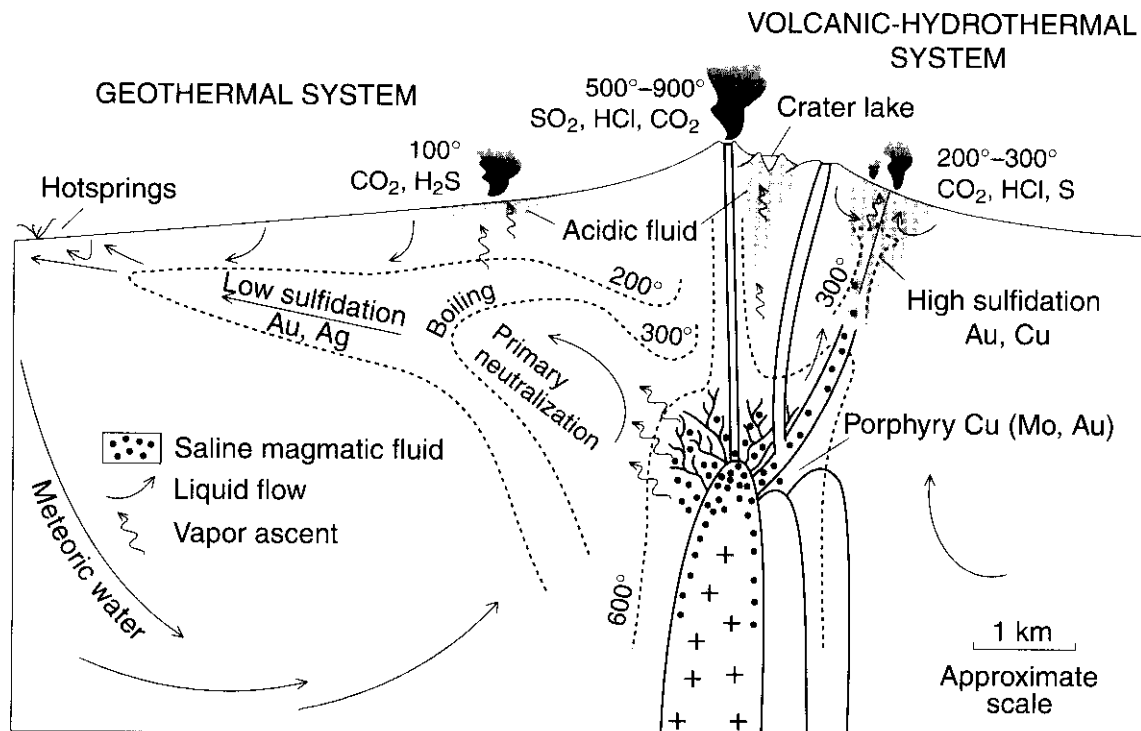


FIG. 1. Cartoon to illustrate schematically the various processes deduced for volcanic-hydrothermal and geothermal systems, and the respective environments of high-sulfidation and low-sulfidation styles of epithermal ore deposits relative to the intrusive engine. We do not infer necessarily this spatial relationship between all systems (from Hedenquist and Lowenstern, 1994, integrated from many sources, including Sillitoe, 1975; Giggenbach, 1981; Hemley and Ellis, 1983).

These volcanic-hydrothermal systems have distinctly different characteristics from their geothermal counterparts, although both can coexist nearly side by side. In some cases, there is a transition downward from geothermal to volcanic-hydrothermal environments at depths of only 1 to 2 km, where hypogene acid fluid rises along fractures or shallow dikes into an overlying neutral-pH geothermal system (Reyes et al., 1993). Typically, this transition is represented by a zone of hydrolysis (Meyer and Hemley, 1967), also called primary neutralization (Giggenbach, 1981), located beneath the epithermal environment (Fig. 1).

Terminology: High- and low-sulfidation styles of epithermal deposit

The two distinctly different chemical environments have also long been recognized for epithermal deposits (Ransome, 1907), although the interpretation of the cause of this difference has varied (Ransome, 1909; Lindgren, 1933; Nolan, 1933; White, 1955). There are a variety of terms that have been used to describe the two end-member styles of epithermal deposits (Table 1). These various terms have caused some confusion at times, and for this reason we review the origins of the terms. The ore-deposit characteristics are strongly influenced by the contrasting compositions of the deep fluid, outlined above, either neutral and reduced or acidic and oxidized. Therefore, the various terms that have been used in the past usually reflected either the different processes or the resulting products of these two end-member types of fluid.

We use the terms low sulfidation and high sulfidation to reflect the two end-member sulfidation states, deduced from the sulfide mineral assemblages (Barton and Skinner, 1979). These terms were first suggested on the basis of the oxidation state of the sulfur in the fluid (Hedenquist, 1987). However, as this is impractical to determine for an ore prospect, the terms are now used to refer to the sulfidation state of the sulfide assemblage. This mineralogical feature reflects the intrinsic nature of the ore fluid (Heald et al., 1987; Hedenquist, 1987; John, 1999), both its origin and also the degree of fluid-rock interaction, rock-dominated for low-sulfidation and fluid-dominated for high-sulfidation systems (Giggenbach, 1992a).

However, just as there is a transition between the end-member active systems, so too do we recognize deposits with features intermediate between the low-sulfidation and high-sulfidation end-members (Sillitoe, 1993a; John et al., 1999; this study). The sulfide assemblage of end-member low-sulfidation epithermal deposits is typically pyrite-pyrrhotite-arsenopyrite and Fe-rich sphalerite. The high-sulfidation deposits, by contrast, are characterized by enargite-luzonite-covellite plus pyrite assemblages. In the past, the low-sulfidation term has been applied to deposits with a sulfide assemblage intermediate between the low-sulfidation and high-sulfidation end-members (i.e., tennantite-tetrahedrite-chalcopyrite and Fe-poor sphalerite). In this study we subdivide the low-sulfidation category into end-member low-sulfidation deposits and those with assemblages indicating an intermediate sulfidation state. This is not simply a distinction of the sulfide assemblage, nor is it intended to confuse

the terminology. Rather, it is an attempt to recognize and distinguish the possibility that intermediate sulfidation-state and end-member low-sulfidation deposits form in different tectonic settings, and have different magmatic affiliations (John, 1999; John et al., 1999).

The hypogene acid fluid that is generated in the volcanic-hydrothermal environment (Fig. 1) leaches the rock, creating a core of residual, commonly vuggy silica that recrystallizes to quartz. These silicic zones invariably form the core of an advanced argillic alteration halo (Fig. 2), and the silicic core may serve as an aquifer for a subsequent ore fluid (White, 1991). The ore fluid, when present, is distinct in composition from the early-stage fluid responsible for leaching; it is less acidic and less oxidized, and is also relatively saline (Hedenquist et al., 1998). Sulfide minerals deposited during this later stage include enargite and pyrite (Table 3).

A shortcoming of the sulfidation-state terminology is the fact that it is based on ore minerals, whereas barren prospects—those that experienced only the early stage of leaching—do not contain diagnostic high-sulfidation minerals such as enargite. Barren prospects are common in the volcanic-hydrothermal environment, similar to the silicic and quartz-alunite lithocaps (Sillitoe, 1995a) that form over degassing intrusions (Hedenquist et al., 1998). In these situations, the high-sulfidation ore fluid either did not form at depth or did not ascend to epithermal depths, thus, leaving an alteration assemblage typical of a lithocap, the term we use for a barren system of this alteration style (Table 1).

The low-salinity liquid responsible for formation of low-sulfidation ore veins and disseminations (Fig. 3) is similar to waters tapped by drilling beneath geothermal hot springs. The low-sulfidation state minerals that form from these reduced, neutral-pH waters are in equilibrium with the host-rock alteration minerals. The salinity of the fluids that form intermediate sulfidation-state deposits is somewhat higher than that of the end-member low-sulfidation systems (Table 3), and the sulfide assemblage indicates a sulfidation state that has not fully been equilibrated with the rock buffer, in contrast to end-member low-sulfidation deposits.

The two end-member styles of deposit, low-sulfidation and high-sulfidation, are also clearly distinguished on the basis of their hypogene alteration mineralogy (Table 2). Quartz-adularia-carbonate veins with sericitic or clay halos commonly host low-sulfidation ore (Fig. 3), in contrast to the leached silicic host of high-sulfidation ores with quartz \pm alunite \pm pyrophyllite \pm dickite halos (Fig. 2). The silicate alteration mineralogy of intermediate sulfidation-state deposits is broadly similar to that of end-member low-sulfidation deposits (Table 3), indicating that they also form from a near-neutral pH ore fluid. Distinctions include an abundance of rhodochrosite and anhydrite versus chalcocite and adularia in intermediate sulfidation-state versus end-member low-sulfidation deposits, respectively.

In reality, the alternative nomenclature that developed over the century simply reflects different ways to refer to the same set of observations. Early distinctions were commonly based on a type example (e.g., Goldfield type; Ransome, 1909), and this neatly avoided any genetic discussion, including changes

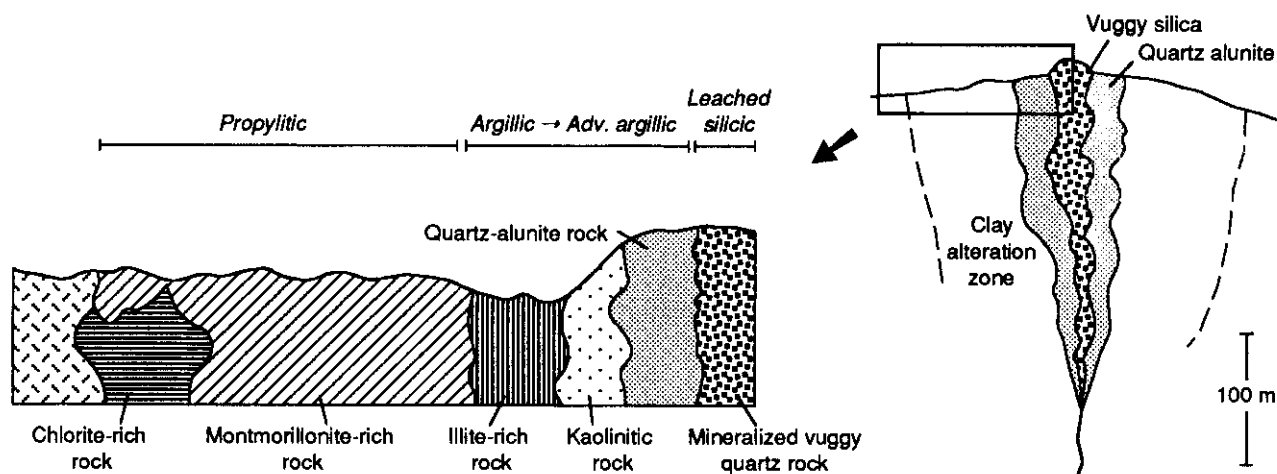


FIG. 2. Section through a typical high-sulfidation orebody, showing the upward-flaring zone of the silicic core (Stoffregen, 1987), with an inset (Steven and Ratté, 1960) illustrating the characteristic alteration zonation outward from the silicic core that may have a texture of vuggy quartz. The silicic core is the principal host to high-sulfidation ore, although portions of the advanced argillic zone can also contain ore, particularly where pyrophyllite dominates over silicic zones (White, 1991). Note that patches of advanced argillic assemblages (e.g., quartz-alunite) may be contained within the silicic core, most likely caused by permeability variations that result in some zones being incompletely leached.

in the perceived origin with time (White and Hedenquist, 1990). Lindgren (1933) distinguished epithermal types on the basis of metal assemblage (Table 1), noting the distinct Au-alunite association from the work by Ransome and others. Nolan (1933) argued for a division that was based on Au-rich or Ag-rich ore, a precursor to the low-sulfidation and intermediate sulfidation-state grouping to distinguish these observed differences. The Russian literature has long recognized the distinctive high-sulfidation style, and used the term secondary quartzite (Fedorov, 1903; Nakovnik, 1933, 1968) to refer to the lithocap of residual quartz associated with volcanism and massive sulfide mineralization, including porphyry-type deposits (Rusakov, 1926; Nakovnik, 1934).

Increased exploration for gold in the late 1970s and 1980s led to a renewed examination of epithermal deposits. The classic low-sulfidation-style veins of the western Americas (Buchanan, 1981), similar to some of the first epithermal deposits ever studied, located in central Europe (Lindgren, 1933), were termed the adularia-sericite (Heald et al., 1987) in recognition of the common presence of vein adularia and halos of sericite (Buchanan, 1981). However, in some deposits adularia is rare or absent, particularly in southwest Pacific examples (White et al., 1995) and others with a high base metal content associated with andesitic volcanism (Sillitoe, 1993a). The counterpart to this term, acid sulfate (Heald et al., 1987), refers to the deduced nature of the altering fluid responsible for generating hypogene advanced argillic assemblages. A problem arises here because an acidic sulfate-rich solution can also be generated in the vadose zone over the top of both styles of system, from syn-hydrothermal steam-heated waters and from post-hydrothermal weathering of sulfides (see below). Thus, advanced argillic minerals (Table 2) can form in three distinct environments (e.g., Rye et al., 1992).

Although mineralogic and textural distinction is possible (Sillitoe, 1993a), a casual reference to an acidic fluid or an advanced argillic assemblage without specifying the mineralogy and determining its origin can lead to some geologists assuming, incorrectly, a high-sulfidation style of system.

The terms low sulfur and high sulfur (Table 1; Bonham, 1986) refer to the total amount of sulfide minerals in a deposit. In contrast, the terms low and high sulfidation do not refer to low and high concentrations of sulfide minerals, but rather reflect the oxidation potential and sulfur fugacity of the fluid that deposited the sulfides (Barton and Skinner, 1979). Although low-sulfidation and high-sulfidation deposits typically have low and high sulfide contents, respectively, there are pyrite-rich low-sulfidation and base metal-rich intermediate sulfidation-state deposits with sulfide contents as high as sulfide-poor high-sulfidation examples (Sillitoe, 1993a). Lastly, hot-spring type is a term that was introduced following the discovery of the McLaughlin deposit, California, beneath silica sinter (Giles and Nelson, 1982). However, this and similar examples are simply a shallow-formed subset of low-sulfidation deposits, typically end-member low-sulfidation deposits, with evidence of paleosurface preserved, and do not constitute a distinct style.

Why do we need to define the various environments of epithermal deposit formation, and understand these variations during exploration? As noted here, the two end-member styles of a system create different alteration and mineralization products with potential for markedly different ore controls and geometries (Sillitoe, 1993a; White and Hedenquist, 1995). For this reason, the correct framework of interpretation is essential, starting from the earliest stage of exploration and assessment. Guidelines to distinguish the style of epithermal environment can all be employed in the field, as discussed in the following sections.

TABLE 3. High- and Low-Sulfidation Deposit Characteristics

	Low-sulfidation deposits ¹		High-sulfidation deposits		
	Shallow	Deep	Shallow ²	Intermediate ²	Deep (porphyry) ²
Genetically related volcanic rocks	Andesite-rhyodacite (AR), bimodal rhyolite-basalt (RB), alkalic (A)		Andesite-rhyodacite, dominated by calc-alkalic magmas		
Depth of formation	0–300 m	300–800 m (rarely >1,000 m)	<500	500–1,000 m	>1,000 m
Setting, typical host rock	Domes; pyroclastic and sedimentary rocks	Domes, diatremes (AR, A); pyroclastic and sedimentary rocks	Domes, central vent; pyroclastic and sedimentary rocks	Domes, diatremes; volcanic rocks	Dome-diatreme; porphyry, volcanic, clastic sedimentary rocks
Deposit form	Vein, vein swarm, stockwork, disseminated	Vein, breccia body, disseminated	Disseminated, breccia and veinlet	Massive sulfide veins, breccia, ledges	Dissemination, veinlets, breccia
Ore textures	Fine bands, combs, crustiform, breccia	Coarse bands	Vuggy quartz hosts replacement	Massive sulfide, late veins/breccias	Replacement
Alteration	Alunite-kaolinite blanket, clay halo	Clays, sericite, carbonates; roscoelite, fluorite (A)	Silicic (vuggy), quartz-alunite	Silicic (vuggy), quartz-alunite, pyrophyllite-dickite-sericite	Pyrophyllite-sericite, quartz-sericite
Gangue ³	Chalcedony-adularia-illite-calcite	Quartz-carbonate-rhodonite-sericite-adularia ± barite ± anhydrite ± hematite ± chlorite (AR)	Alunite, barite, kaolinite	Anhydrite, kaolinite, dickite	Sericite, pyrophyllite
Sulfides ³	Cinnabar, stibnite; pyrite/marcasite-arsenopyrite, Au-Ag selenides, Se sulfosalts, pyrrhotite, Fe-rich sphalerite (RB)	Pyrite-Au-Ag sulfides/sulfosalts, variable sphalerite, galena, chalcopryrite, tetrahedrite/tennantite (AR)	Enargite/luzonite, covellite, pyrite	Enargite/luzonite, chalcopryrite, tetrahedrite/tennantite, sphalerite, late covellite, pyrite	Bornite, digenite, chalcocite, covellite
Metals ³	Au-Ag-As-Sb-Se-Hg-Tl (RB), low Ag:Au; <0.1–1% base metals	Ag-Au-Pb-Zn, Ba, Mn, Se (AR), high Ag:Au; 2–10 (20+) % base metals	Au-Ag, Cu leached (Hg overprint)	Cu-Au-Ag-Bi-Tc-Sn	Cu-Au
Notable features	Sinter, chalcedony blanket	Some intermediate sulfidation-state veins adjacent to high-sulfidation ore	Steam-heated blanket	Vuggy quartz host	Overprinted on porphyry features
Fluids	<1% NaCl, gas-rich, <220°C (RB)	3–10+% NaCl, 220–280°C+ (AR)	<2 wt% NaCl	4–15+ wt% NaCl	Variable
Examples	McLaughlin, Midas, Round Mountain, Sleeper, Hishikari (low-sulfidation)	Comstock, Tonopah, Creede, Fresnillo, Casapalca, Victoria (intermediate sulfidation-state)	Yanacocha, Pueblo Viejo, Pierina, La Coipa, Tambo, Pascua, Paradise Peak, Summitville, Rodalquilar, Kasuga	El Indio, Lepanto, Chinkuashih, Goldfield, Lahóca	Bisbee, MM, Chuquicamata

Based on Lindgren, 1933; Buchanan, 1981; Heald et al., 1987; Sillitoe, 1993a, 1999; White et al., 1995; John et al., 1999; this study

¹Includes both intermediate-sulfidation and end-member low-sulfidation deposits

²Sillitoe, 1999

³Intermediate sulfidation-state deposits associated with andesite-rhyodacite (AR) tend to form at greater depths than the end-member low-sulfidation deposits associated with rhyolite-basalts (John et al., 1999); a deep to shallow transition from the AR to RB association, or from intermediate sulfidation-state to low-sulfidation character, is not implied

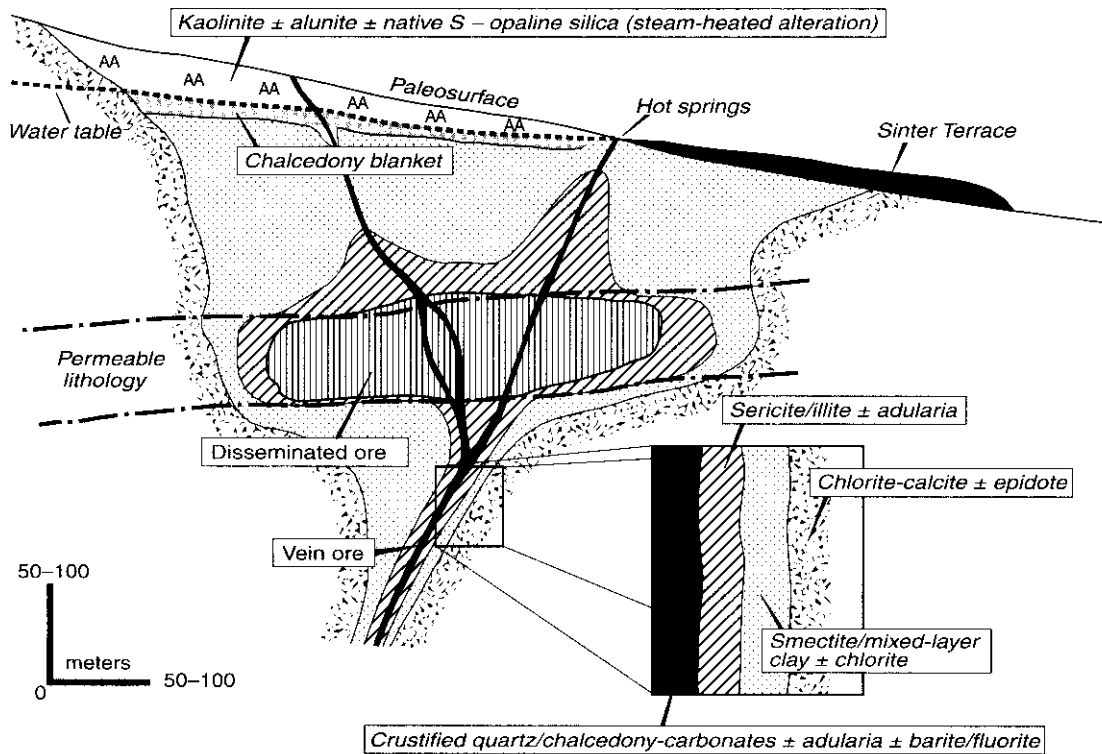


FIG. 3. Schematic section that generalizes patterns of alteration in a low-sulfidation system, showing the variable form with increasing depth, and the typical alteration zonation, including the distribution of sinter, a blanket of advanced argillic (AA) steam-heated alteration, and water table silicification (Buchanan, 1981; Sillitoe, 1993a). The geologic variation between deposits accounts for many deviations from this generalization.

Genetic Framework

Epithermal fluids and processes

Drilling into numerous geothermal systems around the world has provided abundant information on temperature-depth relationships (Henley et al., 1984). The principal upflow in most systems has a thermal gradient indicating boiling conditions, in some cases adjusted for the gas-rich nature of the fluid (Fig. 4a). High gas contents contribute to the vapor pressure, thus pushing a given isotherm to greater depths. For example, high concentrations of CO_2 in solution can cause a 300°C liquid to start to boil as deep as 1,500 m depth, instead of 1,000 m for pure water (Henley et al., 1984). In contrast, the resistance to flow in a fracture typically results in hydrodynamic pressures 10 percent greater than hydrostatic (Hedenquist and Henley, 1985a). In some cases mineral deposition may result in the flow becoming restricted, causing the pressure to increase to as much as twice that of the hydrostatic profile (e.g., at Sulphur Bank and McLaughlin, California; Moore et al., 2000). This will result in the isotherms being compressed closer to the surface. The higher pressure will also contribute to the likelihood of hydraulic fracturing, hydrothermal brecciation, and possible hydrothermal eruption, such as has occurred in the geothermal systems of Yellowstone, Wyoming, and Waitapu

and Waimangu, New Zealand (Hedenquist and Henley, 1985b). Conversely, high salinity decreases the depth to a given boiling temperature, but this effect is limited at the low salinities that are characteristic of Au-rich epithermal fluids (Hedenquist and Henley, 1985a).

Boiling and mixing are the two principal processes that occur in geothermal systems (Giggenbach and Stewart, 1982), together with vapor condensation near the surface. In high-flux hydrothermal systems, fluid ascent is sufficiently rapid that the liquid will intersect its boiling-point-for-depth curve. Because a high fluid—and metal—flux is a requirement for the formation of a hydrothermal ore deposit (Henley, 1985), it follows that boiling should be common in the upflow zones of an ore deposit. Deep fluid typically mixes with shallow water on the margins of the system, either cool ground water or its steam-heated equivalent, and this invariably quenches boiling (Fig. 4a).

In geothermal systems and their low-sulfidation analogues, the principal control on fluid pH is the concentration of CO_2 in solution, together with the salinity (Henley et al., 1984). Thus, boiling and loss of CO_2 to the vapor results in an increase in the pH (eq. 1). In turn, this causes a shift from illite to adularia stability (eq. 2). The loss of CO_2 also leads to the deposition of calcite (eq. 3). This explains the common occurrence of adularia and bladed calcite as gangue minerals in low-sulfidation ore veins.

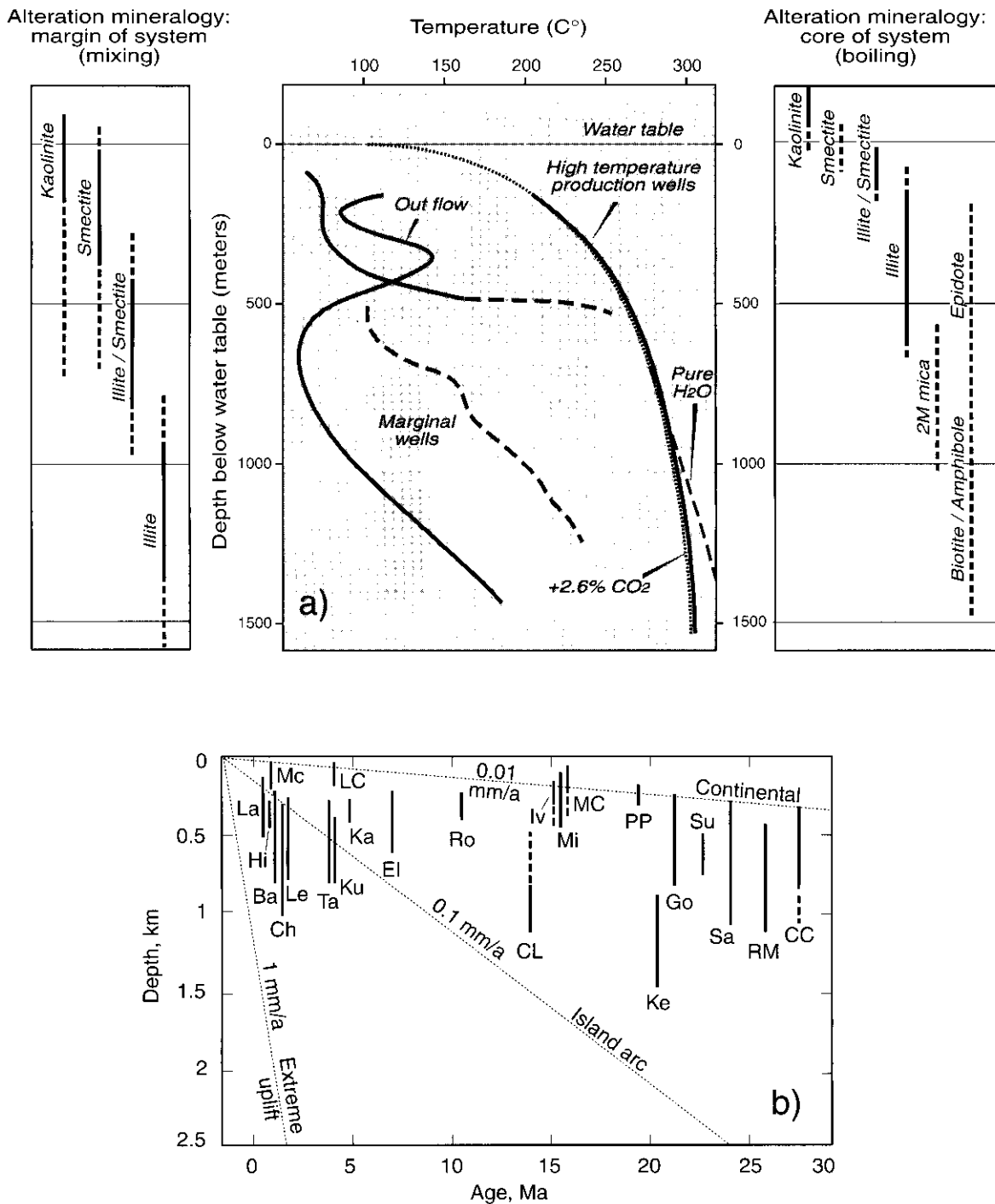
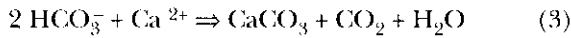
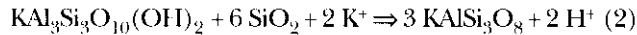
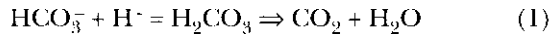
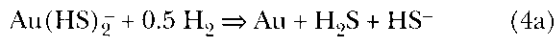


FIG. 4. a. Boiling point for depth curve, for pure water and a gas-rich solution at close to hydrostatic pressure. The schematic distribution of alteration minerals in the wall rock proximal to the boiling upflow portion of the system is shown to the right; many of these minerals form as the result of boiling, and the distribution of some reflect their temperature dependence (Fig. 8). The pattern of temperature-dependent minerals that form on the cooler margins of the system where mixing predominates is shown schematically to the left. b. Depth ranges for epithermal ore versus age (modified from Hedenquist et al., 1996), showing the typical erosion rates for continental and island arcs settings, and the rate for extreme uplift (e.g., Papua New Guinea; Sillitoe, 1994). See Fig. 9 for low-sulfidation or high-sulfidation character. Ba = Baguio; CC = Cripple Creek; Ch = Chinkuashih; CL = Comstock Lode; El = El Indio; Go = Goldfield; Hi = Hishikari; Iv = Ivanhoe; Ka = Kasuga; Ke = Kelian; Ku = Kushikino; La = Ladolam; LC = Lewis-Crofoot (Sulphur); Le = Lepanto; Mc = McLaughlin; Mi = Midas; MC = Mule Canyon; PP = Paradise Peak; Ro = Rodalquilar; RM = Round Mountain; Sa = Sado; Su = Summitville; Ta = Taio; mm/a = millimeter per year erosion rate.

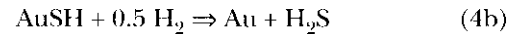


Boiling in the focused upflow of a hydrothermal system (Simmons and Christenson, 1994) is a critical process in the epithermal environment, because boiling and the associated gas loss are the principal causes of gold precipitation from bisulfide complexes. Gold saturation occurs due to the loss of the sulfide ligand to the vapor (eq. 4a; Buchanan, 1981; Brown, 1986; Cooke and Simmons, 2000). This is particularly relevant to the deposition of high-grade gold ore in low-sulfidation veins. Thus, evidence for boiling to have occurred in an epithermal prospect is desirable.



The evidence that boiling occurred in epithermal deposits includes alteration blankets of steam-heated origin (Buchanan, 1981), adularia and bladed calcite in low-sulfidation veins (Simmons and Christenson, 1994), fluid inclusion relations (Roedder, 1984) and, indirectly, hydrothermal breccias that indicate hydraulic fracturing and pressure release (Hedenquist and Henley, 1985b). Boiling also causes cooling and concentration of dissolved species such as silica, leading to quartz supersaturation and the formation of silica colloids. The colloids deposit as a gel that recrystallizes to fine, colloform-banded chalcedony, a feature typically associated with gold dendrites in high-grade low-sulfidation veins. Dendrites grow from gold colloids, another indication of supersaturation caused by boiling (Saunders, 1994). In addition, truscottite (a hydrated Ca-Al silicate) is a mineral that is found most commonly in association with high-grade gold ore. It is stable only when the silica concentration exceeds quartz saturation, another indirect line of evidence for boiling (Izawa and Yamashita, 1995). Although steam-heated blankets and hydrothermal breccias are common in high-sulfidation deposits (Sillitoe, 1999) as well as low-sulfidation deposits, the other features are restricted to low-sulfidation systems.

The high-sulfidation ore fluid typically deposits sericite, dickite, or kaolinite where silicate minerals are able to form adjacent to the vuggy quartz rock that typically hosts ore. These minerals indicate a lower pH than is typical for low-sulfidation fluids that deposit adularia and calcite, about pH 4 to 5 versus pH 6 to 7, respectively (Hedenquist et al., 1998). In the more acid high-sulfidation environment, the dominant gold complex may be bisulfide (Benning and Seward, 1996; Giggenbach, 1997), even though the fluid is relatively oxidized and has a moderate salinity (Hedenquist et al., 1998). By contrast, within the epithermal environment the solubility of the gold-chloride complex will be very low except under acidic and oxidized conditions (i.e., typical of the fluid that may be responsible for early leaching and production of the preore lithocap).

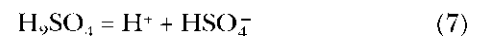


In contrast to the many indications for boiling in low-sulfidation deposits, there is evidence for mixing based on the O and H isotope data from some high-sulfidation deposits. Trends of isotopic data indicate a ground-water diluent during alunite formation (Arribas, 1995), consistent with the temperature-salinity trend of enargite-hosted fluid inclusions from the Julcani, Perú (Deen et al., 1994) and Lepanto, Philippines (Mancano and Campbell, 1995) high-sulfidation deposits. Dilution by cool ground water at Lepanto would have prevented boiling during enargite deposition, because mixing always causes a boiling fluid to fall below the boiling curve (Fig. 4a). Because gold deposition followed closely after enargite deposition in this and similar deposits, boiling may not be the sole precipitation mechanism of sulfides and gold in high-sulfidation deposits. Regardless, boiling must occur at some stage in the life of high-sulfidation systems, as indicated by the common occurrence of hydrothermal breccias, some of which predate mineralization, and evidence, prior to erosion, of steam-heated alteration blankets (Sillitoe, 1999). Our lack of knowledge about the ore fluid and mineralization process responsible for forming high-sulfidation deposits reflects the fewer studies of such deposits and their lack of silicate alteration minerals, plus our inability to drill into their highly reactive and dynamic present-day volcanic equivalents.

Origin of acid waters in the epithermal environment

During exploration it is essential to distinguish the origin of advanced argillic alteration (Bethke, 1984), because the alteration products have distinctly different spatial and genetic relationships to the potential ore zone. For this reason, we elaborate here on the origins of the acid waters that generate advanced argillic assemblages. As mentioned above, there are three principal sources of natural acidity: hypogene magmatic condensates, steam-heated oxidation, and supergene oxidation. The first is responsible for forming the advanced argillic alteration of barren lithocaps as well as high-sulfidation deposits, whereas the latter two can create blankets of advanced argillic alteration over both high-sulfidation and low-sulfidation deposits (Fig. 5; Sillitoe, 1993a; White and Hedenquist, 1995).

The proximal volcanic setting of high-sulfidation deposits accounts for the presence of hypogene acidic species, including HCl, SO₂, and HF in decreasing abundance (Hedenquist, 1995). Dissociation of the dominant acidic species, HCl and H₂SO₄ (eqs. 5 and 7, respectively) occurs at less than 300° to 350°C, subsequent to absorption of the high-temperature magmatic vapor by ground water and disproportionation of the SO₂ (eq. 6).



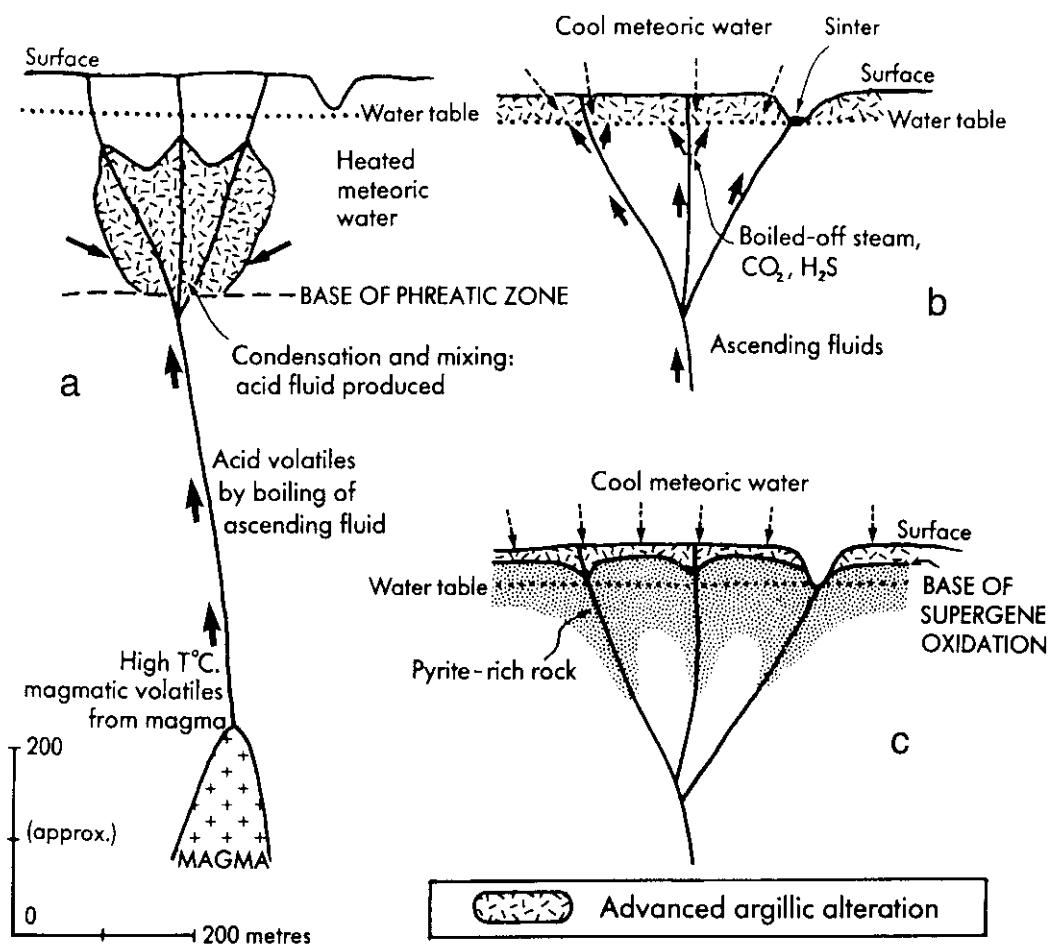
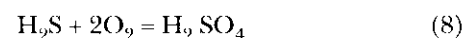


FIG. 5. Schematic illustration of the environment of formation of the three types of acid waters, hypogene (eqs. 5–7), steam-heated (eq. 8), and supergene (eq. 9) (Schoen et al., 1974; Giggenbach, 1992a; Sillitoe, 1993a). In a, acidity derives from ascending and cooling HCl and SO₂, the latter after it has been condensed into water and has formed sulfuric acid (cf. Fig. 2). In b, acidity derives from oxidation of H₂S gas that condenses within the vadose zone (cf. Fig. 3). By contrast, in c, acidity derives from post-hydrothermal oxidation of pyrite within the vadose zone. From Sillitoe (1993a)

This process results in the development of hypogene hydrochloric-sulfuric acid water with a pH of about 1, sufficiently acid to leach most components, including Al, from the rock. This leaching leaves a siliceous residue that soon recrystallizes to quartz, sometimes with a vuggy texture (vuggy quartz, or porous quartzite in the Russian literature), and also forms the advanced argillic alteration halo typical of the lithocaps that host high-sulfidation deposits (Fig. 2). The creation of acid conditions is dependent on absorption of magmatic vapors by ground water; thus, the zone of silicic and advanced argillic alteration has a sharp lower boundary (Stoffregen, 1987; White, 1991), coincident with an aquifer (Giggenbach, 1992a). Cooling results in the solution becoming increasingly reactive, causing the alteration to flare upward as reactive fluid is continually supplied at depth (Figs. 2 and 5). Where the acid fluid intersects a permeable lithology or structure, flow will occur along the most permeable channel, down the hydraulic gradient. If this results in lateral flow, the distribution of this alteration type may be asymmetric.

High-temperature, high-pressure vapors apparently transport metals such as Cu, Au, and As (Heinrich et al., 1999), possibly as molecular S and Cl complexes. Aerosols similar to those accompanying volcanic eruption (Hedenquist, 1995) may also be responsible for transporting metals into the epithermal environment during the formation of this type of advanced argillic alteration (Muntean and Einaudi, 2000). Despite vapor transport of metals at high pressure, hypogene silicic and advanced argillic alteration generated by condensation of relatively low-pressure, metal-poor vapor (Hedenquist, 1995) should not be highly anomalous in metals, as noted in barren lithocaps without subsequent introduction of sulfide minerals.

In both high-sulfidation and low-sulfidation systems, H₂S is present and will oxidize to sulfate in the presence of atmospheric O₂ within the vadose zone (Schoen et al., 1974). This process forms steam-heated acid sulfate water.



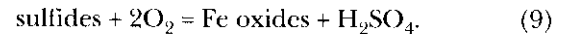
Deeply circulating cold ground water contains a maximum of 10 ppm dissolved O_2 , insufficient to create an acid sulfate water beneath the vadose zone. Thus, because steam-heated acid-sulfate water forms only within the vadose zone, its distribution mimics that of the ground-water table, forming a blanket of alteration (Figs. 3 and 5), although locally the acid water can descend along fractures if the water table is perched. Flow will occur down the hydraulic gradient towards drainages (Schoen et al., 1974; Sillitoe, 1993a). The thickness of the zone of acid-sulfate water is only several meters and, thus, temperatures seldom exceed 100° to 120°C. During syn-hydrothermal erosion, however, the zone of steam-heated alteration will fall with the water table, leading to the development of a thick blanket of acid-sulfate alteration that may overprint deeper alteration. This was the case at Sulphur, Nevada (Ebert and Rye, 1997), which owes its name to the sulfur deposits, formed by steam-heated activity, that cap the ore veins.

The pH of steam-heated acid-sulfate water is typically 2–3, owing to the lack of HCl and the generally low sulfate concentrations. This fluid readily dissolves volcanic glass and many minerals, but because Al tends to remain insoluble at pH greater than 2, it will be fixed in minerals such as kaolinite and alunite. This precludes the widespread development of residual silica (>95% SiO_2) within steam-heated blankets of alteration, although a friable and silica-rich opaline cap of residual material may form locally. Subsurface lateral flow of this water causes silicification of the aquifer as the acid solution reacts with wall rock and is neutralized,

forming blankets of chalcedony at the water table and zones of opal above the water table (Table 4).

In a related situation, condensation of vapor below the vadose zone forms a CO_2 -rich steam-heated water (eq. 1; Hedenquist, 1990) that occurs as a discontinuous umbrella across the top and draped over the margins of the upflow plume. The mildly acid (pH 4–5) water creates argillic halos of smectite and inter-layered clays plus kaolinite, siderite, and other carbonates, and extends locally to 1,000 m depth (Simmons and Browne, 2000). Incursion of this water during the late-stage collapse of a system may form carbonates, some Mn-rich, that are commonly barren (Simmons et al., 2000).

The third environment in which acidic solutions form is related to post-hydrothermal supergene oxidation of sulfide minerals:



Supergene oxidation has many of the same controls as that of steam-heated oxidation, because it occurs only within the vadose zone and is controlled by the position of the water table. Temperature is limited to a maximum of 30° to 40°C, and in addition to the formation of secondary clays, kaolinite, halloysite, and alunite, jarosite is common and Fe oxides are ubiquitous. Acid waters may drain downward locally along faults and open fractures (Fig. 5). Weathering of sulfides improves the recovery of refractory gold in low-grade epithermal deposits that otherwise would be

TABLE 4. Types of Silicic Alteration

Type	Formation	Where?	Significance	Metals	LS or HS
Sinter	From near-neutral pH hot springs	Only at surface	Paleosurface, topographic (hydrologic) depression, focus of upflow	Var. As, Sb, Hg, Tl (Au, Ag if flared vent)	LS only
Residual silica (opaline)	Moderate leaching, pH ~2–3, 80–90% SiO_2	In vadose zone	Steam-heated origin, above paleowater table	Hg, unless overprint	LS or HS
Chalcedony horizon	Silica remobilized from steam-heated zone; deep fluid may contribute to outflow	At water table, up to 1–2+ km from source	Paleowater table, may be distal from source	Hg if only steam-heated, As, Sb, Au, Ag if deep fluid	LS or HS
Chalcedony veins, colloform bands; cryptocrystalline veins	From low-T fluid, colloids; recrystallized from gel	Shallow depth, <150 m	<200°C, rapidly cooling fluid, boiling at depth; cryptocrystalline at ~200°C	As, Sb, Se, Au, Ag	LS or late HS
Quartz veins, vugs	From cooling solution	>150 m depth	>200°C	Au, Ag, base metals	LS, late HS
Residual silica (vuggy quartz)	Extreme leaching at pH <2, >95% SiO_2	Core of volcanic-hydrothermal system	Permeable core, principal host to high-sulfidation ore	Barren, or Cu, As, Au, Ag	HS only
Silicification	From cooling water	Surface to 500 m, massive <150 m depth	Shallow portion of system, pervasive flow	Trace Au, Ag	LS, mid to late HS

Abbreviations: LS = low sulfidation, HS = high sulfidation

uneconomic, and also leads to the removal of Cu, which is a metal that increases cyanide consumption during heap-leach recovery of Au. The tectonic and climatic setting of the western Americas is more favorable than the southwest Pacific for such economically advantageous supergene oxidation (Sillitoe, 1999).

Setting and General Characteristics of Epithermal Ore Deposits

Low-sulfidation deposits, and their tops, bottoms, and sides

Buchanan (1981) and White et al. (1995) reviewed the host rocks and other characteristics of 60 and 137 epithermal deposits in the southwestern United States and southwestern Pacific, respectively (Fig. 6). The majority of the deposits reviewed, particularly those in southwestern United States, were low-sulfidation in style, both end-member low-sulfidation as well as intermediate sulfidation state. Although Buchanan (1981) focused on the physical characteristics of the deposits, White et al. (1995) also examined the controls on ore, and how they varied in different settings. White et al. (1995) noted some differences between the two regions, with the southwest Pacific deposits having characteristics that suggest a greater depth of formation, a larger degree of lateral flow—as a result of higher relief in stratovolcanic settings—and a more rapid rate of uplift. All these factors probably reflect the island-arc settings of most of these deposits, relative to the continental setting of the southwestern United States. Like their geothermal analogues, low-sulfidation deposits typically are distant from contemporaneous central vents. However, they commonly occur within a dome setting, such as Cerro Crucitas, Costa Rica (Pease, 1999), and Castle Mountain, California (Capps and Moore, 1991), and most end-member low-sulfidation deposits in northern Nevada are associated with rhyolite domes (John et al., 1999). Low-sulfidation deposits are affiliated with a wide range of rock types, from alkalic to calc-alkalic (Sillitoe, 1993a). End-member low-sulfidation deposits may be restricted to bimodal basalt-rhyolite settings, in contrast to the andesite-rhyodacite setting noted for intermediate sulfidation-state deposits in Nevada (Table 3; John et al., 1999). A specific style of low-sulfidation deposit is associated with alkalic rocks (Table 3; Richards, 1995).

Examples of end-member low-sulfidation epithermal deposits in the western United States include gold-rich veins at Midas and Sleeper, and disseminated deposits such as Round Mountain, all in Nevada (Buchanan, 1981; John et al., 1999). Martha Hill and Golden Cross in New Zealand (White et al., 1995; de Ronde and Blattner, 1988), some of the mines in the Baguio district, Philippines, and Hishikari, Kushikino, Seigoshi, and Sado mines in Japan (Izawa et al., 1990) are also examples of this style of epithermal deposit (Fig. 6). Other large deposits in the western Pacific region include Gunung Pongkor and Kelian in Indonesia, and Ladolam in Papua New Guinea. The Ag ± base metal-rich deposits of Comstock Lode, Nevada and Creede, Colorado, as well as Pachuca and Fresnillo in Mexico (Simmons, 1991), possess sulfide assemblages that indicate an inter-

mediate sulfidation state, as noted above. The higher fluid salinity in these intermediate sulfidation-state deposits (Table 3) accounts for their high Ag and base metal concentrations (Henley, 1990; Simmons, 1995). Other regions with major end-member low-sulfidation plus intermediate sulfidation-state deposits and potential for discovery include central America (e.g., Cerro Crucitas, Costa Rica), the Andes (e.g., Portovelo, Ecuador, and several prospects in Perú), eastern Australia (Vera Nancy), Patagonia (Cerro Vanguardia, Argentina), the western Mediterranean (Almagrera and Mazarrón, Spain), the Carpatho-Balkan arc (e.g., Rosia Montana and Baia Mare in Romania, and Madjarovo in Bulgaria), Central Asia (Arharlik in Kazakhstan and Baleyskoye in Russia) and the Magadan (Kubaka and Dukat) and Kamchatka regions (Asacha and Aginskoye) of Far East Russia (Fig. 6).

Gold ore in low-sulfidation deposits is commonly associated with quartz and adularia plus calcite or sericite as the major gangue minerals (Sillitoe, 1977; Buchanan, 1981; Berger and Eimon, 1983; White et al., 1995; Tables 3 and 5). The form of the deposit can vary from vein (Sleeper, Midas and Hishikari in Japan) to stockwork (McLaughlin, Cerro Crucitas) to disseminated (Round Mountain) (White and Hedenquist, 1990; Sillitoe, 1993a; Fig. 7). The alteration halos to the zone of mineralization, particularly in vein-controlled mineralization, include a variety of temperature-sensitive clay minerals (Fig. 8). The areal extent of such clay alteration may be two orders of magnitude larger than the actual ore deposit. This is usually the case with the shallow, lower-temperature alteration (e.g., Hishikari; Izawa et al., 1990), that mushrooms near the surface owing to the intersection of an aquifer by basement feeder structures, the latter potentially being host to high-grade ore. Thus, even after a large alteration system is found, it may still be difficult to assess where the ore is located.

Wall-rock alteration assemblages include illite, chlorite, albite, epidote, zeolites, and pyrite, in addition to quartz, adularia, and calcite (Tables 3 and 5). These minerals reflect the near neutral-pH and reduced composition of the ore fluid. Interstratified illite-smectite and smectite clays plus kaolinite occur on the margins of the system, as well as within the ore zone, in some cases as supergene alteration products of hydrothermal sericite. End-member low-sulfidation deposits contain very minor base metal (Zn-Pb) sulfides, in contrast to intermediate sulfidation-state deposits.

Tops of low-sulfidation ore deposits: The most distinctive paleosurface feature of low-sulfidation systems is sinter, which forms finely laminated terraces of amorphous silica around neutral pH hot springs. Aprons may extend in the direction of the drainage for several hundreds of meters. Finely laminated air-fall or lacustrine sediments that have been silicified, in many cases by an outflow of steam-heated water, may be mistaken for sinter. The presence of plant fragments, common in sinter, is not diagnostic, as such material also accumulates with a variety of finely laminated sediments. The only diagnostic criterion is that of the vertical structures that form due to algal growth as well as evaporation (White et al., 1989). Care must be taken in distinguishing true sinter

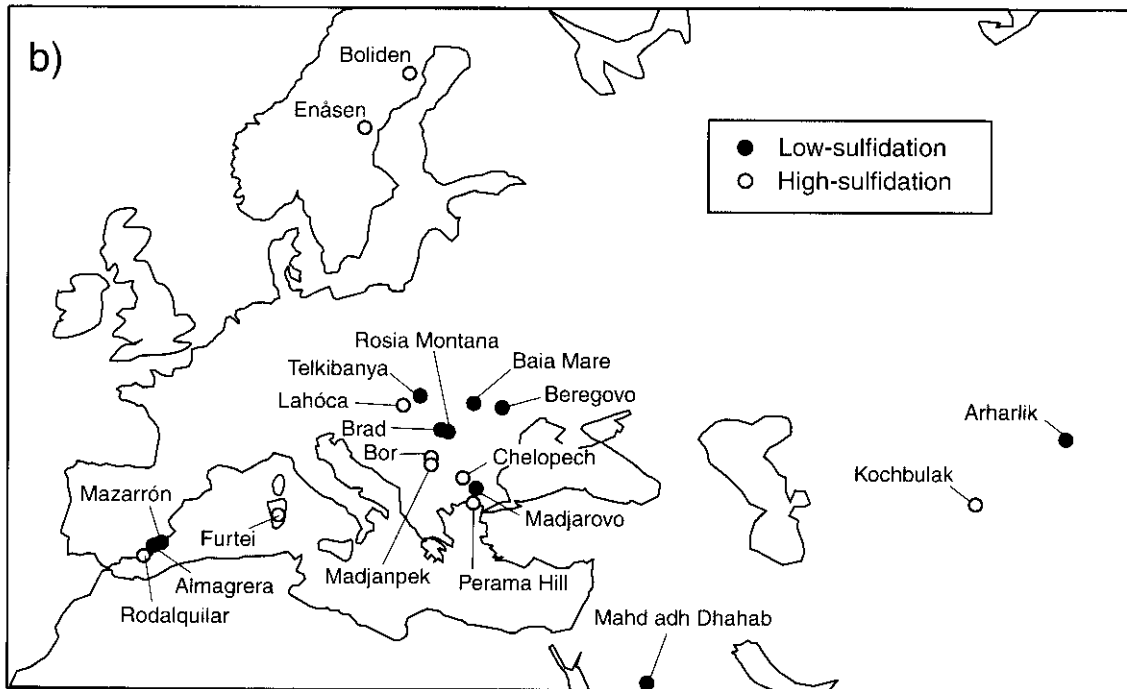
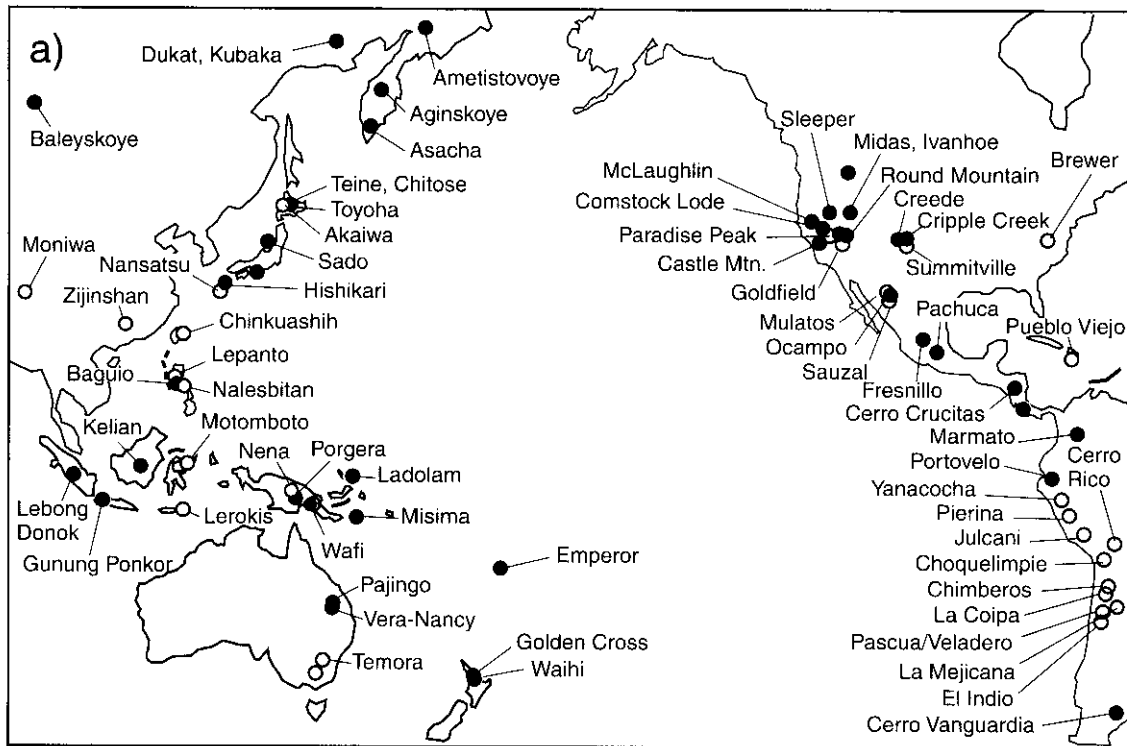


FIG. 6. Distribution of principal low-sulfidation and high-sulfidation epithermal deposits in the circum-Pacific region (a) and Europe-Central Asia (b).

TABLE 5a. Depth and Temperature Controls on Hypogene Alteration Characteristics of Low-Sulfidation Deposits

Approx. depth (m)	Paleosurface or paleowater table	0–150 m	150–300 m	300–500 m	500–1,000 (1,500) m
Temp. (°C)	100–120	100–200	200–230	230–260	260–300+
Advanced argillic alteration	Steam-heated (kaolinite-alunite) blanket over water table; residual opal	Steam-heated (kaolinite-alunite) overprint along fractures	Rare overprint	None	None
Silicic alteration	Sinter around hot springs, horizon of chalcedony at water table	Colloform chalcedony veins, wall-rock silicification	Qtz veins, fine bands, open-space filling	Qtz veins w/ coarse bands	Minor qtz veins, minimal silicification
Argillic alteration	Kaolinite-smectite	Smectite, mixed illite/smectite; marginal halo to 1,000 m depth	Minor illite/smectite to illite	Illite	Illite, illite/smectite halo on margins
Sericitic alteration	None	None	None	Illite ± chlorite	2M mica
T-sensitive or T indicative	Opal	Chalcedony	Loss of interlayer clays	Illite, epidote presence	Biotite-amphibole, >800–1,000 m
Associated minerals	Kaolinite, alunite	(Adularia)	Adularia	Adularia	K feldspar
Carbonates, others	None	Bladed calcite, late Mn carbonates	Bladed calcite, rhodochrosite + barite (AR)	Calcite	Calcite
Sulfides (RB)	Pyrite-marcasite at base	Pyrite	Pyrite, arsenopyrite, pyrrhotite, Ag selenides	Pyrite, arsenopyrite, pyrrhotite (Ag selenides)	Minor base metal sulfides, sphalerite-galenachalcocopyrite
Sulfides (AR)	Pyrite-marcasite at base	Pyrite	Tetrahedrite-tennantite	Major base metal sulfides	Base metal sulfides
Propylitic		Chloritic	Chloritic, trace epidote	Chlorite ± epidote	Epidote

TABLE 5b. Depth and Temperature Controls on Hypogene Alteration Characteristics of High-Sulfidation Deposits

Depth (m)	Paleosurface-water table	0–150 m	150–300 m	300–500 m	500–1,000 (1,500) m
Temp. (°C)	100–120	100–200	200–230	230–260	260–300+
Advanced argillic alteration	Steam-heated kaolinite-alunite blanket; residual opal, cristobalite	Quartz-alunite halo to silicic core	Wide quartz-alunite halo, depending on host	Flaring upward, quartz-alunite; pyrophyllite, diaspore	Narrow zones, pyrophyllite, sericite
Silicic alteration	Chalcedony horizon at water table	Chalcedony veins, residual silica, vuggy quartz, silicification	Residual silica, vuggy quartz, silicification	Residual silica, vuggy quartz, silicification	Quartz D-type porphyry veins
Argillic alteration	Kaolinite-smectite	Smectite-interlayered illite/smectite	Transitional to sericitic	On margins	On margins
Sericitic alteration	None	None	None	Illite-sericite, to sericite-pyrophyllite	Sericite-pyrophyllite to 2M mica
T-sensitive or indicative	Opal	Chalcedony	Loss of interlayered clays	Pyrophyllite	Biotite
Sulfides	Pyrite-marcasite at base	Pyrite-marcasite at base, transitional to Cu-As-Sb sulfides/sulfosalts	Enargite-luzonite with later Au+tennantite-tetrahedrite-chalcocopyrite	Enargite-luzonite with later Au+tennantite-tetrahedrite-chalcocopyrite	Enargite, bornite, digenite, chalcocite, covellite

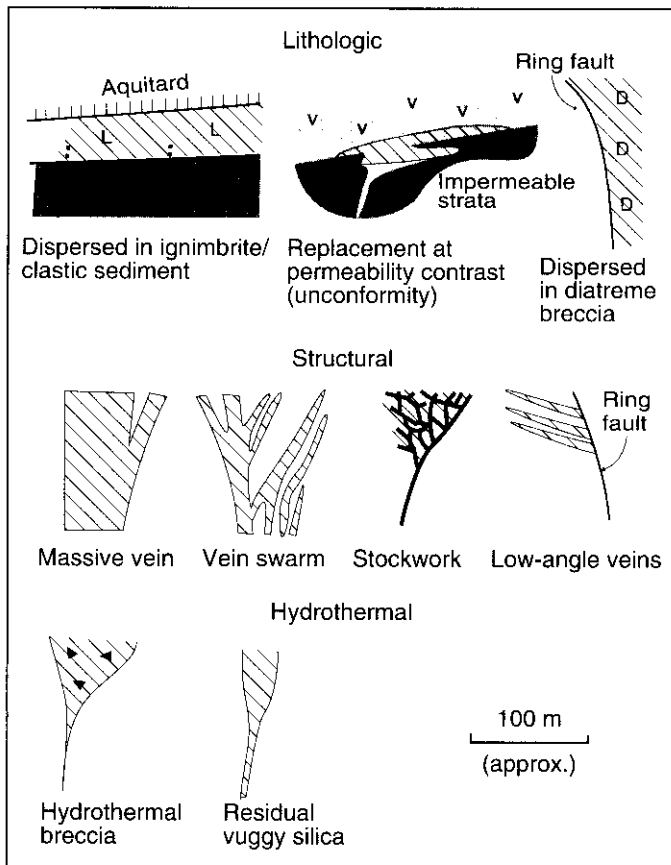


FIG. 7. Examples of orebody form, controlled by the host-rock lithology, by structure, and by hydrothermal processes (from Sillitoe, 1993a). Ore shown by angled lines.

from a variety of finely laminated, dense siliceous beds that form in the epithermal environment. The presence of true sinter proves that the system is low sulfidation, fixes the position of the paleosurface, and, most importantly, identifies the location of a principal upflow channel of boiling fluid (Fig. 3).

Steam-heated water forms blankets of kaolinite, cristobalite, smectite, and, locally, alunite and native sulfur (Figs. 3 and 5). There is no direct relationship to ore, but such blankets typically overlie the ore system in the hanging wall, for example, at McLaughlin (Sherlock et al., 1995) and Kushikino, Japan (Matsuhisa et al., 1985). This alteration assemblage may overprint ore in cases where the groundwater table collapses (e.g., Sulphur and Mt. Muro, Indonesia; Ebert and Rye, 1997; Simmons and Browne, 1990). Where these acid fluids descend along fractures and are heated (Fig. 5), such as in high-relief Philippine geothermal systems (Reyes, 1990), they may form narrow zones of higher-temperature minerals such as pyrophyllite and diaspore. Zoning of alteration minerals in the steam-heated zone, including silica minerals, alunite, kaolinite, and smectite, may be present (Schoen et al., 1974), although this surficial alteration blanket is rarely preserved except in the youngest systems.

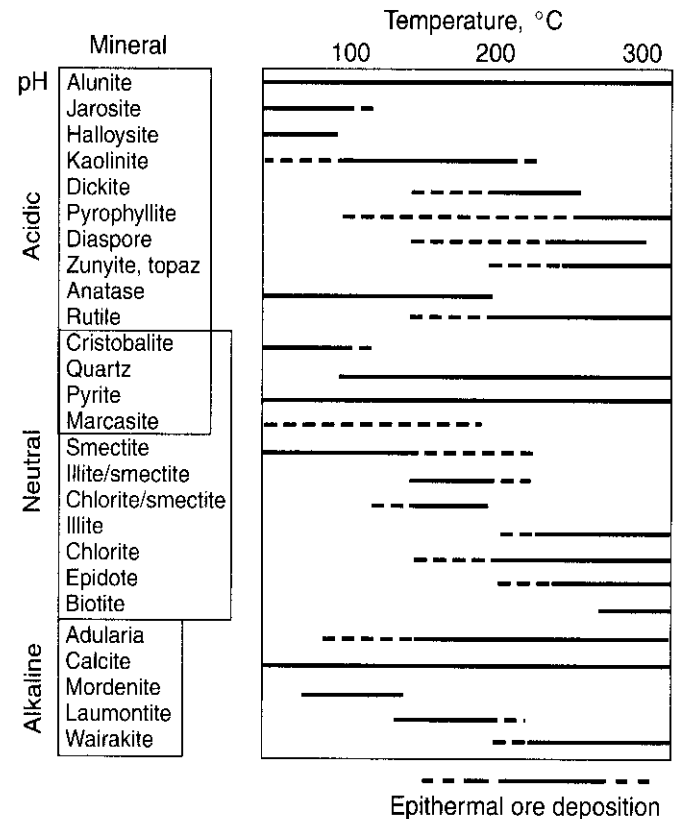


FIG. 8. Thermal stability of various hydrothermal minerals that occur in the epithermal environment under acid and neutral-pH conditions, and the typical temperature range for deposition of epithermal ore (cf. Fig. 4; Reyes, 1990; Hedenquist et al., 1996).

Steam-heated water collects at the water table and creates aquifer-controlled stratiform blankets of dense silicification, typically chalcedony (Schoen et al., 1974), up to several kilometers distant from the zone of generation of steam-heated water (Sillitoe, 1993a). Although the horizon of silicification may be no more than a few meters thick, a falling—or rising—water table resulting from erosion or in-filling of the drainage point can form a siliceous blanket up to 25 m or more in thickness. These acid-sulfate solutions are also Fe-rich and, thus, may form pyrite and marcasite near the base of the horizon, in proximal positions, where H_2S is available. This acid water may pond in depressions, such as eruption craters, leading to the formation of laminated siliceous deposits interbedded with lacustrine sediments.

The low-pressure, $\sim 100^\circ C$ vapor that is involved in generating steam-heated acid sulfate waters does not transport NaCl or metals, except for Hg, which is particularly volatile. Indeed, Hg has been mined from such occurrences over epithermal ores, for example, in the Sulphur and Ivanhoe districts (low-sulfidation) and at the Paradise Peak (high-sulfidation) deposit, all in Nevada. Thus, metal anomalies should not be generated by steam-heated alteration in either low-sulfidation or high-sulfidation systems (Sillitoe, 1993a), except where preexisting metal deposition is over-

printed by a falling water table (Ebert and Rye, 1997). In addition, silicification formed at the ground-water table will not create metal anomalies, aside from Hg, except in cases where an ascending metal-rich liquid also discharges along the same ground-water outflow.

Sides of low-sulfidation deposits: Low-sulfidation deposits grade outward, in some cases sharply, to argillic halos whose widths relate to the primary permeability of the host rocks—narrow halos around structurally focused ore, or wide areas in permeable rocks. The argillic assemblage is transitional outward to propylitic assemblages that may be district-wide in extent.

The argillic alteration forms as the result of the generation of CO₂-rich steam-heated water on the margin of the system (Hedenquist, 1990). The mildly acidic (pH 4–5) water creates halos of illite, interstratified clays, and smectite as well as kaolinite and siderite that extend locally to 1,000 m depth (Fig. 8; Simmons and Browne, 2000). The outward and upward decrease in temperature from the fluid channelway is recorded by temperature-dependent clay minerals (Figs. 7 and 8) that form a zoned halo around ore zones. For example, at Hishikari, clays are increasingly interstratified with distance from the veins (Izawa et al., 1990). Incursion of this marginal water during the late-stage collapse of a system may account for the formation of calcite veins and Mn-rich carbonates that are commonly barren (Simmons et al., 2000).

Bottoms of low-sulfidation deposits: The vertical interval of low-sulfidation ore zones typically averages about 300 m (Buchanan, 1981), but may be as large as 600 to 800 m for intermediate sulfidation-state deposits (Fig. 4b), or in the case of high-grade, end-member low-sulfidation deposits, may be as little as 100 to 150 m. Quartz veins may simply pinch out with increasing depth or change to narrow carbonate stringers, or they may lose gold grade, resulting in sharp bottoms of high-grade ore (e.g., at Sleeper and Hishikari). Buchanan (1981) notes that the precious metal zones in many deposits have roots rich in base metal sulfides. This is also documented in the low-sulfidation deposits of the Metaliferi Mountains of Romania, such as Brád, where a transition over several 100 meters towards higher base metal concentrations eventually leads to porphyry-style base metal mineralization. A genetic relation between porphyry and end-member low-sulfidation epithermal mineralization has not been demonstrated, and in northern Nevada, it appears that they are mutually exclusive (John et al., 1999). By contrast, intermediate sulfidation-state deposits occur in districts that also host deep porphyry deposits (John et al., 1999), and there may be a relationship. This is one of the reasons to distinguish between end-member low-sulfidation and the intermediate sulfidation-state deposits.

High-sulfidation deposits, and their tops, bottoms, and sides

Arribas (1995) and Sillitoe (1999) summarize the setting of 43 high-sulfidation deposits. These deposits range in size from the greater than 35 Moz gold deposit at Yanacocha, to several small 0.3 Moz deposits (e.g., Kasuga, Japan and

Rodalquilar, Spain; Figs. 6 and 9). The most common setting is that associated with domes (22 of 43), primarily in a dome complex, but also summit domes in a central-vent volcano. A central-vent volcano setting is the next most common (12), whereas a spatial association with diatremes or calderas accounts for only three deposits each; in some cases, the association overlaps two of these settings. For ten deposits there is insufficient information to determine their geologic setting. Rodalquilar, one of the smallest deposits considered here, is related spatially to a caldera, whereas the large Goldfield, Nevada, deposit is associated with domes that are influenced by a caldera, as at Rodalquilar (Arribas et al., 1995a).

The high-sulfidation style of epithermal deposit forms in a position transitional between the surface and a shallow degassing intrusion, in places associated with porphyry deposits. Ore bodies commonly are located proximal to volcanic vents and are hosted by structural conduits or permeable lithologies. As mentioned above, a critical requirement is the existence of an effective plumbing system in the volcanic-hydrothermal system that will allow fast ascent of magmatic fluid to epithermal depths. The compositional range of rocks genetically related to high-sulfidation deposits is relatively narrow, primarily intermediate calc-alkaline compositions, in contrast to the wider range of rock types associated with low-sulfidation deposits, both intermediate sulfidation-state and end-member low-sulfidation (Sillitoe, 1993b; Arribas, 1995; John et al., 1999).

High-sulfidation deposits are most well preserved in Cenozoic volcanic arcs of the Circum-Pacific, in the Mediterranean region, including the Cretaceous to Tertiary Carpatho-Balkan belt (e.g., Chelopech), and in the Paleozoic belt of Central Asia (e.g., Kochbulak, Uzbekistan). High-sulfidation deposits occur in two principal settings, in island arcs and at continental margins (Arribas, 1995; White et al., 1995). Older deposits are Mesozoic to Paleozoic in age (e.g., in China, Uzbekistan, and Australia), and highly metamorphosed examples (Dubé et al., 1998) have even been recognized in Proterozoic shields of Canada (Hope Brook; Dubé et al., 1998) and the Baltic region (Enåsen, Sweden; Hallberg, 1994). The large proportion of young high-sulfidation (as well as low-sulfidation) deposits simply reflects the greater likelihood of the epithermal environment being eroded with time.

The principal host rocks to high-sulfidation deposits are variable. Andesitic and dacitic volcanic rocks, both flows and breccias, host a part or all of 16 deposits out of the 43 reviewed by Sillitoe (1999); poorly to moderately welded ignimbrite hosts 10 deposits. Ignimbrite seems to be a more common host in South America, whereas other volcanic units are more common in Asia and Europe. Dacitic to andesitic porphyry and other intrusions host ten deposits, and sedimentary rocks of various types (mudstone, sandstone, calcareous units, metamorphosed graywacke) are host to parts or all of nine deposits. Despite the common association of high-sulfidation deposits with domes, surprisingly the domes themselves host a portion of only a small number of deposits. This may reflect their syn-hydrothermal timing of intrusion.

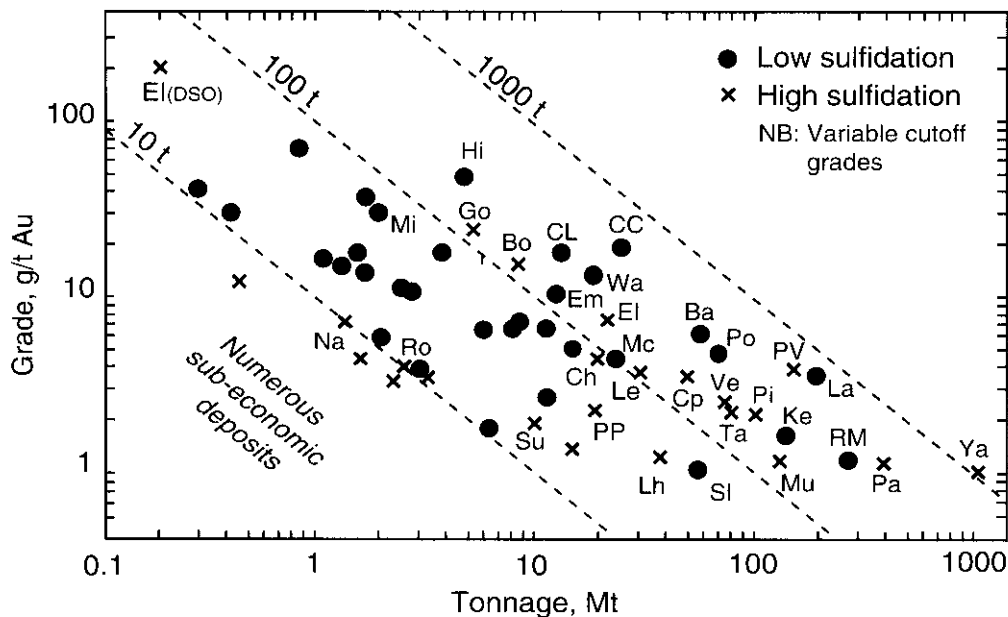


FIG. 9. Grade-tonnage plot of low-sulfidation and high-sulfidation epithermal ore deposits (modified from Hedenquist et al., 1996). Includes both production and reserves, but with various cut-off grades. Note that there are many prospects that contain low tonnages and/or grades, plotting in the lower left corner, which are largely uneconomic to mine. Intermediate sulfidation-state deposits such as Comstock Lode and Kelian are shown with the low-sulfidation symbol. Ba = Baguio, Philippines; Bo = Boliden, Sweden; CC = Cripple Creek, Colorado; Ch = Chinkuashih, Taiwan; CL = Comstock Lode, Nevada; Cp = Chelopech, Bulgaria; El = El Indio, Chile; El D.S.O. = El Indio direct shipping ore; Em = Emperor, Fiji; Go = Goldfield, Nevada; Hi = Hishikari, Japan; Ke = Kelian, Indonesia; La = Ladolam, Papua New Guinea; Lh = Lahóca, Hungary; Le = Lepanto, Philippines; Mc = McLaughlin, California; Mi = Midas, Nevada; Mu = Mulatos, Mexico; Na = Nansatsu district deposits, including Kasuga, Japan; Pa = Pascua, Chile; Pi = Pierina, Perú; Po = Porgera, Papua New Guinea; PP = Paradise Peak, Nevada; PV = Pueblo Viejo (oxide-sulfide), Dominican Republic; Ro = Rodalquilar, Spain; RM = Round Mountain, Nevada; SI = Sleeper, Nevada (average ore); Su = Summitville, Colorado; Ta = Tambo, Chile; Ve = Veladero, Argentina; Wa = Waihi, Martha Hill, New Zealand; Ya = Yanacocha, Perú.

The form of high-sulfidation deposits varies from disseminations or replacements to veins, stockworks, and hydrothermal breccia bodies (Fig. 7). Similar to low-sulfidation deposits, lithologic and structural controls determine the individual deposit form. Sillitoe (1999) discussed the characteristics of high-sulfidation deposits at porphyry, deep epithermal, and shallow epithermal levels (Table 3). In particular, he recognized a diversity in styles of high-sulfidation ore that is controlled largely by the changing nature of the permeability from the surface to greater than 1 km depth. The largest, although lowest-grade, deposits form at shallow depths, where the system mushrooms into permeable lithologies such as volcanoclastic rocks, lacustrine sediments and, in particular, ignimbrite. The pyroclastic host rocks have a range in the degree of welding; where welded they are brittle and fracture easily. By contrast, high-grade vein deposits, typically with massive accumulations of pyrite and sulfosalts minerals, tend to form at greater depths and are, thus, exposed in more deeply eroded terrain. As noted above, high-sulfidation ore deposits commonly show a large degree of structural control, even within the massive zones of vuggy quartz and sulfides, reflecting the fracture-related roots of these systems. These fractures reflect regional-scale features in some cases, whereas in other cases the fractures appear to be caused by

emplacement of the shallow intrusions to which high-sulfidation deposits are related.

One of the most common characteristics of high-sulfidation deposits is the alteration zoning outward from the orebody. Figure 2 shows the alteration section for the Summitville deposit, Colorado (Steven and Ratté, 1960). Ore is hosted by a vuggy rock consisting of quartz recrystallized from residual silica, with grades decreasing at the edge of the silicic core. Outward from the vuggy quartz is a zone of advanced argillic alteration consisting of quartz-alunite and the kandite minerals, including kaolinite, nacrite, or dickite, in places with pyrophyllite or diasporite. This assemblage may also occur in patches within the silicic zone, possibly preserved from complete leaching by local zones of lower permeability. The silicic and advanced argillic altered core is followed in an outward direction by argillic alteration of illite or interstratified clays; an outermost zone of propylitic alteration includes chlorite. The total thickness of the zone of advanced argillic alteration can be as narrow as 1 m but may be as wide as 100 m. This pattern of alteration zonation indicates progressively less acidic conditions outward from the pathway of acid fluid flow (Hemley et al., 1969, 1980; White, 1991). The silicic (\pm vuggy quartz) and quartz-alunite zones, where they can be observed, pinch downward, with a sericitic or argillic zone surrounding the structural feed

zone (Fig. 2; Stoffregen, 1987). In some deposits, the silicic zone is absent, and gold ore is hosted by quartz-pyrophyllite, such as at Pueblo Viejo (White, 1991; White et al., 1995).

Gold mineralization in high-sulfidation ore deposits is associated most commonly with enargite or its lower-temperature dimorph, luzonite. Such high sulfidation-state copper sulfides typically form early in the paragenesis, with relatively low contents of gold, and are cut by gold ore (e.g., Lepanto as well as El Indio, Chile). The post-enargite gold ore is associated with pyrite, tennantite-tetrahedrite, chalcopyrite, and tellurides; these sulfides indicate a lower sulfidation state than enargite. At Summitville, there is a transition from enargite to tetrahedrite with increasing depth (Stoffregen, 1987). Alunite is commonly an early alteration and gangue mineral, whereas anhydrite and barite are relatively late.

Tops of high-sulfidation ore deposits: Many high-sulfidation deposits, particularly those formed at shallow depth, occur near domes and stratovolcano edifices (Sillitoe, 1999; Table 3). Both volcanic environments may host acid lakes, either in summit craters or as moats around the margins of domes. Evaporation increases the sulfate concentration by up to an order of magnitude, decreasing the pH to values as low as 0.0 (e.g., in the Kawah Ijen crater lake of Java; Delmelle and Bernard, 1994). Native S typically forms pools at the lake bottom, along with pyrite \pm alunite \pm kaolinite.

The ponding of acid water also leads to the formation of laminated siliceous deposits interbedded with lacustrine sediments (e.g., Yanacocha; Sillitoe, 1999). Like in low-sulfidation systems, hydraulic fracturing and hydrothermal brecciation is common, and where these breccia zones reach the paleosurface, the eruption vent commonly is filled by water. In these situations, the shallow portion of the breccia vent may contain laminated sediments consisting of breccia fragments and alteration minerals such as alunite, kaolinite, and barite, as noted at Akaiwa, Japan, and Rodalquilar (Arribas et al., 1995a). In places, the silicic alteration zone that hosts ore at depth is seen to pinch upwards, for example, at the Panshan orebody, Chinkuashih, Taiwan (White, 1991).

Where the paleosurface is nearly preserved (e.g., at La Coipa, Pascua and Tambo, Chile, plus Yanacocha; Sillitoe, 1999), a blanket of bleached, friable opaline quartz (originally cristobalite)-kaolinite-alunite alteration is common. This blanket formed from steam-heated waters by the same mechanism discussed for low-sulfidation deposits, and native sulfur is common. Distinguishing this style of acidic alteration from hypogene quartz-alunite \pm dickite-pyrophyllite must be done on textural observations and field occurrence, rather than on mineralogy alone, as the mineralogy of hypogene and steam-heated alteration overlaps. Although pyrophyllite typically forms at high temperature in porphyry deposits, it can form at less than 150°C when the concentration of silica exceeds that of quartz solubility (Hemley et al., 1980).

Again similar to low-sulfidation systems, chalcedony forms at the base of the vadose zone along the water table (Fig. 5). Thus, the features related to steam-heated alteration are the

same in high-sulfidation and low-sulfidation systems. Conversely, the acid pH of deep waters in the high-sulfidation environment precludes the precipitation of silica directly from hot springs due to kinetic effects, and for this reason, silica sinters do not form in high-sulfidation systems.

Sides of high-sulfidation deposits: The early, reactive solution typically is controlled by a fracture, but where this fluid intersects a zone of permeability, such as a tuff bed, an unconformity, a fault, or their intersection (cf. Lepanto), lateral outflow may occur in the direction of the hydraulic gradient. Such lateral flow can result in a regionally extensive but asymmetric zone of hypogene silicic alteration. Close to the upflow, leaching leaves residual silica that may be silicified by later solutions, forming a massive unit. By contrast, the distal outflow may consist of massive chalcedony with no residual silica, and this chalcedony will be difficult to distinguish from that related to outflow of steam-heated water. The extent of lateral flow is a function of the permeability and hydraulic gradient, the latter determined by the paleorelief, and may extend for several kilometers.

Epithermal veins occur close to several high-sulfidation deposits (Sillitoe, 1993a), within a few kilometers and near the margins of alteration of the volcanic-hydrothermal system. Such veins typically possess features of intermediate sulfidation-state deposits. In a few cases, these veins form economic deposits, for example, Victoria, south of Lepanto, and Chiufen, west of Chinkuashih, both within 1 km of the high-sulfidation deposits. A lack of detailed study and precise dating for most deposits precludes an unambiguous determination of the timing or relationship of these adjacent ore bodies.

Bottoms of high-sulfidation deposits: High-sulfidation deposits typically are located above or marginal to intrusions, some associated with porphyry Cu-Au deposits. Well-known porphyry systems are recognized beneath Lahóca, Hungary and Lepanto, and a porphyry-style system was recently found beneath Yanacocha. Porphyry systems also occur below or adjacent to many other similar high-sulfidation deposits (Sillitoe, 1999).

Drilling beneath some high-sulfidation orebodies (Hedenquist et al., 1994; Arribas et al., 1995a) indicates that the silicic and advanced argillic zones commonly pinch downward (Stoffregen, 1987). The roots can be narrow, unmineralized quartz-pyrite veins with narrow to broad halos of sericitic alteration (Fig. 2), in some instances with variable amounts of pyrophyllite. Although pyrophyllite is an advanced argillic mineral, it is not restricted to forming only as a halo of the silicic and quartz-alunite advanced argillic alteration. This mineral can also form by cooling of a sericite-stable fluid (Hemley and Hunt 1992), explaining the common upward transition of sericite to pyrophyllite over the tops of porphyry deposits, but below the quartz-alunite lithocaps (Hedenquist et al., 1998). Where fluid inclusion studies have been conducted, the lower limit of advanced argillic alteration corresponds to the upper limit of hypersaline, daughter-mineral bearing liquid inclusions, for example, at Rodalquilar (Arribas et al., 1995a) and Lepanto (Hedenquist et al., 1998). This suggests that the dense

hypersaline liquid is confined largely to the upper reaches of the porphyry environment unless catastrophic events cause it to be propelled to epithermal depths (Muntean and Ein-audi, 2000).

Exploration of Epithermal Prospects: Relevant Observations and Useful Tools

As discussed above, epithermal mineralization is the result of a series of hydrothermal processes that are relatively well understood. The products resulting from these processes (e.g., silica sinter, steam-heated alteration, vuggy quartz, etc.) provide the framework for unraveling epithermal prospects and assessing their potential to contain an orebody. This section focuses on the processes and resulting products that are most relevant to exploration. Throughout the section, comments are made on the most useful assessment tools.

Methodology

During the assessment of a prospect, the first goal is to determine if it is epithermal, and if so, its style, low sulfidation or high sulfidation. This determination will define in part the questions to be asked, such as the relationship between alteration zoning and the potential ore zone (Figs. 2 and 3). The ore, gangue, and alteration mineralogy, commonly combined with textural features, typically allow a straightforward distinction between the end-member low-sulfidation and high-sulfidation styles. To a lesser extent, the metal suite and form of the deposit are also distinctive (Table 3; Buchanan, 1981; Sillitoe, 1993a; White et al., 1995; Hedenquist et al., 1996).

Once the style of prospect is determined, the field geologist can focus on understanding the geologic and structural characteristics of the prospect and determine the geometry of the system and its size in terms of ore potential. At this early stage, it is critical that one or more working hypotheses are developed, and then refined or discarded after testing with observations and ground facts. As mentioned above, a clear understanding of the economic requirements of a given project (i.e., definition of the critical path to achieve exploration success) is essential at any stage.

For epithermal deposits, as with all other deposits of hydrothermal origin, one of the main challenges for exploration is to identify the location of the paleoflow channels, and to determine whether or not there is ore potential. From a practical point of view, the objective is to find ore, particularly high-grade ore, and understanding the controls on ore will make this job much easier.

Textures and their interpretation

Lindgren's (1922) definition of the epithermal environment was based on vein textures (Table 3) and these criteria still serve us well. Textures in low-sulfidation veins include fine, crustiform bands of chalcedony, bladed quartz (due to carbonate-replacement) and open-space fillings. Indeed, even epizonal Archean lode-Au veins (e.g., Wiluna deposits, western Australia; Groves et al., 1998) possess similar textures, indicating their shallow depth of formation. These

textures are less common in disseminated low-sulfidation deposits, although they are still present in crosscutting veins. Colloform bands are most commonly formed at very shallow depths, caused by colloidal silica accumulations (Lindgren, 1933), and which host gold dendrites in high-grade low-sulfidation veins (Saunders, 1994). Hydrothermal brecciation of veins and wall rock is common in both low-sulfidation and high-sulfidation deposits, typically showing a jigsaw pattern of fragments.

By contrast, high-sulfidation deposits may lack many of these textures, except in late veins. Textures in high-sulfidation deposits are dominated by massive to vuggy silicic zones that are residual in origin, with the abundance and size of vugs dependent on the original nature of the host rock and the degree of silicification. Extreme acid leaching of a rock with coarse to fine phenocrysts or pumice and/or lithic fragments will develop a vuggy appearance, recognizable even if massively silicified. By contrast, an originally fine-grained and massive rock will maintain its appearance. Breccias are common in high-sulfidation deposits, as they are in low-sulfidation deposits, and it may be difficult to distinguish between altered breccias of volcanic or phreatomagmatic origin and breccias of syn-hydrothermal origin. It is common in high-sulfidation deposits to find comparatively higher ore grades in syn-hydrothermal breccia and its immediately surrounding rock.

The majority of the textures that are diagnostic of the epithermal environment involve silica in some form, and silicic zones tend to be more resistant to erosion than argillic alteration, thus forming topographic highs. This is particularly notable with silicic cores of vuggy quartz. Therefore, topographic highs, no matter how high or how steep, must be thoroughly examined for evidence of silicic alteration and textures, and for gold anomalies.

Host rocks, structure, and deposit form

Early in the assessment of any epithermal prospect, the likely controls on grade (i.e., the potential form of the body) must be determined, as this is one of the most basic characteristics of any ore deposit. Both high-sulfidation and low-sulfidation styles commonly have a strong structural control, although in disseminated or replacement deposits the structure may be concealed. Epithermal deposits are extremely variable in form (Fig. 7). Much of this variability is caused by strong permeability differences in the near-surface environment, resulting from lithologic, structural, and hydrothermal controls. Some of the largest high-sulfidation deposits (e.g., Yanacocha) have a large component of ore contained within favorable lithologies and can be considered disseminated. This is also true for the large low-sulfidation deposit of Round Mountain.

The grade-tonnage characteristics of both low-sulfidation and high-sulfidation deposits (Fig. 9) correlate closely with deposit form (Fig. 7). The largest tonnage deposits are dominated by disseminated or replacement ore as the result of lithologic control on fluid flow. In addition, these large deposits are economic most commonly because of supergene oxidation of the sulfides (e.g., at Yanacocha and

Pierina, Perú, and La Coipa and Pascua). By contrast, several deposits with greater than 3 Moz Au at grades in excess of 20 to 30 g/t are hosted by fractures. These high-grade vein deposits include end-member low-sulfidation (Hishikari, Midas, Sleeper), intermediate sulfidation-state (Comstock Lode), and high-sulfidation (El Indio, Goldfield, Chinkuashih) styles.

Practical recommendation: An efficient technique to quickly reach a preliminary, broad understanding of the order of magnitude potential of an epithermal prospect, including host rocks, structural control, deposit form, alteration relations, and ore, is to start exploration with a preliminary field transect of the entire exposed hydrothermal system. Assuming the terrain and/or vegetation cover allow, this should be done by traversing from fresh or weakly altered rock on one side of the system, to fresh or weakly altered rock on the opposite side, via the perceived core(s) of mineralization. Combined with preliminary alteration and mineralogical identification, possibly supported by PIMA (Portable Infrared Mineral Analyzer; see below) and previously generated photo-geologic maps, this early traverse will provide a wealth of information and help to focus exploration efforts.

Hydrothermal alteration

Assemblages and zoning: Alteration zoning is distinctly different between the two styles of epithermal system, low sulfidation and high sulfidation (Figs. 2 and 3). Although we generalize about such zoning, each prospect is an individual and, given the geologic variability, may vary from such generalized schema (cf. Buchanan, 1981). Although it is common to group alteration minerals into the standard assemblages (Table 2; Meyer and Hemley, 1967), we recommend first identifying the minerals present and their zonation, rather than mapping simply on the basis of an alteration type. There is much information to be gained from knowing specific mineralogy, information that is lost when lumping into a broad category, or perhaps never making the distinction in the first place ("split in the field, as you can always lump in the office" is good advice; see below). Furthermore, geologists are known to use such alteration terminology differently.

Type of advanced argillic alteration: Two of the most critical interpretations to make in the field are the origin of advanced argillic alteration (i.e., hypogene, steam-heated, or supergene) and the origin of silicic alteration (e.g., residual silica or silicification). If the advanced argillic alteration is of hypogene origin, the prospect may either be a barren lithocap or a mineralized high-sulfidation system. As discussed above and elsewhere (Meyer and Hemley, 1967; Rye et al. 1992; Sillitoe 1993a; Arribas 1995), hypogene advanced argillic alteration is defined by the presence of assemblages containing some or all of the following minerals: quartz, alunite, kandite minerals (kaolinite, nacrite, dickite), diaspore, pyrophyllite and zunyite. Advanced argillic alteration of steam-heated origin contains many of these minerals, particularly if these acid fluids descend along fractures and are heated (Reyes, 1990). In this con-

text, two observations may provide diagnostic evidence for a hypogene origin: (1) the presence of true residual silica (e.g., vuggy quartz formed by leaching of a porphyritic igneous rock), and (2) crystalline alunite with tabular to platy crystal habit and white to pink color (Table 6); these crystals can be recognized most easily in thin section or with the aid of a hand lens in vugs or open spaces.

There is a clear distinction between the two styles of deposit in the spatial relationship of the advanced argillic assemblage to the ore zone. In high-sulfidation deposits, the ore zone occurs within a silicic core and is hosted by a laterally varying halo of advanced argillic minerals. Closest to the residual quartz core is the common zone of quartz-alunite (Fig. 2). By contrast, the advanced argillic zone of low-sulfidation deposits forms a blanket above the water table due to steam heating and, hence, lies over the ore zone, although it may overprint ore during conditions of a falling water table (e.g., at Sulphur and Mt. Muro). Because of its shallow origin, the associated alteration minerals constitute opal-cristobalite, kaolinite (the low-temperature kandite mineral), and alunite (Fig. 8). Although steam-heated alteration may result in a porous texture, complete residual silica of the type formed by extremely acid hypogene leaching should not be found within steam-heated blankets other than in overprinted situations. Alunite formed in the steam-heated environment, as well as supergene alunite, shows a pseudocubic or rhombohedral crystal habit and is rarely coarse enough for individual crystals to be recognized with the aid of a hand lens (Table 6).

Steam-heated alteration blankets also form over high-sulfidation systems (e.g., La Coipa). Therefore, even if the low-temperature advanced argillic minerals form a blanket, this morphology does not distinguish the style of system. Neither does such a morphology and mineral assemblage constitute evidence for steam-heating, because supergene processes are also controlled by the water table, and result in a similar assemblage of minerals. In addition to the different textures between hypogene and steam-heated or supergene alunite (Sillitoe, 1993a), secondary or low-temperature minerals such as jarosite, scorodite, or halloysite (low temperature polymorph of kaolinite) help to identify supergene alteration (Fig. 8, Table 6). In some situations it is also possible to use sulfur isotope ratios to distinguish readily between hypogene and supergene or steam-heated advanced argillic alteration (Rye et al., 1992; Arribas et al., 1995a)

Silica and silicification: It is critical to distinguish the various origins of silicic alteration or silicification (Table 4) for two main reasons: (1) to understand the geometry and spatial relations of the hydrothermal system, and (2) to better understand epithermal ore, as ore is closely associated with various forms of silica products. In this context, the rapid cooling of a boiling fluid as it ascends near the surface results in silica deposition, either as quartz (at >200°C) or as a less-ordered polymorph such as chalcedony at less than 150° to 200°C or amorphous silica at 100° to 150°C (Fig. 4a). In addition, quartz deposition occurs only from relatively neutral-pH solutions, such as those ascending in low-sulfidation and intermediate sulfidation-state

TABLE 6. Characteristics of Three Types of Acid Alteration, Including Alunite, and their Distinction

	Hypogene chloride-sulfate	Steam-heated sulfate	Supergene
Genetic terminology (Rye et al., 1992)	Magmatic-hydrothermal	Steam-heated	Supergene
Origin (eqs. 5-9)	Condensation of high-T magmatic vapor with HCl + SO ₂ ascending	Atmospheric oxidation of H ₂ S in vadose zone above water table	Atmospheric oxidation of fine-grained sulfide within surficial weathering zone
Temperature range (°C)	300°-350° to 100°, ascending hypogene fluid	100°-120°, up to 150°+, descending fluid	20°-40°, overprinting descending fluid
Assemblages	Quartz, alunite, kaolinite, dickite, diaspore, pyrophyllite, zunyite	Kaolinite, alunite, opal, cristobalite, native sulfur	Kaolinite, halloysite, jarosite
Relationship to ore	Potentially ore-bearing, typically forms envelope to ore	Barren, above ore or overprint	Unrelated to sulfide ore, related to oxide ore
Alunite texture and crystal habit	Coarse to fine-grained crystalline aggregates; tabular or bladed crystals	Fine (<20-50 µm) powdery aggregates, chalky, porcellaneous masses; rhombohedral crystals with nearly cubic angles	Chalky, porcellaneous masses and fine (<20-50 µm) powdery aggregates; rhombohedral crystals with nearly cubic angles
Alunite color	Colorless, pink, white, cream, yellow, tan, brown	Generally white	White, cream, yellow-brown (jarosite staining)
Alunite isotopic composition (see Rye et al., 1992)	δ ³⁴ S alunite >> δ ³⁴ S associated sulfide (e.g., 20 vs. 0 per mil)	δ ³⁴ S alunite ~δ ³⁴ S associated sulfide (e.g., 3 and 2 per mil)	δ ³⁴ S alunite = δ ³⁴ S of precursor sulfide

Literature sources include Schoen et al. (1974), Anderson (1982), Rye et al. (1992), Sillitoe (1993a), Itaya et al. (1996)

veins. By contrast, hypogene leaching leaves a silica residue that recrystallizes to quartz, and this zone also may be silicified in later stages, typically before and/or during high-sulfidation ore introduction. Chalcedony deposits as a blanket at the water table, whereas opal forms within the vadose zone (Figs. 3 and 5). Lastly, silica sinter, which precipitates in the form of amorphous silica but eventually recrystallizes to quartz, provides unequivocal evidence for the location of the paleosurface at the time of mineralization.

Mineralogy: Knowledge of the temperature range over which epithermal alteration minerals are stable (Fig. 8), coupled with an interpretive geometry of fluid distributions in low-sulfidation and high-sulfidation systems, allows the alteration mineral distribution to be used to establish the paleoisotherms and, thus, paleofluid flow directions (Fig. 1). The latter aspect can help to establish the center of fluid flow and direction to mineral potential, and may indicate whether or not flow was structurally controlled (tight paleoisotherms) or pervasive (broad alteration halos). White et al. (1995) also stress the potential for lateral fluid flow in settings of high topographic relief, and an asymmetric alteration zonation may indicate this.

As an example of using minerals to estimate isotherms, the zonation of hydrothermal clays from smectite to illite (Fig. 2) matches the pattern of increasing thermal gradient in the Broadlands geothermal system, New Zealand (Simmons and Browne, 2000). Izawa et al. (1990) recognized a similar zonation of clay minerals around the surface projection of the high-grade veins at Hishikari. Paleotemperature esti-

mates can also provide information on the erosion level, although one must remember that isotherms drape over the upflow. If there are indications that a portion of the system was boiling, then the estimate of the paleotemperature may be used to estimate the paleodepth (Fig. 4a). Some of these indications, at least in a qualitative and relative sense, are summarized in Tables 4 and 5. Mapping the zonation of alteration minerals can be applied to early-stage projects, but it is typically a more useful approach during exploration of established deposits or districts, where a variety of three-dimensional mineralogical information is available.

Alteration minerals can provide much information about the composition of the fluids. As already noted, there is a suite of advanced argillic minerals that indicate a fluid was acid (Fig. 8), whereas the presence of adularia and calcite suggest relatively alkaline fluid, perhaps generated from a neutral-pH fluid by the loss of CO₂ during boiling. Zeolites also indicate somewhat alkaline conditions and, along with the occurrence of epidote, provide evidence for relatively low gas contents in the fluid. Calcite forms in place of zeolites from fluids of high CO₂ content. This observation of the presence of Ca minerals may be relevant if several properties are being compared, because it can be argued that indications of a high gas content are favorable for ore formation because this implies a high H₂S content and, thus, high gold solubility (eq. 4).

Practical recommendations: Remote sensing tools that can assist in broad-scale alteration studies include satellite-based thematic mapping (TM) and the new generation of

hyper-spectral airborne spectrometers, such as AVIRIS (Airborne Visible Infrared Imaging Spectrometer) and the SFSI (Short Wave Infra-Red Full Spectrum Imager; Neville and Powell, 1992; Neville et al., 1995; Staenz et al., 1999). TM provides a useful tool to survey large regions and identify areas of potentially favorable alteration. However, with one possible exception at Pierina (Volkert et al., 1998), TM has not been credited as a crucial tool in the discovery of an epithermal deposit (Sillitoe, 1995b). The high spatial and spectral resolution of the most recent hyper-spectral airborne instruments, such as SFSI, allow for the fast (although costly) production of detailed district and property-wide mineral maps at resolutions of 5 m or better. For example, SFSI operates in the short wave infra-red between 1,208 and 2,445 nm, and has 10 nm resolution, making it ideal for mapping the narrow absorption features of clay minerals associated with hydrothermal alteration (Staenz et al., 1999). These remote sensing techniques require good rock exposure, but also allow identification of lithologic and structural features that may help to focus field mapping on a critical set of questions or areas.

The most common tools for mineralogical analysis are X-ray diffractometry (XRD) and short-wave infrared (SWIR) spectroscopy (e.g., PIMA II). Both techniques are widely used and each has its advantages and disadvantages (cf. Thompson et al., 1999). The PIMA II is field portable and requires minimal sample preparation, allowing for a large number of measurements to be made rapidly and at relatively low cost. Alteration mapping can be done at the same time as mapping or drilling, improving the effectiveness of alteration data (Thompson et al., 1999). Samples for XRD must be submitted to central laboratories with longer turn-around time and higher costs. However, there are trade-offs; for example, portable SWIR spectrometers are limited in terms of the number of minerals they can identify. For this and other reasons, SWIR data must be collected by a skilled operator who has geologic training. The use of each technique (XRD or PIMA) reflects the preferences and previous experience of the explorationist. We

recommend that either tool be used primarily to help field geologists directly recognize in the field the alteration minerals relevant to the prospect being explored. Secondly, PIMA or XRD data will provide mineralogical information that can be processed later in different ways. In general, epithermal prospects can be understood more quickly and completely when mineralogical data are processed, interpreted, and simplified into relevant categories. Table 7 shows an alteration classification based on visual observations and field-generated PIMA data (about 150 measurements a day) that we used successfully in grassroots exploration of epithermal deposits in remote locations.

Ore controls and their indications

During exploration for low-sulfidation epithermal deposits, recognition of features such as silica sinter, steam-heated alteration, and the structural fabric of the prospect area will assist in the reconstruction of the paleo-geothermal system, and identification of the most favorable drilling or sampling locations. For example, evidence for boiling indicates proximity to upflow channels. This includes mineralogical evidence such as adularia, and in some situations, truscottite, and textures such as bladed calcite, commonly replaced by quartz pseudomorphs. In addition to identifying flow channels, evidence for boiling also indicates the occurrence of the mechanism that we argue is most favorable for deposition of gold in the low-sulfidation environment (eq. 4a). However, in many deposits there is a spatial separation between the gold ore zone and indicators of boiling (Simmons and Browne, 2000). This may be caused either by the delay in gold saturation on initiation of boiling, or in the extreme case, by physical transport of gold colloids from the site of saturation, as indicated by gold dendrites. In addition, barren bladed calcite also forms late in the life of the system from marginal waters collapsing inward (Simmons et al., 2000), when there are no metals present in the fluid.

High-grade epithermal deposits, both low-sulfidation and high-sulfidation, typically are fracture controlled and

TABLE 7. Alteration Assemblages Based on Visual Identification Aided by PIMA

Mineral assemblage	Alteration type
Quartz dominant (e.g., >90%)	Silicic
Alunite ± quartz dominant	Advanced argillic ¹
Mixtures of quartz with alunite, and other acid minerals such as dickite, kaolinite, pyrophyllite, diaspore	Advanced argillic ²
Pyrophyllite ± dickite ± quartz dominant	Advanced argillic ³
Illite ± quartz dominant	Argillic (high temperature)
Interstratified illite-smectite dominant	Argillic (moderate temperature)
Smectite dominant	Argillic (low temperature)
Epidote present	Propylitic (high temperature)
Chlorite ± calcite present	Propylitic (low temperature)

¹Typically hypogene; note textures

²May be hypogene, steam-heated, or supergene; note textures and morphology

³Deep hypogene

have sharp lower and upper limits, suggesting the influence of a process such as boiling on gold saturation. In the case of deposits with dendritic gold, Saunders (1994) concludes that deposition occurred from gold colloids when a rapidly ascending fluid slowed near the surface, or in the case of Hishikari, reached a permeable unit. In addition, some deposits can have low-grade tops to the ore zone in the form of veins (e.g., Comstock Lode) or disseminated bodies (Ivanhoe). In the case of the Comstock Lode, outcrops of the Oriental vein consist of quartz and adularia, but contain only 50 to 150 ppb Au, despite the quartz-calcite and quartz-adularia ore zones rising to within 30 m of these outcrops (D.M. Hudson, pers. commun., 2000). Individual ore zones in this district have vertical intervals of about 150 m, but their tops vary by as much as 800 m (although much of this range may have been caused by fault offset after vein formation; D.M. Hudson, pers. commun., 2000). In the case of the Ivanhoe district, the newly discovered Clementine and Gwenivere vein systems in basement rocks represent the high-grade feeders to low-grade, volcanic-hosted disseminated ore of the type previously mined at the Hollister deposit (Northern Miner, 2000).

A good understanding of the origin of the faults and fractures is always important, but it may be critical during exploration for fault-controlled high-grade deposits. The alteration halos of these fractures or faults may be narrow and of little use from a prospecting point of view. In the case of deposits with two contrasting lithologies occurring within the ore zone, such as older metamorphic basement and younger volcanic cover, no hard rules can be applied as to the location of high-grade ore. At Hishikari (Izawa et al., 1990) and Ivanhoe (John et al., 1999), the high-grade ore that constitutes much of the contained gold is found in the Cretaceous and Ordovician metasedimentary basement rocks, respectively. However, the opposite is true at Sleeper (Nash and Trudel, 1996) and Midas (John et al., 1999), with the high-grade gold occurring within the Tertiary volcanic cover. The thickness of the volcanic package and depth to basement are likely factors that control the host of high-grade veins, located within or beneath a disseminated, low-grade halo.

In high-sulfidation deposits, ore typically is controlled by hydrothermal products such as hydrothermal breccia or bodies of residual vuggy quartz (Fig. 7; Sillitoe, 1993a). Both provide permeability that may be localized along lithologic or structural boundaries. Examples of lithologic controls include gold dispersed in ignimbrite (Yanacocha, Pierina) or permeable clastic or carbonate sedimentary rocks (La Coipa, and San Gregorio, Perú, respectively), as a replacement at a lithologic unconformity (Lepanto), or dispersed in diatreme breccia. Examples of structural control include massive veins or vein swarms (El Indio, Chinkuashih, Lepanto) and low-angle veins (Goldfield).

Practical recommendation: Estimating the degree and direction of hydraulic gradients is critical in both low-sulfidation and high-sulfidation prospects, whether permeability is lithologic or structural, because the ascending fluid is influenced by topography soon after it reaches epithermal

depths. This may not be a trivial exercise, but the potential rewards make it a worthwhile project, particularly in advanced exploration targets where a larger amount of information may be available. In low-relief paleo-settings the amount of lateral flow may be relatively small, with surface discharge to topographic lows such as lakes and streams. In high-relief settings such as stratovolcanoes, however, lateral flow can extend for many kilometers (Henley and Ellis, 1983), resulting in a large and asymmetric zone of alteration. The secret is in determining where the upflow was focused, because this is most likely the place to start the search for ore.

Depth of erosion and paleodepth indicators

With the discovery of the low-sulfidation deposit of McLaughlin, beneath a silica sinter, the "hot-spring deposit" term (Table 1) was coined and the rush was on for exploration geologists to identify sinters in their search for other such deposits. As discussed above, the criteria for identifying silica sinter (White et al., 1989) extend beyond finely laminated siliceous material, because this feature is also typical of silicified air-fall tuffs and lacustrine sediments. For this reason, care must be taken in interpreting the origin of such outcrops, because misidentification can have negative consequences to the model of the prospect being constructed. An example would be deciding that a prospect was low sulfidation on the basis of sinter, when in fact the fine siliceous laminae had a lacustrine origin in a volcanic-hydrothermal crater with high-sulfidation potential. In both cases, the potential for ore would be at depth, but the guides for its location could differ significantly.

Silica sinters define the paleosurface of the low-sulfidation environment and also indicate the location of a principal upflow conduit. Sinter may contain anomalous As, Sb, Hg, and Tl, and even Au and Ag, although whether or not these elements are concentrated may depend on local features, such as vent shape, more than on basic characteristics, such as the metal content of the thermal water (Hedenquist, 1991).

In both low-sulfidation and high-sulfidation systems, the discontinuous blanket of steam-heated kaolinite-smectite \pm alunite \pm native S produces an opaline rock that is friable and easily eroded. Thick blankets may form where there is a syn-hydrothermal fall in the water table, and this alteration may also descend along permeable zones. At the base of the vadose zone, finely laminated sedimentary units such as air-fall tuffs and other permeable units serve as aquifers for outflow of the acid water along the paleowater table, causing silicification in the form of chalcedony (Fig. 3). The opaline cap to the chalcedony horizon forms only in the area where steam-heated water is generated, whereas the chalcedony horizon can extend far outside the system as the result of outflow along aquifers, forming a thick and laterally extensive horizon.

As just noted, finely laminated, in some cases crossbedded siliceous deposits accumulate in acid lakes in both low-sulfidation and high-sulfidation systems, and these lakes may fill and thus define hydrothermal or even volcanic

eruption craters. These and other lakes also accumulate clastic sediments that may be interbedded with the siliceous material, the latter depositing from a colloidal suspension of silica in an acid lake.

Disseminated and replacement deposits, both low-sulfidation and high-sulfidation, are dominated by quartz, either residual (high sulfidation) and/or silicification (low sulfidation, high sulfidation) in origin. As noted above, if erosion exposes these shallow-formed zones, these bodies will be very resistant and topographically steep, as seen at Chinkuashih, Yanacocha, and the Mexican deposits of Sauzal and Mulatos, plus the Ladera-Farellón orebodies at La Coipa, among others. However, where erosion is slight, as indicated by evidence of paleowater table preservation, the silicic core may not crop out, or it may crop out only along permeable horizons, such as an unconformity some distance from the upflow (e.g., Lepanto, Pierina, and Coipa Norte orebody at La Coipa). Similarly, low-sulfidation quartz veins will also crop out, particularly because their argillic halos erode relatively quickly.

Without additional information from a geologic reconstruction or indications of paleotemperature, vein textures provide a relative indication of the paleodepth. Open-space cockade textures decrease with increasing depth, and the abundance of chalcedony also decreases with a concomitant increase in the crystallinity of quartz; fine banding also disappears with increasing depth. White et al. (1995) note that quartz is more common than chalcedony in Philippines deposits compared to those in Japan and New Zealand, suggesting a greater depth of formation and more erosion in the high-relief Philippines setting. By contrast, the lack of silicification, or the presence of hydrothermal minerals such as biotite and amphibole (Fig. 8), indicates high temperature and, thus, deep erosion, close to or below the lower limit of the epithermal environment. However, this is a guide only, because some deposits (Comstock Lode and Kelian, Indonesia) apparently formed at greater than 280°C, a higher temperature, and thus, a greater depth than is typical for most epithermal deposits (Figs. 4b and 8).

The deeper portion of the high-sulfidation environment may encroach on porphyry systems. In porphyry copper environments, lithocaps may be separated by as much as 1 km from the underlying porphyry ore, or the lithocap may overprint the orebody if there has been sufficient telescoping and collapse of shallow alteration onto the deep system (Sillitoe, 1995a). The occurrence of high-temperature minerals such as pyrophyllite, andalusite, and corundum, and lodes of enargite or bornite, indicate that the exposed portion of the lithocap is deep, and near an underlying porphyry system. The absence of a porphyry stock or early veins of vitreous, anhedral quartz, commonly termed A-veins, suggests that the porphyry may be deeper, or laterally offset. The occurrence of pyrophyllite ± sericite as a laterally extensive or thick zone of alteration, with or without sulfide veins, may indicate a position beneath the quartz-alunite lithocap. However, the intrusive complexity of porphyry systems must be kept in mind. Added to this complexity is the rapid erosion or even sector collapse of some

volcanoes that can cause an epithermal style of overprint on a deep porphyry system (Sillitoe, 1994).

Potential for related deposits within a district

There is potential for a variety of related deposits in epithermal districts. For example, there is clear evidence for a spatial, and in some cases genetic relationship between high-sulfidation epithermal deposits and underlying or adjacent porphyry deposits (Sillitoe, 1983, 1999; Arribas et al., 1995b). Skarns are present in some districts with calcareous lithologies. Calcareous rocks have also been recognized recently to host ore adjacent to high-sulfidation deposits (San Gregorio; Fontboté and Bendezú, 1999). There is also speculation on the relationship between porphyry-epithermal deposits and some Carlin-like ore bodies in Nevada (Sillitoe and Bonham, 1990). By contrast, end-member low-sulfidation deposits appear to form in a geologic environment with little porphyry or high-sulfidation potential (John et al., 1999).

There is increasing recognition of the potential for base metal ± Au-Ag veins adjacent to high-sulfidation deposits, for example, the recent discovery of the multi-million oz gold veins of Victoria, adjacent to Lepanto (Cuizon et al., 1998; Claveria et al., 1999). These quartz veins have sericite halos, rhodochrosite and rhodonite gangue, and a sulfide assemblage similar to intermediate sulfidation-state deposits elsewhere (Table 3). In addition, the ore-mineral assemblage is very similar in composition to the post-enargite gold-rich stage of Lepanto (Hedenquist et al., 1998). This raises the speculation that there is a similar chemical evolution of the fluid responsible for intermediate sulfidation-state veins and the post-enargite, Au-rich stage of high-sulfidation deposits.

Practical recommendation: Epithermal veins with intermediate sulfidation-state characteristics have been found adjacent to some high-sulfidation ore deposits. This empirical observation of a spatial association (Sillitoe, 1993a, 1999) should encourage exploration around known high-sulfidation deposits for such intermediate sulfidation-state veins, and even in the vicinity of barren lithocaps. Conversely, one may also predict potential for high-sulfidation and/or porphyry deposits to occur near known intermediate sulfidation-state veins.

Weathering and supergene processes

Supergene oxidation, and in places enrichment, have affected the large replacement-type high-sulfidation deposits, as well as disseminated low-sulfidation deposits in regions of arid and semi-arid climate, for example, the western Americas (Sillitoe, 1999). By contrast, relatively few epithermal deposits have been equally affected in the tropics, where water tables lie close to the surface. The extent of supergene oxidation is controlled by the climate, permeability of the rock, rates of uplift, etc. For reasons of permeability, such oxidation extends to depths of 400 m at Yanacocha within the silicic (vuggy quartz) zones, whereas fresh sulfide occurs at the surface in adjacent argillic zones (Harvey et al., 1999). The oxide-sulfide interface typically is subhorizontal because the position of the paleowater table is a critical control (Chávez, 2000).

The effect of oxidation on metal recovery by cyanide leaching can be dramatic, from a recovery of more than 90 percent gold for oxidized ore, to less than 35 percent for non-oxidized ore. If the gold is distributed mainly along fissures and fractures (e.g., Yanacocha and Cerro Crucitas), rock that is only partly oxidized may still yield acceptable recovery rates.

Supergene oxidation allowed the economic mining of part of the Pueblo Viejo high-sulfidation orebody, because it affected about one-third of the overall deposit. The balance of the deposit, about 100 Mt at ~4 g/t Au, occurs as refractory sulfide material and remains a resource only, awaiting technological advances that will allow economic processing of the gold-bearing sulfide. Soil samples collected here at the bedrock surface are a good indication of the gold content of the underlying oxidized rock, whereas silver was significantly depleted (Russell et al., 1981). At the La Coipa high-sulfidation Ag-Au deposit most of the silver occurs as secondary halides and native silver. Nevertheless, despite the steep topography, both gold and silver anomalies in talus fines define the area of the ore bodies (Oviedo et al., 1991), indicating that these two metals were not highly mobile in the supergene environment. By contrast, there is evidence that copper was significantly remobilized during supergene alteration.

The El Indio high-sulfidation Au-Ag-Cu deposit consists of a series of veins (Jannas et al., 1990, 1999) that have been little affected by supergene processes except within several meters of the surface. Talus fines are highly anomalous in Au, Ag, As, and Pb, and correlate well with each other, whereas Cu and Zn were depleted because of supergene leaching. Arsenic is more widely dispersed than gold owing to its occurrence as scorodite and other supergene arsenates, formed when the veins of massive enargite and pyrite were oxidized (Siddeley and Araneda, 1986).

Turner (1997) compared the geochemistry of primary sulfide and supergene oxide samples from Yanacocha core. He found little evidence for enrichment or depletion of gold in the oxide zone, whereas Ag can exceed 1,000 ppm in some areas, despite concentrations of only ~10 ppm in sulfide ore. There is a weak correlation of Au with As, Sb, Hg, and Ba, with the poorest correlation in deposits that are all oxide, suggesting some secondary mobility of this suite of elements. Barite is an insoluble gangue mineral that commonly accompanies high-sulfidation sulfide mineralization, and it remains even after complete oxidation of sulfides.

Supergene modification of low-sulfidation veins, particularly those with a low sulfide content, is limited, particularly in regions of rapid uplift (i.e., most of the prospective arcs of the circum-Pacific). Supergene minerals are rare and low in abundance in epithermal deposits of the tropical southwestern Pacific (White et al., 1995), and are only slightly more common in southwestern North American deposits (Buchanan, 1981). The most typical minerals in arid areas are cerargyrite (AgCl) and iodargyrite (AgI), both in high-sulfidation and low-sulfidation deposits, whereas limonite, scorodite, jarosite, and goethite occur primarily in high-sulfidation deposits, largely in arid terrain but also locally in some tropical areas along with Mn oxides.

Atmospheric oxidation of sulfides during weathering of high-sulfidation deposits (e.g., Goldfield, Rodalquilar, Summitville) has led to the development of surficial acid-sulfate blankets. Supergene alunite at Rodalquilar and elsewhere forms white to yellow porcellaneous veins consisting of fine-grained (commonly less than 50 μm) pseudo-cubic alunite with subordinate kaolinite, jarosite, and hydrated amorphous silica (Table 6). Oxidation at Rodalquilar occurred as a result of dramatic drop of the water table and increased erosion during partial desiccation of the Mediterranean Sea during the late Miocene. In the Andes as well as the western United States, formation of supergene alunite during the middle to late Miocene appears to be related to regional tectonic episodes (Sillitoe and McKee, 1996).

Noble et al. (1997) suggested that hypogene oxidation caused the destruction of sulfides at Pierina, and also introduced gold. A similar oxidation process has been suggested for the Sauzal deposit in Mexico. Our field observations on both deposits do not suggest processes any different from those ascribed to weathering of many other high-sulfidation deposits. Sillitoe (1999) outlines a series of arguments against hypogene oxidation, and notes that the covellite rims on remnant sulfide bodies at Pierina are equivalent to the oxidation front in any copper deposit that has been affected by supergene processes. Similar rims or patinas were observed in drill core beneath the well-documented supergene zone at Rodalquilar (Arribas et al., 1995a).

Practical recommendation: Documentation of supergene alteration processes and features should be a critical component of exploration programs for high-sulfidation deposits. As mentioned above, however, the presence of sulfide at the surface within the alteration zone (e.g., with clay-altered rock) may not indicate the depth of the oxidation front within the silicic core (which extends to depth of 400 m at Yanacocha Sur). Distinction of advanced argillic alteration of supergene origin is important because it commonly masks the core of hypogene alteration, and can also result in a large and misleading alteration anomaly due to oxidation of fine-grained disseminated pyrite within a weakly argillic-altered and barren zone.

Geochemistry

A review of the discovery history of 54 hydrothermal deposits in the circum Pacific (Sillitoe, 1995b) noted that geochemical sampling played a principal role in the discovery of 37 of the deposits. However, of the 20 epithermal low-sulfidation and high-sulfidation discoveries, geochemistry was significant in all but two, with the discoveries at Bullfrog, Nevada and Hishikari based in part on minor workings from nearly a century earlier. Of the 18 deposits where geochemical sampling was a factor, rock chip sampling was important at 16 prospects, and soil, talus, and drainage sampling at four, three, and two, respectively; a combination of geochemical anomalies was employed at eight of the discoveries.

Drainage sampling can be an effective tool to first identify epithermal and porphyry prospects, particularly where cover, either rock or vegetation, masks the alteration zone. This technique was a major factor leading to discovery of

Kelian and Gunung Pongkor, both in the tropics of Indonesia. The -80 or -200 mesh fractions of stream sediments and panned concentrates were both anomalous at Kelian, but only the -80 mesh fraction was anomalous at Gunung Pongkor (Basuki et al., 1994). The BLEG (bulk leach extractable gold) method, developed for the arid drainages of Australia, served to identify the Batu Hijau porphyry Cu-Au system, as well as the seafloor high-sulfidation gold orebody of Lerokis, both deposits in eastern Indonesia. A regional drainage survey over 25,000 km² in northern Peru by the British Geological Survey first identified Cu and Pb-Zn-Ag anomalies in the region including Yanacocha. There are variable stream-sediment anomalies in the Yanacocha area for elements such as As, Pb, and Cu, and an early survey of gold showed that it is anomalous in both -80 mesh and BLEG samples within several kilometers of the orebodies (Turner, 1997).

Geochemical anomalies have been used successfully during epithermal exploration despite the range of settings, from arid, temperate, or tropical climates, although talus sampling is confined to the arid portions of the Andes. Such talus sampling was critical in definition of the high-sulfidation deposits at El Indio, Pascua and La Coipa, whereas rock-chip sampling was the method used at Yanacocha. Indeed, at Yanacocha only three out of every 100 rock samples first collected contained over 500 ppb Au, and did not define coherent anomalies. It was not until chipping was expanded to encompass 6 m centers that the surface anomalies defined the present ore bodies (Turner, 1997). This is now the standard method to assess the potential of silicic cores of high-sulfidation prospects, in some cases chipping over 12 m centers or more in order to homogenize the micro-fracture control on gold distribution (e.g., at Pierina). Of interest is the early introduction of anomalous gold at Yanacocha, most likely during the initial leaching stage, with broad areas of 50 to 200 ppb Au accompanying the silicic event with minor pyrite (Harvey et al., 1999).

Blankets of chalcedony are common in low-sulfidation and high-sulfidation prospects, and in most cases they are not geochemically anomalous except for Hg, commonly as cinnabar or meta-cinnabar. Exceptions can occur when a falling water table results in the chalcedony horizon overprinting a zone of underlying mineralization. Nevertheless, a chalcedony blanket in a position proximal to the upflow zone may create a barren cover, causing an orebody to be blind.

Practical recommendation: In addition to analyzing for Au and Ag, a standard suite of As, Sb, Hg \pm Tl, Se is typically employed during epithermal exploration, expanded in high-sulfidation terrain to include elements such as Ba, Cu, Zn, Pb, Mo, Te, Sn, and Bi. It is not possible to assign a suite of elements that is always anomalous, either in the epithermal deposit itself or in dispersion trains; nor is it possible to generalize the concentrations of individual elements that constitute anomalous levels. The variation within the epithermal environment causes a huge variability in geochemical signatures. These include differences between high-sulfidation and low-sulfidation systems, the wide variety of basement and host-rock types, the compositional vari-

ations of related igneous activity (Sillitoe et al., 1998), and the variable effects of climate, supergene oxidation, erosion rate, depth of erosion, etc. For these and other reasons, an orientation survey is essential at the start of exploration in every district to determine what type(s) of samples, metal suite, and threshold values are appropriate.

Geophysics

Geophysical techniques that are applicable to the assessment of epithermal deposits are outlined below; White et al. (1995) and Sillitoe (1995b) also review these techniques. The timing and appropriateness of their use varies from prospect to prospect, and depends on many factors from regional setting to deposit style. The important aspect to remember is that geophysics provides a variety of tools, one or more of which may contribute information to the overall exploration model being constructed.

Sillitoe (1995b) notes that geophysics played a surprisingly minor overall role in the discovery histories of 54 hydrothermal deposits in the circum-Pacific, with only seven of the discoveries resulting directly from the integration of geophysical results with geologic or geochemical surveys. Of these, only one such discovery was epithermal, the Hishikari low-sulfidation vein deposit. By contrast, geophysical surveys were more useful in the post-discovery delineation stage, being considered important in one-quarter of the developments. The use of a geophysical tool was critical in three of the epithermal delineation projects. All three were low-sulfidation veins where resistivity anomalies defined with induced polarization (IP) surveys were used to help site drillholes. For example, an IP survey at Gunung Pongkor helped to define the strike and continuity of low-sulfidation quartz veins hosted by argillic alteration (Basuki et al., 1994).

Airborne magnetic surveys provide information about regional setting, geologic boundaries, and structures. Magnetic highs are associated with magnetite-bearing I-type intrusions, and the method also allows the detection of magnetite associated with the biotite cores of porphyry systems. In addition, the method provides information about hydrothermal alteration, such as the destruction of primary magnetite signatures of igneous rocks by high-sulfidation and low-sulfidation systems, resulting in a magnetic low. Ground magnetic profiles over hydrothermally altered rocks are smooth owing to destruction of near-surface magnetite in the epithermal systems.

Electrical methods are particularly well suited to assessing the silicic core of high-sulfidation deposits and also barren lithocaps, because the vuggy quartz is typically a strong electrical resistor in dry conditions. This silicic core can be detected clearly with the IP method, although chalcedony blankets may interfere. The method can determine the shape, and lateral or depth extent, when using 3-D interpretation tools. The lateral extent of the silicic body can be estimated from IP traverses, although several traverses are necessary, particularly on the uphill side of the target. There may be an IP chargeability response from pyrite if sulfides are not oxidized, and conductive clays in the

advanced argillic and argillic zones should also be detected. Argillic halos to low-sulfidation veins also can be detected.

Giant Epithermal Gold Deposits: Are There Fundamental Controls?

There are at least 18 epithermal deposits around the Pacific rim that are known to contain a total of greater than 200 tonnes Au (~7 Moz; Sillitoe, 1997). Six of these deposits are high sulfidation in style (Pueblo Viejo, Yanacocha, Pierina, El Indio, and Pascua-Lama plus Veladero, Argentina) and, given the recent exploration successes for this style of deposit, more are sure to be discovered. Are there fundamental controls on the formation of such giant epithermal gold deposits, controls that may help guide exploration efforts?

Most of the largest porphyry Cu-Au deposits are similar to each other in overall character, whereas the large epithermal deposits are highly variable, reflecting the geologic variability within the near-surface, epithermal environment. This variability includes the lithology and permeability of the rocks themselves, the structural development, the nature of the volcanic edifice, topographic relief, and the reactivity of the fluid between low-sulfidation and high-sulfidation systems. Sillitoe (1997) speculated on the fundamental controls on formation of large deposits, and suggested that there are no unambiguous answers, for porphyry or epithermal systems, although both appear to be associated with atypical arc settings near the end of their tectonic development. In order to form a huge gold deposit, the starting point is most likely in the mantle, which is the ultimate source of metals (Hedenquist and Lowenstern, 1994). Partial melting of stalled lithospheric slabs in the mantle, soon after collision or arc migration, may promote the oxidation of the mantle and release of gold. Decompression will facilitate melting of the mantle, possibly related to isostatic changes. In turn, rapid uplift and crystallization at shallow crustal depths cause the exsolution of Au-bearing fluid, unless gold is incorporated in magnetite, pyrrhotite, or chalcopyrite prior to fluid saturation (Rowins, 2000, and references therein).

A disproportionate number of giant gold deposits are associated with relatively rare, highly potassic igneous rocks. Richards (1995) examined the association of epithermal gold deposits with alkalic igneous rocks. These magmas contain a high alkali-element content, and also tend to be high in volatiles, with particularly high SO_2 . Rocks as diverse as syenites, trachytes, phonolites, and shoshonites are all alkalic, although these rock types, and the characteristics of the associated deposits, are not separated by a hard boundary from other igneous rocks and epithermal deposits. The ore deposits associated with alkalic rocks include Ladolam, Porgera, Mount Kare in Papua New Guinea, Emperor in Fiji, those in the Montana alkalic province and Colorado Mineral Belt, and Cripple Creek, Colorado. Jensen and Barton (2000) discuss deposits associated with alkalic rocks in more detail.

The alkalic magmas associated with these large gold deposits appear to be products of post-subduction tectonic adjustments, or back-arc activity. Partial melting of meta-

somatized mantle, as also noted by Sillitoe (1997), may be important in creating a hydrous, oxidized melt at mantle depths, with ascent of this melt to shallow crustal depths facilitated by intersections of trans-crustal or trans-lithospheric structures. The oxidized magma prevents sulfide saturation, leaving metals available to be exsolved upon water saturation. Their oxidized nature is reflected in their $\text{SO}_2/\text{H}_2\text{S}$ ratios, which are typically as high as 10, compared to ratios of 1 to 10 for calc-alkaline volcanic discharges (Hedenquist, 1995); in addition there is a low HCl content from alkaline magmatic discharges. The low-sulfidation epithermal deposits associated with alkalic magmatism have isotopic signatures indicating a large magmatic component in the hydrothermal system, larger than is typical for other types of low-sulfidation epithermal systems (O'Neil and Silberman, 1974; Simmons, 1995). To date, high-sulfidation counterparts to these low-sulfidation deposits have not been recognized.

Lithocaps of silicic and advanced argillic alteration occur in little-eroded portions of the Bolivian Sn-Ag belt, where mineralization is centered on felsic domes (Sillitoe et al., 1998). The vuggy quartz portion of the lithocap at Potosí contained the world's largest Ag resource, 86,000 tonnes (2,800 Moz) prior to exploitation. Tin and base metal-bearing massive sulfide veins are hosted by sericitic alteration underlying these lithocaps, although the sulfides differ from Au-rich high-sulfidation systems in having a relatively low-sulfidation state. Sillitoe et al. (1998) argue that magma chemistry, reduced ilmenite-versus oxidized magnetite-type intrusions, is the fundamental control on whether a lithocap is mineralized by Ag-Sn-Sb or Au-Cu-As, respectively.

Both high-sulfidation and intermediate sulfidation-state epithermal deposits occur in andesite-rhyodacite arc terrain, commonly in association with porphyry deposits. By contrast, true end-member low-sulfidation deposits have no apparent spatial association with high-sulfidation deposits. Indeed, there are no high-sulfidation deposits in the bimodal rhyolite-basalt rift terrain of northern Nevada (John et al., 1999). John (1999) found that the oxidation state of magmas associated with low-sulfidation deposits (Sleeper, Midas) of the northern Nevada rift are 3 to 4 orders of magnitude more reduced than magmas associated with the high-sulfidation and intermediate sulfidation-state deposits of the western andesite arc of Nevada (Goldfield, Comstock Lode). John noted that this difference in magma oxidation state matches the difference in oxidation state of the sulfide assemblages, pyrite-pyrrhotite-arsenopyrite in the low-sulfidation deposits, compared with enargite-tennantite-chalcopyrite-pyrite and tennantite-chalcopyrite-pyrite in the high-sulfidation and intermediate sulfidation-state deposits, respectively.

Clearly magmas do more than simply drive convection cells of meteoric water during the formation of epithermal deposits. There is abundant evidence for magmatic components in the ore fluid during formation of high-sulfidation as well as low-sulfidation deposits (Simmons, 1995; Arribas, 1995; Cooke and Simmons, 2000). In addition, there is also a fundamental relationship between oxidized versus reduced magmas and the style of epithermal deposit, high-

sulfidation and intermediate sulfidation-state versus end-member low-sulfidation, respectively. Metal complement, Au-Cu-As versus Ag-Sn-Sb, is also related to the oxidation state of the magma. These empirical observations allow the explorationist to predict the style of epithermal deposit to expect in a district. Continued research on the relation between magma chemistry and epithermal as well as porphyry mineralization may eventually lead to criteria that can help to identify whether or not a magma exsolved an ore fluid, information that may help rank the prospectivity of a district. Such research may also eventually identify the factors that combine to form the giant deposits.

Concluding Remarks

A recent review of the discovery histories of 54 hydrothermal ore deposits around the Pacific Rim focused on the factors that led to discovery (Sillitoe, 1995b). Twenty epithermal deposits discovered between 1970 and 1995 were examined, including 14 low-sulfidation deposits and 6 high-sulfidation deposits. The lessons learned from these and other case studies are surprising but nevertheless reassuring in this increasingly high-technology world. Geologists working in the field on long-term projects in prospective regions with consistent support from management, leading up to and including a liberal amount of drilling, were responsible for a majority of recent discoveries. Although some of the greatest discoveries were made in virgin terrain, there was much success from exploring in known districts near known deposits.

For effective exploration it is essential to maximize the time in the field of well-trained and experienced geologists using tried and tested methods. Understanding the characteristics of the deposit style being sought allows multiple working hypotheses for a prospect to be constructed; this leads to ways to efficiently test each model of the prospect, using the tools appropriate for the situation. Technological panaceas should be avoided, because there is no evidence that they have yet replaced creative thinking and hard work by geologists in the field.

Epithermal precious-metal deposits form a grouping, albeit with large variations. The term epithermal was introduced almost 80 years ago (Lindgren, 1922), but remains applicable to the same deposits and to the many more deposits discovered since then. The basis for the strength of the term, and the validity of dealing with these deposits as a separate group, is the recognition of hydrothermal processes that occur near the surface, related in most cases to the shallow emplacement of igneous intrusions.

The early deduction of epithermal processes was based on careful observations and remarkable intuition. Today this understanding has been augmented by a genetic framework based on fundamental research into active geothermal and volcanic-hydrothermal systems, as well as studies of ore deposits. Understanding these hydrothermal processes and their products are directly applicable to the exploration for and assessment of epithermal deposits. However, such technical expertise must be combined with an understanding of the economic objectives of a company and then driven by an enthusiasm to imagine, pursue and test the unknown.

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REFERENCES

- Anderson, J.A., 1982. Characteristics of leached capping and techniques of appraisal, *in* Titley, S., ed., *Advances in geology of porphyry copper deposits: Southwestern North America*: Tucson, The University of Arizona Press, p. 275–296.
- Arribas, A., Jr., 1995. Characteristics of high-sulfidation epithermal deposits, and their relation to magmatic fluid: *Mineralogical Association of Canada Short Course Handbook*, v. 23, p. 419–454.
- Arribas, A., Jr., Cunningham, C.G., Rytuba, J.J., Rye, R.O., Kelly, W.C., Podwysocki, M.H., McKee, E.H., and Tosdal, R.M., 1995a. Geology, geochronology, fluid inclusions, and isotope geochemistry of the Rodalquilar gold-alunite deposit, Spain: *Economic Geology*, v. 90, p. 795–822.
- Arribas, A., Jr., Hedenquist, J.W., Itaya, T., Okada, T., Concepción, R.A., and García, J.S., Jr., 1995b. Contemporaneous formation of adjacent porphyry and epithermal Cu-Au deposits over 300 ka in northern Luzon, Philippines: *Geology*, v. 23, p. 337–340.
- Ashley, R.P., 1982. Occurrence model for enargite-gold deposits: U.S. Geological Survey Open-File Report 82-795, p. 144–147.
- Barton, P.B., Jr., and Skinner, B.J., 1979. Sulfide mineral stabilities, *in* Barnes, H.L., ed., *Geochemistry of hydrothermal ore deposits*, 2nd ed.: New York, Wiley-Interscience, p. 278–403.
- Basuki, A., Aditya Sumanagara, A., and Sinambela, D., 1994. The Gunung Pongkor gold-silver deposit, West Java, Indonesia: *Journal of Geochemical Exploration*, v. 50, p. 371–391.
- Benning, L.G., and Seward, T.M., 1996. Hydrosulfide complexing of Au(I) in hydrothermal solutions from 150°–400°C and 500–1500 bar: *Geochimica et Cosmochimica Acta*, v. 60, p. 1849–1871.
- Berger, B.R., and Eimon, P., 1983. Conceptual models of epithermal precious metal deposits, *in* Shanks, W.C., ed., *Cameron volume on unconventional mineral deposits*: New York, AIME Society of Mining Engineers, p. 191–205.
- Berger, B.R., and Henley, R.W., 1989. Advances in understanding of epithermal gold-silver deposits, with special reference to the western United States: *Economic Geology Monograph* 6, p. 405–423.
- Bethke, P.M., 1984. Controls on base and precious metal mineralization in deeper epithermal environments: U.S. Geological Survey Open-File Report 84-890, 40 p.
- Bonham, H.F., Jr., 1986. Models for volcanic-hosted epithermal precious metal deposits: A review: *International Volcanological Congress, Symposium 5, Hamilton, New Zealand, February 1986, Proceedings*, p. 13–17.
- Brown, K.L., 1986. Gold deposition from geothermal discharges in New Zealand: *Economic Geology*, v. 81, p. 979–983.
- Buchanan, I.J., 1981. Precious metal deposits associated with volcanic environments in the Southwest: *Arizona Geological Society Digest* 14, p. 237–262.
- Capps, R.C., and Moore, J.A., 1991. Geologic setting of mid-Miocene gold deposits in the Castle Mountains, San Bernardino County, California and Clark County, Nevada: *Geological Society of Nevada, Symposium Proceedings*, v. 2, p. 1195–1219.
- Chávez, W.X., Jr., 2000. Supergene oxidation of copper deposits: Zoning and distribution of copper oxide minerals: *Society of Economic Geologists Newsletter*, no. 41, p. 1, 10–21.

- Claveria, R.J.R., Cuisson, A.G., and Andam, B.V., 1999, The Victoria gold deposit in the Mankayan mineral district, Luzon, Philippines: PacRim '99, Bali, Indonesia, 10–13 October, Proceedings, p. 73–80.
- Cooke D.R., and Simmons, S.F., 2000, Characteristics and genesis of epithermal gold deposits: Society of Economic Geologists, Reviews in Economic Geology, v. 13, p. 221–244.
- Cuisson, A.L.G., Claveria, R.J.R., and Andam, B.V., 1998, The discovery of the Lepanto Victoria gold deposit, Mankayan, Benguet, Philippines: Geocoin 98, 11th Annual GSP Geological Convention, Manila, Philippines, 1998, Manuscripts Volume, p. 211–219.
- Deen, J.A., Rye, R.O., Munoz, J.L., and Drexler, J.W., 1994, The magmatic hydrothermal system at Julcani, Peru: Evidence from fluid inclusions and hydrogen and oxygen isotopes: Economic Geology, v. 89, p. 1924–1938.
- Delmelle, P., and Bernard, A., 1994, Geochemistry, mineralogy, and chemical modelling of the acid crater lake of Kawah Ijen volcano, Indonesia: Geochimica et Cosmochimica Acta, v. 58, p. 2445–2460.
- de Ronde, C.E.J., and Blatner, P., 1988, Hydrothermal alteration, stable isotopes, and fluid inclusions of the Golden Cross epithermal gold-silver deposit, Waihi, New Zealand: Economic Geology, v. 83, p. 895–917.
- Dubič, B., Dunning, G., and Lauzière, K., 1998, Geology of the Hope Brook mine, Newfoundland, Canada: A preserved Late Proterozoic high-sulfidation epithermal gold deposits and its implications for exploration: Economic Geology, v. 93, p. 405–436.
- Ebert, S.W., and Rye, R.O., 1997, Secondary precious metal enrichment by steam-heated fluids in the Crofoot-Lewis hot spring gold-silver deposit and relation to paleoclimate: Economic Geology, v. 92, p. 578–600.
- Fedorov, Ye.S., 1903, Host rocks of the Kedabek (Armenia): Zapiski Akademii Nauk, series 8, v. 15.
- Fontboté, L., and Bendezi, R., 1999, The carbonate-hosted Zn-Pb San Gregorio deposit, Colquijirca district, central Peru, as part of a high-sulfidation epithermal system. *in* Stanley, C.J., et al., eds., Mineral deposits: Processes to processing: Rotterdam, Balkema, p. 495–498.
- Giggenbach, W.F., 1981, Geothermal mineral equilibria: Geochimica et Cosmochimica Acta, v. 45, p. 393–410.
- 1992a, Magma degassing and mineral deposition in hydrothermal systems along convergent plate boundaries: Economic Geology, v. 87, p. 1927–1944.
- 1992b, Isotopic shifts in waters from geothermal and volcanic systems along convergent plate boundaries and their origin: Earth and Planetary Sciences, v. 113, p. 495–510.
- 1997, The origin and evolution of fluids in magmatic-hydrothermal systems, *in* Barnes, H.L., ed., Geochemistry of hydrothermal ore deposits, 3rd ed.: New York, Wiley, p. 737–796.
- Giggenbach, W.F., and Stewart, M.K., 1982, Processes controlling the isotopic composition of steam and water discharges from steam vents and steam-heated pools in geothermal areas: Geothermics, v. 11, p. 71–80.
- Giles, D.L., and Nelson, C.F., 1982, Epithermal lode gold deposits of the circum-Pacific: Circum Pacific Energy and Mineral Resources Conference, 3rd, Honolulu, August 1982, Transactions, p. 273–278.
- Groves, D.L., Goldfarb, R.J., Gebre-Mariam, M., Hagemann, S.G., and Robert, F., 1998, Orogenic gold deposits: A proposed classification in the context of their crustal distribution and relationship to other gold deposit types: Ore Geology Reviews, v. 13, p. 7–27.
- Hallberg, A., 1994, The Enäsen gold deposit, central Sweden: 1. A paleoproterozoic high-sulfidation epithermal gold mineralization: Mineralium Deposita, v. 29, p. 150–162.
- Harvey, B.A., Myers, S.A., and Klein, T., 1999, Yanacocha gold district, northern Peru: PacRim '99, Bali, Indonesia, 10–13 October, Proceedings, p. 445–459.
- Heald, P., Foley, N.K., and Hayba, D.O., 1987, Comparative anatomy of volcanic-hosted epithermal deposits: Acid-sulfate and adularia-sericite types: Economic Geology, v. 82, p. 1–26.
- Hedenquist, J.W., 1987, Mineralization associated with volcanic-related hydrothermal systems in the Circum-Pacific Basin: Circum Pacific Energy and Mineral Resources Conference, 4th, Singapore, August 1986, Transactions, p. 513–524.
- 1990, The thermal and geochemical structure of the Broadlands-Ohaaki geothermal system: Geothermics, v. 19, p. 151–185.
- 1991, Boiling and dilution in the shallow portion of the Waiotapu geothermal system, New Zealand: Geochimica et Cosmochimica Acta, v. 55, p. 2753–2765.
- 1995, The ascent of magmatic fluid: Discharge versus mineralization: Mineralogical Association of Canada Short Course Handbook, v. 23, p. 263–289.
- Hedenquist, J.W., and Henley, R.W., 1985a, The importance of CO₂ on freezing point measurements of fluid inclusions: Evidence from active geothermal systems and implications for epithermal ore deposition: Economic Geology, v. 80, p. 1379–1406.
- 1985b, Hydrothermal eruptions in the Waiotapu geothermal system, New Zealand: Their origin, associated breccias, and relation to precious metal mineralization: Economic Geology, v. 80, p. 1640–1668.
- Hedenquist, J.W., and Lowenstern, J.B., 1994, The role of magmas in the formation of hydrothermal ore deposits: Nature, v. 370, p. 519–527.
- Hedenquist, J.W., Matsuhisa, Y., Izawa, E., White, N.C., Giggenbach, W.F., and Aoki, M., 1994, Geology and geochemistry of high-sulfidation Cu-Au mineralization in the Nansatsu district, Japan: Economic Geology, v. 89, p. 1–30.
- Hedenquist, J.W., Izawa, E., Arribas, A., Jr., and White, N.C., 1996, Epithermal gold deposits: Styles, characteristics, and exploration: Poster and booklet, Resource Geology Special Publication 1, 17 p. (with translations to Spanish, French, Japanese, and Chinese).
- Hedenquist, J.W., Arribas, A., Jr., and Reynolds, T.J., 1998, Evolution of an intrusion-centered hydrothermal system: Far Southeast-Lepanto porphyry-epithermal Cu-Au deposits, Philippines: Economic Geology, v. 93, p. 373–404.
- Heinrich, C.A., Günther, D., Audéat, A., Ulrich, T., and Frischknecht, R., 1999, Metal fractionation between magmatic brine and vapor, determined by microanalysis of fluid inclusions: Geology, v. 27, p. 755–758.
- Hemley, J.J., and Hunt, J.P., 1992, Hydrothermal ore-forming processes in the light of studies in rock-buffered systems: II. Some general geological applications: Economic Geology, v. 87, p. 23–43.
- Hemley, J.J., Hostetler, P.B., Gude, A.J., and Mountjoy, W.T., 1969, Some stability relations of alunite: Economic Geology, v. 64, p. 599–612.
- Hemley, J.J., Montoya, J.W., Marinenko, J.W., and Luce, R.W., 1980, Equilibria in the system Al₂O₃-SiO₂-H₂O and some general implications for alteration/mineralization processes: Economic Geology, v. 75, p. 210–228.
- Henley, R.W., 1985, The geothermal framework of epithermal deposits: Society of Economic Geologists, Reviews in Economic Geology, v. 2, p. 1–24.
- 1990, Ore transport and deposition in epithermal ore environments, *in* Herbert, H.K., and Ho, S.E., eds., Stable isotopes and fluid processes in mineralization: Perth, University of Western Australia, Geology Department Publication 23, p. 51–60.
- 1991, Epithermal gold deposits in volcanic terranes, *in* Foster, R.P., ed., Gold metallogeny and exploration: London, Blackie, p. 133–164.
- Henley, R.W., and Ellis, A.J., 1983, Geothermal systems, ancient and modern: Earth Science Reviews, v. 19, p. 1–50.
- Henley, R.W., Truesdell, A.H., and Barton, P.B., Jr., 1984, Fluid-mineral equilibria in hydrothermal systems: Society of Economic Geologists, Reviews in Economic Geology, v. 1, 267 p.
- Itaya, T., Arribas, A., Jr., and Okada, T., 1996, Argon release systematics of hypogene and supergene alunite based on progressive heating experiments from 100° to 1000°C: Geochimica et Cosmochimica Acta, v. 60, p. 4525–4535.
- Izawa, E., and Yamashita, M., 1995, Truscottite from the Hishikari mine, Kagoshima prefecture [abs.]: Resource Geology, v. 45, no. 252, p. 251.
- 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: Journal of Geochemical Exploration, v. 36, p. 1–56.
- Jannas, R.R., Beane, R.E., Abler, B.A., and Brosnahan, D.R., 1990, Gold and copper mineralization at the El Indio deposit, Chile: Journal of Geochemical Exploration, v. 36, p. 233–266.
- Jannas, R.R., Bowers, T.S., Petersen, U., and Beane, R.E., 1999, High-sulfidation deposit types in the El Indio district, Chile: Society of Economic Geologists Special Publication 7, p. 219–266.
- Jensen, F.P., and Barton, M.D., 2000, Gold deposits related to alkaline magmatism: Society of Economic Geologists, Reviews in Economic Geology, v. 13, p. 279–314.
- John, D.A., 1999, Magmatic influence on characteristics of Miocene low-sulfidation Au-Ag deposits in the northern Great Basin [abs.]: Geological Society of America, Annual Meeting, Denver, Colorado, October 24–28, 1999, Abstracts with Programs, v. 30, p. 405.

- John, D.A., Garside, L.J., and Wallace, A.R., 1999. Magmatic and tectonic setting of late Cenozoic epithermal gold-silver deposits in northern Nevada, with an emphasis on the Pah Rah and Virginia ranges and the northern Nevada rift: Geological Society of Nevada, Special Publication no. 29, p. 65–158.
- Lindgren, W., 1922. A suggestion for the terminology of certain mineral deposits: *Economic Geology*, v. 17, p. 292–294.
- 1933. *Mineral deposits*, 4th edition: New York, McGraw-Hill, 930 p.
- McKibben, M.A., and Hardie, L.A., 1997. Ore-forming brines in active continental rifts, in Barnes, H.L., ed., *Geochemistry of hydrothermal ore deposits*, 3rd ed.: New York, Wiley, p. 877–935.
- Mancano, D.P., and Campbell, A.R., 1995. Microthermometry of enargite-hosted fluid inclusions from the Lepanto, Philippines, high-sulfidation Cu-Au deposit: *Geochimica et Cosmochimica Acta*, v. 59, p. 3909–3916.
- Matsuhisa, Y., Morishita, Y., and Sato, T., 1985. Oxygen and carbon isotope variations in gold-bearing hydrothermal veins in the Kushikino mining area, southern Kyushu, Japan: *Economic Geology*, v. 80, p. 283–293.
- Meyer, C., and Hemley, J.J., 1967. Wall rock alteration, in Barnes, H.L., ed., *Geochemistry of hydrothermal ore deposits*: New York, Holt, Rinehart and Winston, p. 166–235.
- Moore, J.N., Adams, M.C., and Anderson, A.J., 2000. The fluid-inclusion and mineralogic record of the transition from liquid- to vapor-dominated conditions in The Geysers geothermal system, California: *Economic Geology*, v. 95 (in press).
- Munteau, J., and Einaudi, M., 2000. Porphyry gold deposits of the Refugio district, Maricunga belt, northern Chile: *Economic Geology*, v. 95, (in press).
- Nakovnik, N.L., 1933. New data about the so-called secondary quartzites: *Problemy sovetskoy geologii*, N6, p. 228–242.
- 1934. Kounrad and its secondary quartzites: *Problemy sovetskoy geologii*, N4, p. 29–42.
- 1968. Secondary quartzites of the USSR and associated mineral deposits, 2nd ed.: Moscow, Nedra, 335 p. (in Russian)
- Nash, J.T., and Trudel, W.S., 1996. Bulk mineable gold ore at the Sleeper Mine, Nevada—Importance of extensional faults, breccia, fiamoids, and oxidation: Geological Society of Nevada Symposium, Reno/Sparks, Nevada, April 1995, Proceedings, p. 235–256.
- Neville, R.A., and Powell, I., 1992. Design of SFSI: An image spectrometer in the SWIR: *Canadian Journal of Remote Sensing*, v. 18, p. 210–222.
- Neville, R.A., Rowlands, N., Marois, R., and Powell, I., 1995. SFSI: Canada's first airborne SWIR imaging spectrometer: *Canadian Journal of Remote Sensing*, v. 21, p. 328–336.
- Noble, D.C., Park-Li, B., Henderson, W.B., and Vidal, C.E., 1997. Hypogene oxidation and late deposition of precious metals in the Pierina high-sulfidation deposit and other volcanic and sedimentary rock-hosted gold systems: *Sociedad Geológica del Perú Congreso Peruano de Geología Publicación Especial 1*, 9th, Lima, August 1997, Extended Abstracts, p. 121–127.
- Nolan, T.B., 1933. Epithermal precious-metal deposits of the western states (Lindgren volume): New York, American Institute of Mining and Metallurgical Engineers, p. 623–640.
- Northern Miner, 2000. Great Basin expands Ivanhoe: *The Northern Miner*, v. 86, no. 3, 13 March, 2000.
- O'Neil, J.R., and Silberman, M.L., 1974. Stable isotope relations in epithermal Au-Ag deposits: *Economic Geology*, v. 69, p. 902–909.
- Oviedo, L., Fuster, N., Tschichow, N., Ribba, L., Zuccone, A., Grez, E., and Aguilar, A., 1991. General geology of La Coipa precious metal deposit, Atacama, Chile: *Economic Geology*, v. 86, p. 1287–1300.
- Pease, R.B., 1999. The Cerro Crucitas deposit, Costa Rica: Geological Society of America Cordilleran Section Meeting, Berkeley, California, June 1–3, 1999, Abstracts with Programs, p. A-85.
- Ransome, F.L., 1907. The association of alunite with gold in the Goldfield district, Nevada: *Economic Geology*, v. 2, p. 667–692.
- 1909. The geology and ore deposits of Goldfield, Nevada: U.S. Geological Survey, Professional Paper 66, 258 p.
- Reyes, A.G., 1990. Petrology of Philippine geothermal systems and the application of alteration mineralogy to their assessment: *Journal of Volcanology and Geothermal Research*, v. 43, p. 279–309.
- Reyes, A.G., Giggenbach, W.F., Saleras, J.R.M., Salonga, N.S., and Vergara, M.C., 1993. Petrology and geochemistry of Alto Peak, a vapor-cored hydrothermal system: *Geothermics*, v. 22, p. 479–519.
- Richards, J.P., 1995. Alkaline-type epithermal gold deposits: *Mineralogical Association of Canada Short Course Handbook*, v. 23, p. 367–400.
- Roedder, E., 1984. Fluid inclusions: *Reviews in Mineralogy*, v. 12, 644 p.
- Rowins, S.M., 2000. Reduced porphyry copper-gold deposits: A new variation on an old theme: *Geology*, v. 28, p. 491–494.
- Rusakov, M.P., 1926. Secondary quartzites and porphyry copper of the Kirgizskaya Step: *Vestnik Geolcoma*, N3, p. 27–28.
- Russell, N., Scaward, M., Rivera, J., McCurdy, K., Kesler, S.E., and Cloke, P.L., 1981. Geology and geochemistry of the Pueblo Viejo gold-silver oxide ore deposit, Dominican Republic: *Transactions of the Institution of Mining and Metallurgy*, v. 90, p. B153–B162.
- Rye, R.O., Bethke, P.M., and Wasserman, M.D., 1992. The stable isotope geochemistry of acid-sulfate alteration: *Economic Geology*, v. 87, p. 225–267.
- Saunders, J.A., 1994. Silica and gold textures in bonanza ores of the Sleeper deposit, Humboldt County, Nevada: Evidence for colloids and implications for epithermal ore-forming processes: *Economic Geology*, v. 89, p. 628–638.
- Schoen, R., White, D.E., and Hemley, J.J., 1974. Argillization by descending acid at Steamboat Springs, Nevada: *Clays and Clay Minerals*, v. 22, p. 1–22.
- Sherlock, R.L., Tosdal, R.M., Lehrman, N.J., Graney, J.R., Losh, S., Jowett, E.C., and Kesler, S.E., 1995. Origin of the McLaughlin mine sheeted vein complex: Metal zoning, fluid inclusion, and isotopic evidence: *Economic Geology*, v. 90, p. 2156–2181.
- Siddeley, G., and Areneda, R., 1986. The El Indio-Fambo gold deposits, Chile: *Gold '86*, Konsult International, Toronto, 1986, Proceedings, p. 445–456.
- Sillitoe, R.H., 1975. Lead-silver, manganese, and native sulfur mineralization within a stratovolcano, El Queva, northwest Argentina: *Economic Geology*, v. 70, p. 1190–1201.
- 1977. Metallic mineralization affiliated to subaerial volcanism: A review: *Geological Society of London Special Publication 7*, p. 99–116.
- 1983. Enargite-bearing massive sulfide deposits high in porphyry copper systems: *Economic Geology*, v. 78, p. 348–352.
- 1993a. Epithermal models: Genetic types, geometrical controls, and shallow features: *Geological Association of Canada Special Paper 40*, p. 403–417.
- 1993b. Giant and bonanza gold deposits in the epithermal environment: *Society of Economic Geologists Special Publication 2*, p. 125–156.
- 1994. Erosion and collapse of volcanoes: Causes of telescoping in intrusion-centered ore deposits: *Geology*, v. 22, p. 945–948.
- 1995a. Exploration of porphyry copper lithocaps: *Australasian Institute of Mining and Metallurgy Publication Series No. 9/95*, p. 527–532.
- 1995b. Exploration and discovery of base- and precious-metal deposits in the circum-Pacific region during the last 25 years: *Resource Geology Special Issue 19*, 119 p.
- 1997. Characteristics and controls of the largest porphyry copper-gold and epithermal gold deposits in the circum-Pacific region: *Australian Journal of Earth Sciences*, v. 44, p. 373–388.
- 1999. Styles of high-sulphidation gold, silver, and copper mineralization in the porphyry and epithermal environments: *PacRim '99*, Bali, Indonesia, 10–13 October, Proceedings, p. 29–44.
- Sillitoe, R.H., and Bonham, H.F., 1990. Sediment-hosted gold deposits—Distal products of magmatic-hydrothermal systems: *Geology*, v. 18, p. 157–161.
- Sillitoe, R.H., and McKee, E.H., 1996. Age of supergene oxidation and enrichment in the Chilean porphyry copper province: *Economic Geology*, v. 91, p. 164–179.
- Sillitoe, R.H., Steele, G.B., Thompson, J.F.H., and Lang, J.R., 1998. Advanced argillic lithocaps in the Bolivian tin-silver belt: *Mineralium Deposita*, v. 33, p. 539–546.
- Simmons, S.F., 1991. Hydrologic implications of alteration and fluid inclusion studies in the Fresnillo district, Mexico: Evidence for a brine reservoir and a descending water table during the formation of hydrothermal Ag-Pb-Zn ore bodies: *Economic Geology*, v. 86, p. 1579–1601.
- 1995. Magmatic contributions to low-sulfidation epithermal deposits: *Mineralogical Association of Canada Short Course Handbook*, v. 23, p. 455–477.
- 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: *Geology, geochemistry, origin, and exploration: Journal of Geochemical Exploration*, v. 35, p. 63–104.

- 2000, Hydrothermal minerals and precious metals in the Broadlands-Ohaaki geothermal system: Implications for understanding low-sulfidation epithermal environments: *Economic Geology*, v. 95, p. 971–999.
- Simmons, S.F., and Christenson, B.W., 1994, Origins of calcite in a boiling geothermal system: *American Journal of Science*, v. 294, p. 361–400.
- Simmons, S.F., Archart, G., Simpson, M.P., and Mauk, J.L., 2000, Origin of massive calcite veins in the Golden Cross low-sulfidation epithermal Au-Ag deposit, New Zealand: *Economic Geology*, v. 95, p. 99–112.
- Staeenz, K., Neville, R.A., Lévesque, J., Szeredi, T., Singhroy, V., Borstad, G.A., and Hauff, P., 1999, Evaluation of CASI and SFSI hyperspectral data for environmental and geological applications—Two case studies: *Canadian Journal of Remote Sensing*, v. 25, p. 311–322.
- Steven, T.A., and Ratté, J.C., 1960, Geology of ore deposits of the Summitville district, San Juan Mountains, Colorado: U.S. Geological Survey Professional Paper 343, 70 p.
- Stoffregen, R.E., 1987, Genesis of acid-sulfate alteration and Au-Cu-Ag mineralization at Summitville, Colorado: *Economic Geology*, v. 82, p. 1575–1591.
- Thompson, A.J.B., Hauff, P.L., and Robitaille, A.J., 1999, Alteration mapping in exploration: Application of short-wave infrared (SWIR) spectroscopy: *Society of Economic Geologists Newsletter*, no. 39, p. 1, 16–27.
- Turner, S.J., 1997, The Yanacocha epithermal gold deposits, northern Peru: High-sulfidation mineralization in a flow dome setting: Unpublished Ph.D. thesis, Golden, Colorado School of Mines, 326 p.
- Volkert, D.F., McEwan, C.J.A., and Garay, M.E., 1998, Pierina Au-Ag deposit, Cordillera Negra, north-central Peru: *Pathways '98*, Vancouver, January 27–30, 1998, Extended Abstracts, p. 33–35.
- White, D.E., 1955, Thermal springs and epithermal ore deposits: *Economic Geology 50th Anniversary Volume*, p. 99–154.
- White, N.C., 1991, High sulfidation epithermal gold deposits: Characteristics, and a model for their origin: *Geological Survey of Japan Report 277*, p. 9–20.
- White, N.C., and Hedenquist, J.W., 1990, Epithermal environments and styles of mineralization: Variations and their causes, and guidelines for their exploration: *Journal of Geochemical Exploration*, v. 36, p. 443–474.
- 1995, Epithermal gold deposits: Styles, characteristics, and exploration: *Society of Economic Geologists Newsletter*, no. 23, p. 1, 9–13.
- White, N.C., Wood, D.G., and Lee, M.C., 1989, Epithermal sinters of Paleozoic age in north Queensland, Australia: *Geology*, v. 17, p. 718–722.
- White, N.C., Leake, M.J., McCaughey, S.N., and Parris, B.W., 1995, Epithermal deposits of the southwest Pacific: *Journal of Geochemical Exploration*, v. 54, p. 87–136.

Questions

1. What are the various but synonymous terms used to describe epithermal deposits that are hosted by silicified altered rock, with an intimate association with hypogene advanced argillic assemblages? Hosted by quartz-adularia veins with a halo of clay minerals? Disseminated in silicified rock and accompanied by adularia or illite? Lying beneath a silica sinter?
2. What are the typical silicate and sulfide mineral assemblages for low-sulfidation deposits in general? For end-member low-sulfidation deposits, and intermediate sulfidation deposits? For high-sulfidation deposits?
3. What are some indicators that paleosurface is nearly preserved in an epithermal prospect, for low-sulfidation systems; for high-sulfidation systems?
4. What are the distinctions in the nature of advanced argillic alteration, and silicic alteration, to be expected between low-sulfidation and high-sulfidation prospects that have suffered erosion to about 200 m below paleosurface?
5. What are some of the distinctions between alunite of hypogene and steam-heated origin; and supergene origin?
6. Name three types of deposit that can form in a spatial and/or genetic association with high-sulfidation deposits. All three deposit types are potential exploration targets in high-sulfidation environments.
7. Identify at least three types of ore mineralization in epithermal deposits and provide examples. Add some detail with respect to geological controls, metallurgy, mining, cut-off grade, etc.

Answers

(Note: columns and rows of tables referred to include the top row and left-most column)

1. See Table 1, column 2. For all the subsequent questions, see Table 1, column 1.
2. See Table 3, columns 2 and 3. Table 3, column 1 versus column 2. Table 3, columns 4–6.
3. See Table 4, rows 2–4, and Table 5a, column 2. Table 4, rows 3–4, and Table 5b, column 2.
4. Compare Tables 5a and 5b, column 4, row 3; and row 4.
5. Compare Table 6, column 2 versus 3, rows 6–8; column 4, rows 6–8.
6. Porphyry, intermediate sulfidation and skarn or carbonate replacement deposits.
7. For example: (1) lithologically controlled and disseminated, non-refractory, open-pittable and low strip ratio high-sulfidation deposit (e.g., Yanacocha, Pierina). The cut-off may be as low as ~0.2 g/t Au. (2) Structurally controlled, underground, bonanza low-sulfidation or high-sulfidation vein deposit (e.g., El Indio-high-sulfidation; Hishikari, Midas-low-sulfidation). (3) Structurally controlled, underground or open pittable breccia and/or replacement deposit with high As, refractory ore. For example, Chelopech or Pueblo Viejo. Here the cut-off grade may be 4 g/t Au or greater, more than an order of magnitude higher than the low-cost deposits discussed above.