

Linkages between volcanotectonic settings, ore-fluid compositions, and epithermal precious-metal deposits

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Abstract

Epithermal Au and Ag deposits of both vein and bulk-tonnage styles may be broadly grouped into high-sulfidation (HS), intermediate-sulfidation (IS), and low-sulfidation (LS) types based on the sulfidation states of their hypogene sulfide assemblages. The HS and LS types may be subdivided using additional parameters, particularly related igneous rock types and metal content.

Most HS deposits are generated in calc-alkaline andesitic-dacitic arcs characterized by near-neutral stress states or mild extension, although a few major deposits also occur in compressive arcs characterized by the suppression of volcanic activity. Rhyolitic rocks generally lack appreciable HS deposits. Highly acidic fluids produced the advanced argillic lithocaps that presage HS mineralization, which itself is due to higher-pH, moderate- to low-salinity fluids. Similar lithocaps in the Bolivian Sn-Ag belt, some mineralized with Ag and Sn, accompany reduced, ilmenite-series magmatism.

IS epithermal deposits occur in a broadly similar spectrum of andesitic-dacitic arcs, but commonly do not show such a close connection with porphyry Cu deposits as do many of the HS deposits. However, igneous rocks as silicic as rhyolite are related to a few IS deposits. IS deposits form from fluids spanning broadly the same salinity range as those responsible for the HS type, although Au-Ag, Ag-Au, and base-metal-rich Ag-(Au) subtypes reveal progressively higher ore-fluid salinities.

Most LS deposits, including nearly 60 percent of the world's bonanza veins, are associated with bimodal (basalt-rhyolite) volcanic suites in a broad spectrum of extensional tectonic settings, including intra-, near-, and backarc, as well as postcollisional, rifts. Some LS deposits, however, accompany extension-related alkaline magmatism, which unlike the bimodal suites, is capable of generating porphyry Cu deposits. Extensional arcs characterized by active andesitic-dacitic

volcanism do, however, host a few LS deposits. LS deposits genetically linked to bimodal volcanism are formed from extremely dilute fluids, whereas modestly saline contributions account for the LS deposits in alkaline centers.

Early lithocap-forming and HS fluids, as well as LS fluids in deposits associated with alkaline igneous rocks, display clear evidence for a close genetic relationship to magmatism and, although the linkage is less intimate, late HS fluids responsible for much of the Au introduction along with similar IS fluids also seem to owe much to their magmatic parentage. Where ascending IS fluids enter lithocaps, they evolve to HS fluids. Eventual neutralization and lowering of sulfidation state by wallrock interaction can convert HS back to IS fluids, as confirmed by both spatial and paragenetic transitions from HS to IS mineralization. In contrast, LS fluids other than those of alkaline affiliation lack such clear-cut connections to magmatism, although Giggerbach's work on the geothermal fluids associated with the Taupo Volcanic Zone in New Zealand suggests that a deep magmatic source different to that of fluids in andesitic arc terranes is probable. In addition, at least in some regions, there appears to be a correlation between the reduced sulfide assemblages of LS deposits and the reduced nature of the volcanic rocks with which they are associated. Therefore, it may be argued that the defining characteristics of epithermal deposits are related directly to their magmatic roots, notwithstanding the existence of important unanswered questions, especially regarding the source of LS fluids.

This review puts forward several exploration guidelines for epithermal precious-metal deposits. Exploration activity in andesitic-dacitic arcs should be restricted to HS and potentially related IS deposits containing Au and/or Ag, whereas a variety of rift-related bimodal suites and convergent-margin alkaline rocks offer the prime environments for Ag-deficient, LS Au deposits ($Ag/Au < \sim 15$). Bonanza Au veins are more likely to be of LS type and to be discovered at relatively shallow paleodepths in bimodal rift settings, where rhyolitic and/or basaltic rocks may be proximal to Au ore. Even tholeiitic basalts in emergent mid-ocean ridge or hotspot settings might possess underappreciated epithermal Au potential. Subaerial extensions to some VMS belts may possess LS Au potential because of the broadly similar volcanotectonic settings for both deposit types. The reduced, ilmenite-series volcanic rocks of the Bolivian Sn-Ag belt are unfavorable for epithermal Au. Deficiency of volcanic rocks in epithermal provinces is typical of highly compressive arcs (HS-IS deposits) and some rifts swamped by fluviolacustrine sedimentation with silica sinter occurrences (LS deposits). In contrast to HS and IS deposits, exploration for LS Au deposits, even where exposed, may be hampered by the visually subtle nature of many outcropping examples.

Introduction

Early metallogenic analysis using plate-tectonic theory assigned epithermal precious-metal deposits to subduction-related arc terranes (Sawkins, 1972). In addition, the association of a few epithermal deposits with post-subduction extensional regimes characterized locally by bimodal volcanism was also noted (Sillitoe, 1977). The subduction-related context of many epithermal deposits has become generally accepted, although a variety of arc and extra-arc settings characterized by distinctive volcanic rock suites and associated epithermal deposit types are now widely recognized. For example, Sillitoe (1989) subdivided low-sulfidation epithermal Au deposits of the western Pacific region into a sulfide- and base metal-rich type in andesitic-dacitic arcs and a sulfide- and base metal-poor type in rhyolite-bearing extensional arcs. Shortly thereafter, a similar distinction was made for epithermal deposits in the western United States (Albino and Margolis, 1991). Most recently, John (2001) noted the volcanotectonic distinction between sulfide-rich epithermal deposits of the western andesite arc of Nevada and the sulfide-poor, high-grade Au deposits of the northern Nevada rift, hosted by bimodal volcanic rocks. He also highlighted the differences in oxidation state of the associated magmas, and the apparent reflection of the magmatic oxidation state in the ore assemblages.

This report summarizes our present understanding of the volcanotectonic settings of epithermal deposits and some possible consequences. It commences with a brief overview of epithermal deposit types and their salient geologic features, and attempts to correlate the currently accepted classification scheme with those employed in the early decades of the 20th century. Possible linkages between volcanotectonic settings, compositions of hydrothermal fluids, and epithermal deposit types are then explored. Given that the mineralogic differences between epithermal deposits result in large measure from chemical differences between the fluids that deposited them (e.g., Barton et al., 1977; Giggenbach, 1997; Einaudi et al., this volume), speculations are offered on the evolution and ultimate sources of epithermal fluids. In this regard, the variation in volatile content of magma types from different tectonic settings is highlighted. The resulting volcanotectonic classification of epithermal deposits may be used in exploration to guide area selection, especially if the targets are high-grade (bonanza) Au deposits.

Epithermal Deposit Types

Early classification schemes for volcanic-hosted epithermal deposits, almost all of vein type, were based on their dominant elements and minerals (Emmons, 1918; Lindgren, 1933). Even earlier, however, several distinctive epithermal vein varieties had been noted (Lindgren, 1901), with documentation of type examples of Au-alunite (Goldfield, Nevada; Ransome, 1907) and Au-telluride (Cripple Creek, Colorado; Lindgren and Ransome, 1906) deposits being especially important.

During the last 25 years or so, two principal epithermal deposit types (Sillitoe, 1977; Hayba et al., 1985; Bonham, 1986; Heald et al., 1987; Hedenquist, 1987) have been widely recognized in both vein and bulk-tonnage style deposits and referred to by

several synonyms, and with subtypes noted (Table 1). More recent study of the variations in deposit style, from tectonic setting to mineralogy, has led to three environments being identified (John et al., 1999; John, 2001; Table 1). Herein, we use the most widely employed nomenclature: high-sulfidation (HS), intermediate-sulfidation (IS), and low-sulfidation (LS), terms introduced by Hedenquist (1987) and Hedenquist et al. (2000). HS deposits contain sulfide-rich assemblages of high sulfidation state, typically pyrite-enargite, pyrite-luzonite, pyrite-famatinite, and pyrite-covellite (Einaudi et al., this volume; Fig. 1), hosted by leached silicic rock with a halo of advanced argillic minerals. In contrast, LS deposits contain the low-sulfidation pair, pyrite-arsenopyrite, the latter sulfide mineral typically present in only relatively minor quantities, within banded veins of quartz, chalcedony, and adularia plus subordinate calcite. Very minor amounts of Cu (typically <100-200 ppm) are largely present as chalcopyrite or, less commonly, tetrahedrite-tennantite (Einaudi et al., this volume; Fig. 1). Pyrrhotite is present in trace amounts in only some LS deposits (e.g., northern Nevada rift; John and Wallace, 2000; John, 2001). As the name implies, IS deposits possess sulfidation states between those of the HS and LS types, typically with stability of chalcopyrite, tetrahedrite-tennantite, and FeS-poor sphalerite, but lacking appreciable arsenopyrite and pyrrhotite (Einaudi et al., this volume; Fig. 1). Like the early classification schemes for epithermal deposits, the current one is also based on mineralogic criteria, albeit the hypogene sulfide assemblages interpreted in terms of sulfidation state (Barton and Skinner, 1967; Barton, 1970). Hence, it is fairly straightforward to reinterpret the early classifications in the context of the current scheme (Table 2).

On the basis of these apparent correlations between early and current schemes, and taking into account more recent proposals for an alkaline epithermal type (Bonham, 1986, 1988) and Ag-Sn-rich (as opposed to Au-Cu-rich) deposits hosted by lithocaps (Sillitoe et al., 1998), a series of widely recognizable epithermal types and subtypes are proposed. Their defining characteristics and type examples are summarized in Table 3. Assignment of these Bolivian epithermal Ag-Sn deposits to the HS category is somewhat contentious, as they lack high-sulfidation sulfides such as enargite, even though it was recently appreciated that they are hosted by widespread advanced argillic alteration (Sillitoe et al., 1998), similar to their Au-Cu-rich cousins. Indeed, White and Poizat (1995) considered them to be of LS type, like the subjacent mesothermal Sn-base metal veins. Despite the abundance of hypogene alunite and aluminum-phosphate-sulfate (APS) minerals in the host alteration zones, the extremely limited knowledge of the sulfide mineralogy, resulting from supergene oxidation, precludes definitive assignment of sulfidation state. The carbonate-base metal Au deposits distinguished by Leach and Corbett (1994) and Corbett and Leach (1998, Chapter 7) in the southwestern Pacific region, although not confined to the epithermal environment, may be considered as deep, quartz-poor representatives of the Au-rich IS subtype.

There is evidence for a spatial, and in some cases genetic relationship between HS and IS deposits (Sillitoe, 1999a; Hedenquist et al., 2000), and thus these two deposit types may be considered as being potentially transitional (Einaudi et al., this volume). In addition, the presence of sulfide minerals that are indicative of widely varying sulfidation state is common (Buchanan, 1981; White et al., 1995), albeit with

information on paragenesis being rare. Such occurrence of apparently incompatible sulfide minerals (e.g., rare occurrences of arsenopyrite and pyrrhotite in IS and HS deposits or enargite in IS and LS deposits) may indicate appreciable but local fluctuations in sulfidation state during the lives of many hydrothermal systems (Giggenbach, 1992; Einaudi et al., this volume) or, alternatively, that equilibrium was simply not attained. Except for these transient or nonequilibrium events, which can cause some confusion in classification, the potential for a true transition between IS and LS deposit types is considered to be low, given their largely distinct tectonic settings, as outlined below.

Volcanotectonic Settings

Review of worldwide examples of major epithermal Au and Ag deposits suggests a reasonable correspondence between the various epithermal types and subtypes and specific volcanotectonic settings (Sillitoe, 1989, 2002a; Berger and Bonham, 1990; John, 2001), although, as noted, there are a number of relatively minor exceptions to the general scheme. It must be emphasized, however, that there is rarely, if ever, proof of a genetic connection between an epithermal deposit and the spatially and temporally related volcanism and tectonism. Therefore, the genetic relationship generally has to be assumed, a process which may lead sometimes to incorrect assignment of epithermal deposits to particular volcanotectonic settings. Extensional and compressive tectonic settings are defined on the basis of regional structural features (e.g., normal faulting during extension and reverse faulting during compression). Oxidized and reduced volcanic rocks are roughly defined on the basis of their contained opaque oxide minerals, i.e., magnetite- vs. ilmenite-series (Ishihara, 1981). The proposed correlations with volcanotectonic setting are summarized herein by epithermal type.

HS deposits

Typical Au-Cu-bearing HS epithermal deposits occur mainly in calc-alkaline andesitic-dacitic arc terranes (Fig. 2; Table 4), although locally volcanic rocks as felsic as quartz latite are genetically related to the mineralization (e.g., Summitville, Colorado; Steven and Ratté, 1960). Certainly, rhyolitic volcanic rocks normally lack any appreciable HS mineralization (Sillitoe, 1993a). Furthermore, it is evident that arcs subjected to neutral stress conditions or mild extension host many of the world's premier HS deposits. The giant low-grade Yanacocha Au deposit and its underlying porphyry Cu systems in northern Peru, which lie within the extensional Calipuy arc, record this relationship (Fig. 2a; Petford and Atherton, 1994). In addition, the high-grade Au deposit at Goldfield, Nevada, is part of an extensional volcanic terrane characterized by rocks ranging in composition from basalt through andesite and dacite to rhyolite (Seedorff, 1991; John, 2001). Extension, deduced from the graben control of volcanism, also seems to have prevailed during intermediate- to high-K calc-alkaline volcanism and formation of major HS deposits, like Bor and Chelopech, in the Timok-Srednogorie belt of Serbia and Bulgaria, respectively, as summarized by Ciobanu et al. (2002). The La Coipa HS Au-Ag deposit and accompanying dacitic domes in the Maricunga belt of northern Chile developed in the interval from 24 to 20 Ma (Sillitoe et al., 1991; Kay et al., 1994), prior to onset of an episode of compression and related crustal thickening from ~40 or 45 km up to >50 km between

20 and 18 Ma (Kay and Mpodozis, 2001, 2002).

Once extension in either arc, near-arc, or backarc terranes becomes more advanced, as indicated by the bimodal (basalt-rhyolite) composition of the host volcanic-rock suites (e.g., Martin and Piwinski, 1972), HS deposits and their precursor advanced argillic lithocaps are not normally generated. One of the few apparent exceptions is provided by minor HS Au occurrences in the Los Menucos district within the Permo-Triassic backarc of the Patagonian Andes, Argentina, where there is at least a spatial relationship with rhyolite domes (Franco et al., 1999). Alkaline volcanic and associated subvolcanic rocks, whether saturated or undersaturated in silica and in whatever tectonic setting, likewise seldom contain HS precious-metal deposits or barren advanced argillic lithocaps (Sillitoe, 2002b). Nevertheless, several occurrences are known; perhaps the best described is the monzonite-centered, shoshonite-hosted Navisi 3 prospect nearby the Emperor LS Au-telluride vein deposit in Fiji (Eaton and Setterfield, 1993).

Compressive arcs also host both small, high-grade and large, low-grade HS Au-Ag deposits. The prime example is the El Indio belt, northern Chile, where the giant Pascua-Lama and El Indio-Tambo deposits formed since ~9 Ma (Table 4; Fig. 2b; Bissig et al., 2002), as magmatism ceased in response to slab flattening, strong compressive deformation, and crustal thickening to ~60 km (Kay and Mpodozis, 2001, 2002). Compression inhibited eruptive activity (Takada, 1994), leading to the lack of appreciable volumes of coeval subvolcanic and volcanic rocks in association with these HS deposits, although a parent magma chamber is assumed to have existed at depth. The Luzon Central Cordillera, northern Philippines, is one of the few other compressive arcs where an areally extensive advanced argillic lithocap atop several porphyry copper-gold systems hosts a significant HS deposit, at Lepanto. The compression may be attributed to eastward subduction of an aseismic ridge in the South China Sea leading to slab flattening, rapid uplift, and cessation of volcanism (Yang et al., 1996).

Although only a few HS deposits appear to have formed during compression, most of the world's premier porphyry Cu provinces were generated during compressive (or transpressive) tectonic regimes (Sillitoe, 1998), albeit commonly during stress relaxation (Tosdal and Richards, 2001). However, these compressive provinces do not seem to host appreciable HS mineralization of the same ages as the porphyry Cu deposits despite the presence of extensive advanced argillic lithocaps. The Late Eocene-Early Oligocene porphyry Cu belt of northern Chile, for example, is characterized by numerous lithocaps, all of them barren. This fact is difficult to explain simply by assuming that erosion levels are too deep for preservation of epithermal mineralization and, hence, it requires a mechanism for inhibiting generation of an appropriate ore fluid or its ascent to the level of the lithocap where most HS ore forms.

The giant Pueblo Viejo HS Au deposit, Dominican Republic, is generally considered to be an integral part of the Early Cretaceous Los Ranchos Formation, irrespective of the preferred interpretation of the volcanic setting (Russell and Kesler, 1991; Nelson, 2000). If so, then a bimodal basalt-rhyolite suite of island-arc tholeiite rather than

calc-alkaline affinity, generated immediately following collision-related arc reversal (Lebron and Perfit, 1993), was responsible for the HS event. However, if the Late Cretaceous (~77 Ma) $^{40}\text{Ar}/^{39}\text{Ar}$ ages for ore-related alunite reported by Kesler (1998) are not, as generally assumed, spurious, the mineralization would be much younger and part of a well-defined calc-alkaline magmatic belt rather than being related to the host bimodal suite.

A backarc setting subjected to compressive deformation is the site of mineralized advanced argillic lithocaps in the Sn-Ag belt of the Cordillera Oriental of Bolivia (Fig. 2c). The compression is manifested by east-directed thrusting along the eastern side of the Cordillera Oriental (Lamb et al., 1997). The calc-alkaline volcanic and subvolcanic rocks related to these systems are predominantly peraluminous rhyodacite and markedly reduced (ilmenite series; Lehmann, 1994). In contrast to the HS deposits in arc terranes, these backarc variants are characterized by their Ag, Sn, and Sb contents and lack appreciable Cu, Au, and As (Sillitoe et al., 1998). Moreover, the largest of the advanced argillic lithocaps, host to the giant Potosí Ag deposit, is rich in APS minerals but deficient in hypogene alunite, unlike most lithocaps in arc environments (Sillitoe et al., 1998). This difference may be due to the anomalously high phosphorus contents of the volcanic host rocks rather than solely due to the predicted lower amounts of oxidized sulfur evolved from reduced magmas (Sillitoe et al., 1998).

IS deposits

Most IS epithermal Au and Ag deposits, like their HS brethren, occur in calc-alkaline andesitic-dacitic arcs, although more felsic rocks act as hosts locally (Fig. 2; Table 5). A few IS deposits, typically small ones, adjoin advanced argillic lithocaps, with or without associated HS deposits (Sillitoe, 1999a). Probably the largest of these is the Victoria deposit, located alongside the Lepanto HS deposit in northern Luzon (Claveria, 2001; Table 4). The rhodochrosite-bearing IS veins at Chiufen-Wutanshan, Taiwan, produced 29 t Au, and occur 1 km west of the larger HS deposit of Chinkuashih, within the same overall alteration system and of the same age (Tan, 1991). The Río del Medio deposit along strike of the El Indio HS vein system (Table 4) is another, albeit smaller, example (Jannas et al., 1999). The Acupan and Antamok deposits in the Baguio district, Philippines, are other large IS deposits that share a broad spatial and, possibly, temporal (Aoki et al., 1993) relationship to a weakly mineralized lithocap. In the case of Masupa Ria in Central Kalimantan, Indonesia, however, the IS vein cuts an extensive but largely barren lithocap (Thompson et al., 1994).

Some IS deposits show a spatial association with porphyry systems, including the Baguio district (Sillitoe, 1989) and several deposits in the South Apuseni Mountains of Romania (e.g., Valea Morii and Rosia Poeni; Ivascanu et al., 2002). In contrast, no such IS vein and porphyry relationship has been noted in the major epithermal Ag belt of Mexico, discussed below, although there are epithermal veins proximal to porphyry and skarn deposits elsewhere in Mexico (Albinson et al., 2001).

Many major IS deposits formed in arcs with neutral to mildly extensional stress states, a tectonic environment that includes not only several large vein Au-Ag

deposits, but also most of the world's major epithermal vein Ag deposits (Table 5). In the western USA, post-38 Ma extension of varied severity accompanied widespread calc-alkaline andesite-dacite-rhyolite volcanism of subduction origin, which culminated in bimodal igneous activity during the Miocene on termination of subduction (Seedorff, 1991). Major IS precious-metal deposits apparently formed during modest extension in the Great Basin of Nevada include Comstock Lode, in an andesitic-dacitic dome field (Bonham, 1969), and Tonopah, formed at the end of an andesitic-rhyodacitic volcanic event (Bonham and Garside, 1979; Table 5). At broadly the same time in the San Juan Mountains, Colorado, IS precious-metal deposits, also noted for their base-metal production, are widespread and include Creede and Camp Bird (Lipman et al., 1976; Table 5).

Northeast-southwest extension in the Sierra Madre Occidental and contiguous Altiplano central plateau of Mexico during the Late Oligocene and Miocene was a weaker southward continuation of the broadly contemporaneous extensional event in the Great Basin of the western USA (Staupe and Barton, 2001). The associated calc-alkaline arc magmatism gave rise to the world's premier epithermal Ag province based on Ag-dominant (Fresnillo), Ag-Au (Pachuca-Real del Monte and Guanajuato), and Au-Ag (Tayoltita) deposits of IS type (Albinson et al., 2001). A similarly broad andesite-dacite-rhyolite arc terrane was constructed in the Altiplano of southern Peru during the Miocene and again gave rise to several major IS Ag-Au deposits, including Arcata and Orcopampa (Noble et al., 1990).

Extensional basins linked to east-striking transcurrent fault systems generated during tectonic extrusion caused by Alpine collision farther west are the sites of andesitic volcanism in the Carpathian arc from 12 to 14 Ma (Ivascanu et al., 2002). Several major IS vein and breccia-hosted precious- and base-metal deposits, including Sacarimb and Rosia Montana in the South Apuseni Mountains of Romania, are associated with this magmatic activity (Alderton and Fallick, 2000). Neutral-stress to mildly extensional arc segments elsewhere are also characterized by IS deposits, including the Kelian Au deposit in Kalimantan, Indonesia (van Leeuwen et al., 1990), and the Toyoha Ag-Pb-Zn-In deposit in southwestern Hokkaido, Japan (Watanabe, 2002). Nevertheless, compressive arcs are not devoid of IS deposits, in the same way that they also contain some HS systems. For example, the Luzon Central Cordillera of the northern Philippines contains the Victoria deposit (see above) and the giant Baguio Au district. The Altiplano of Bolivia was also under compression (Muñoz and Charrier, 1996; Lamb et al., 1997) during andesite-dacite dome emplacement and development of the San Cristóbal deposit, disseminated IS Ag-Zn-Pb mineralization unusually rich in adularia alteration (Buchanan, 2000).

LS deposits

Only a few LS deposits, as defined herein, are formed in andesite-dacite-rhyolite arc terranes, and perhaps only when extension is fairly advanced and, arguably, about to herald onset of bimodal or mafic magmatism (e.g., Rytuba, 1996). Excellent examples are provided by the giant, bulk-tonnage Round Mountain Au deposit, Nevada, which formed during regional extension and rhyolitic caldera development (Axen et al., 1993; Henry et al., 1996); the high-grade Bodie Au deposit, California, where both LS and IS vein systems formed within 0.12 m.y. of one another during

graben formation and dacitic volcanism (Berger et al., 1999); and at Waihi, North Island, New Zealand, following an arc-wide change from andesitic to rhyolitic volcanism (Brathwaite and Faure, 2002; Fig. 2d; Table 6).

Many LS deposits are formed in a variety of continental and island-arc rifts characterized by bimodal volcanism (Sillitoe, 2002b). The rifts may occur in intra-, near-, and backarc settings during subduction of oceanic lithosphere, postarc settings following cessation of subduction, and postcollisional settings (Fig. 2). Depending on the environment concerned, the rifting is most typically attributed to one or more of the following tectonic situations: slab rollback consequent upon steepening or waning subduction, oblique subduction (that can lead to creation of slab windows), asthenospheric upwelling caused by removal of mantle lithosphere and/or attached underthrust slabs (slab breakoff), gravitationally controlled lateral expansion (tectonic collapse) of thickened crust following episodes of contractional deformation, and hotspot (possibly mantle-plume) activity (e.g., Sonder and Jones, 1999). The accompanying bimodal volcanism comprises basalt or basaltic andesite plus rhyolite or rhyodacite of calc-alkaline and/or tholeiitic affinities, with a notable deficiency of andesitic and dacitic rock compositions.

Bimodal volcanic suites and LS vein Au deposits characterize several extensional arcs. They include the Paleocene arc in the central Andes of northern Chile (Cornejo et al., 1994), where the El Peñón deposit is hosted by a rhyolite dome complex (Fig. 2d; Robbins, 2000), and the Middle to Late Jurassic arc in the Patagonian Andes of southern Argentina where the Esquel veins share a fault zone with broadly synmineralization dikes of basaltic andesite and rhyolite compositions (Sillitoe et al., 2002). Similarly, the Hishikari Au deposit and synchronous rhyodacite domes of Quaternary age in southern Kyushu, Japan, lie along the western shoulder of the Kagoshima graben (Izawa and Urashima, 1989) as part of a near-arc bimodal suite (Y. Watanabe, written communication, 2001).

Most of the LS Au deposits in the northern Great Basin of Nevada, including Ivanhoe, Midas, Mule Canyon, and Sleeper (Fig. 2e; Table 6), were generated from 16 to 14 Ma. The deposits and associated bimodal volcanic rocks lie within and near the northern Nevada rift, a product of backarc extension related to the Yellowstone hotspot and, according to some investigators, of mantle-plume activity (Noble et al., 1988; John et al., 1999; John, 2001). The rhyolites are reduced (ilmenite-series) and assignable to the tholeiitic suite (John, 2001). Extension, bimodal volcanism, and LS vein mineralization, most notably at Bullfrog (Fig. 2g; Table 6; Eng et al., 1996), became more widespread in the Great Basin once subduction ceased at the adjacent Pacific margin (John, 2001).

A somewhat similar situation prevailed in the backarc region of the Patagonian Andes in southern Argentina. Widespread LS vein deposits, including the vast Cerro Vanguardia system (Fig. 2e; Table 6), of Late Jurassic age – broadly coeval with the intra-arc Esquel deposit to the northwest – are hosted by voluminous rhyolitic ignimbrites, small rhyolite domes, and petrochemically related basalt and basaltic andesite (Schalamuk et al., 1997; Panza and Haller, 2002). The bimodal volcanism throughout Patagonian Argentina was coincident with a series of north-northwest-

striking half-grabens that presaged opening of the South Atlantic Ocean and, on petrochemical grounds, is attributed to the far-field effects of mantle-plume activity (Riley et al., 2001). Backarc rifting and bimodal volcanism during the Miocene in the Kitami region of northeastern Hokkaido, Japan, are intimately related to formation of many LS veins, of which Konomai (Table 6) is the largest historic producer. However, oblique subduction – and possible generation of a slab window – rather than mantle-plume activity is believed to be the cause of this volcanism (Watanabe, 1996). The Republic (Table 6) and Wenatchee LS vein deposits of Eocene age in northern Washington also formed in backarc grabens linked to slab steepening (Berger and Bonham, 1990) as andesitic-dacitic volcanism gave way to low-volume basaltic and rhyolitic activity (Wagoner et al., 1994; Margolis, 1989).

Intra- and backarc rifting, bimodal volcanism, and LS vein formation in island arcs, as discussed herein for southern Kyushu and northeastern Hokkaido in Japan, commonly appear to occur along strike from marginal ocean basins. This relationship is particularly well documented for the Taupo Volcanic Zone of North Island, New Zealand, with its many active geothermal systems, dominantly rhyolitic volcanism, and traces of LS mineralization (Barker, 1993). Nevertheless, a direct connection between onland rifting and marginal-basin opening is often disputed, as it is in northeastern Hokkaido (Watanabe, 1995). By contrast, rifting of the North American continental margin in northwestern Mexico and contiguous California as it overrode the East Pacific Rise is widely accepted. The basalt-dominated bimodal volcanism that encircles the Gulf of California contains a number of LS Au occurrences (Staudé and Barton, 2001), of which the Modoc prospect, consisting of both vein and disseminated mineralization, is the best described (Hillemeier et al., 1991).

An excellent example of postcollisional extension is provided by the Early Cretaceous rifts of the Mongol-Okhotsk belt in the Trans-Baikal region of Russia and contiguous northeastern Mongolia. Rifting resulted from tectonic collapse following collision between the Siberian and previously collided Mongolia and North China cratons (Fig. 2j; Zorin et al., 1999). The rifting gave rise to relatively small volumes of rhyolitic and basaltic rocks along with a few LS vein systems, including the giant Baley deposit (Zorin et al., 2001).

The characteristically low topographic relief of these diverse rift settings commonly leads to widespread accumulation of siliciclastic sequences of fluvial and/or lacustrine origin. The sedimentation is broadly contemporaneous with hydrothermal activity, and LS veins commonly occur near the sedimentary rocks, and in some cases are hosted by such units (e.g., Republic and Baley deposits; see above). Locally, LS disseminated Au mineralization accumulated in the fluviolacustrine sequences, either before lithification (e.g., Grassy Mountain, Idaho; Rytuba et al., 1991) or perhaps somewhat later (e.g., Wenatchee deposit; see above). Such settings are especially conducive to formation and preservation of hot-spring sinter, noted in most of the LS vein provinces discussed herein (Table 7), due to surface discharge in hydrologic depressions. By contrast, sinter is apparently absent from most IS epithermal systems – likely because of higher synmineralization relief and the resulting large hydrologic gradients and propensity for groundwater dilution. Sinter is also absent, as expected, from all HS deposits because of the prevailing

acidic conditions (Fournier, 1985).

Other geologic features observed near some rift-related LS deposits include metamorphic core complexes and related detachment faults, e.g., Republic graben, Washington (Holder and Holder, 1998), Trans-Baikal region (Zorin et al., 1999), and Ada Tepe and nearby prospects, eastern Rhodope massif, Bulgaria (Marchev et al., 2002); and rift-bounding listric normal faults, e.g., Lewis-Crofoot deposit, Nevada (Ebert and Rye, 1997). However, such low-angle faults generally do not host the Au veins.

A minority of LS epithermal deposits, albeit including several giants, is associated with a diverse suite of alkaline rocks in a bewildering array of convergent-margin extensional settings, but not normally in well-developed rifts. The alkaline magmatism occurs within or beyond inactive arcs, but also in backarcs and at near-arc termini during active subduction. The alkaline rocks are typically volumetrically restricted. They commonly appear to be generated in response to a wide variety of tectonic readjustments (Richards, 1995), including backarc extension preparatory to formation of the Rio Grande rift in the case of the phonolite and later mafic and ultramafic magmatism and Au-telluride veins at Cripple Creek, Colorado (Fig. 2f; Kelley et al., 1998); extension preparatory to opening of the North Fiji backarc basin in the case of the alkaline basaltic shield volcano and Au-telluride veins at Emperor, Fiji (Eaton and Setterfield, 1993); and backarc extensional faulting linked to either opening of the Manus backarc basin or slab flexure in the case of the trachybasalt, trachandesite, and latite volcanism and Au mineralization at Ladolam, in Lihir island, Papua New Guinea (Müller et al., 2002; Carman, this volume). The alkaline magmatism and LS Au mineralization at Cripple Creek appear to span the transition between subduction-related, calc-alkaline arc construction and postsubduction bimodal magmatism (Kelley et al., 1998). In stark contrast, the Zone VII LS veins and vein breccias at Porgera, Papua New Guinea, are associated with alkaline gabbro and compositionally related feldspar porphyries of intraplate chemistry that were emplaced during development of a fold-thrust belt resulting from arc-continent collision (Fig. 2i; Richards and Kerrich, 1993). Several of these alkaline rock-related deposits, including Cripple Creek, Porgera, and Ladolam, may be considered to possess geologic features that are atypical for LS epithermal deposits (Sillitoe, 2000b).

The exception to the rule that LS Au mineralization is generated during regional extension is provided by the McLaughlin deposit in the Coast Ranges of California, where nearby age-related basaltic andesite is an outlier of the bimodal Clear Lake volcanic field of Plio-Pleistocene age. Although extension prevailed at the deposit scale, the volcanism and related LS Au mineralization are components of a transpressive setting within the San Andreas transform fault system (Fig. 2h; Tosdal et al., 1996).

VMS deposits as submarine analogues of LS settings

Many polymetallic or Kuroko-type volcanic-associated massive sulfide (VMS) deposits occupy submarine equivalents of the subaerial rift settings that host LS epithermal deposits, as emphasized by the bimodal character of their associated

volcanic sequences (Sillitoe, 2002a). Intra-, near, and backarc, as well as postcollisional, rifts have all been identified to contain VMS deposits (Barrett and MacLean, 1999; Sillitoe, 1999b). As with LS veins, VMS deposits are commonly formed in or alongside domes of rhyolitic to rhyodacitic composition. Some, but not all, of these felsic volcanic rocks in both LS epithermal and VMS districts are the products of magmas that tend to be hotter, drier, and more reduced than usual for arc magmas of comparable composition, probably due to the lack of an appreciable subducted-slab contribution to magma genesis (Lentz, 1998; John, 2001).

The commonality between the rift settings of subaerial LS epithermal and submarine VMS deposits raises the possibility that transitional deposit types may exist where shallow-water conditions prevailed. Indeed, the Eskay Creek VMS deposit in British Columbia, which formed during bimodal volcanism and rifting of an andesitic arc, possesses feeder-zone textures and metal content, including bonanza Au and Ag grades and abundant As, Sb, and Hg, reminiscent of the epithermal environment (Roth et al., 1999; Hannington et al., 1999). Subseafloor fluid boiling resulting from the relatively shallow depth of seawater is used to explain deposition of the epithermal suite, especially the abundance of Au. A recent report of Au-rich, base-metal deficient veins with late realgar and orpiment at about 1,050-m water depth atop Conical seamount, 20 km south of the Ladolam LS Au deposit, notes high Au grades contained in dendrites with amorphous silica (Petersen et al., 2002). These authors suggest this system, which is associated with alkaline magmatism sharing the same characteristics and origin as that related to Ladolam, may represent a transition between traditional epithermal and VMS deposits.

Epithermal Ore-Fluid Compositions

The distinct differences between epithermal deposit types, specifically the ore mineralogy and metal complement, are largely controlled by the composition of the ore fluid. This is in part caused by the difference in solubilities of Au, Ag, and base metals as a function of the available complexes (Henley, 1990), with Au and Ag transported as sulfide complexes, and Ag plus base metals soluble as chloride complexes at temperatures of epithermal ore deposition. In addition, the S activity and redox state of the fluid affect the mineralogy (e.g., Barton et al., 1977; Einaudi et al., this volume). Therefore, we review the compositional variation of epithermal fluids, largely the salinity (where wt percent NaCl refers to the equivalent NaCl + KCl + CaCl₂; this value may be overestimated in fluid inclusion studies due to the presence of dissolved CO₂ and other gases; Hedenquist and Henley, 1985), as a first-order estimate in examining the role that volcanotectonic setting has on fluid composition and hence ore characteristics. Some of the nomenclature used in epithermal classification (Table 1) is based on the alteration mineralogy associated with ore bodies, whereas the sulfidation-state terminology refers specifically to some of the ore and related sulfide minerals.

Although alteration and gangue mineral assemblages are an important indicator of fluid composition, particularly of the reactivity (i.e., acidity) of a fluid, it is the ore-mineral assemblage itself that reflects the nature of the ore fluid. Indeed, in some cases, the ore fluid is distinctly different from the fluid responsible for alteration and

gangue mineral deposition. For example, highly acidic condensates of magmatic vapor are responsible for forming the early-stage leached cores with advanced argillic halos that characterize lithocaps (e.g., Sillitoe, 1995). Most HS deposits are hosted by lithocaps, but not all lithocaps, which are themselves essentially barren on initial formation, are subsequently mineralized with HS minerals. The latter requires ascent of a moderate-salinity ore fluid distinctly different from the early vapor (Table 8; Hedenquist et al., 1998). Likewise, the fluid responsible for quartz and carbonate gangue deposition in IS vein deposits is commonly much lower in salinity than the episodic pulses of saline fluid that deposit the ore and related sulfide minerals (Tables 5 and 8; Simmons, 1991; Albinson et al., 2001). There is also a suspicion that the event related to deposition of exceptionally high-grade bands of Au as dendrites in layers of silica gel, with both the Au and silica transported as colloids (Saunders, 1994), may be distinct from the fluid flow in LS veins during much of the lifetime of the hydrothermal system. Saunders et al. (1996) concluded that bonanza Au deposition in northern Nevada was related to episodic processes, such as fracturing, that led to rapid fluid ascent, flashing, and extreme disequilibrium conditions. They also predicted a relationship with rifting and mafic magmas, as subsequently demonstrated by John (2001).

HS ore fluids

HS ore deposits are commonly equated with acidic fluid because of the extreme leaching and quartz-alunite alteration during formation of the lithocap. Leaching requires a fluid with pH <2 to mobilize alumina (Stoffregen, 1987), and the alunite most likely forms at a pH of 2 to 3. Ore is hosted largely by the vuggy quartz zone, in which there are no aluminosilicate minerals left to indicate the pH or to influence any subsequently introduced fluid. However, at El Indio, for example, the post-enargite fluid responsible for formation of the Au-quartz-pyrite veins flowed through fractures that intersected propylitized rock, as well as the enargite-bearing core of the deposit. Where unleached rock was cut, veins have halos of quartz-sericite (Jannas et al., 1990), indicating a pH >4 to 5 for the Au-precipitating fluid. The Cu-stage HS mineral, enargite, is typically followed by an intermediate-sulfidation state, tennantite-dominated assemblage associated with the Au stage (Jannas et al., 1990), indicating a decrease in sulfidation state (Fig. 1), as well as the increase in pH, with time.

The early leaching fluid that precedes HS mineralization is a condensate of magmatic vapor, with a relatively low salinity (<1 wt% NaCl; Rye et al., 1992, Arribas, 1995; Hedenquist et al., 1998). Fluid-inclusion studies indicate that fluids with a wide range of salinities, from <1 to 45 wt percent NaCl, were present during the lives of HS deposits (Arribas, 1995). The hypersaline fluids, however, appear to have been present either early in the paragenesis and/or below the ore deposit, in places associated with sericitic and K-silicate alteration related to deeper intrusions (e.g., Summitville; Rye, 1993; Rodalquilar; Arribas et al., 1995). There are only five studies to date that have examined fluid inclusions hosted by enargite in epithermal deposits: Julcani, Peru (Deen et al., 1994), Lepanto (Mancano and Campbell, 1995), Radtka (Kouzmanov, 2001) and Chelopech (Moritz et al., 2002) in Bulgaria, and Bor, Serbia (Bailly et al., 2002). These studies found salinities with ranges of 8 to 18, 2 to 4.5, 9.6 to 10.2, 6.3 to 20.4, and 4.6 to 8.4 wt percent NaCl, respectively. Quartz associated with enargite from El Indio contains inclusions with salinities of 2.9 to 4.6

wt percent NaCl, whereas the salinity of the later Au stage is <1.7 wt percent NaCl (Jannas et al., 1990). Despite the distinctly different metal complement of lithocap-hosted Ag-Sn-Sb mineralization in Bolivia that is associated with reduced, ilmenite-series rhyodacite domes (Sillitoe et al., 1998), the salinities of the ore fluids are similar to the lithocap-hosted HS Au-Cu-As deposits, e.g., 6 to 13 wt percent NaCl at Cerro Rico, Potosí (Steele, 1996). These intermediate salinities are typical of the 5 to 20 wt percent NaCl range associated with sericitic alteration in porphyry Cu deposits (Reynolds and Beane, 1985). This observation, plus the common presence of sericitic (\pm pyrophyllite) roots to lithocaps in general, and most HS deposits in particular (Sillitoe, 1995; Sillitoe et al., 1998), led Hedenquist et al. (1998) to suggest that the HS ore fluids are affiliated with the sericitic stage of underlying porphyry deposits.

In view of the moderate salinities at the enargite stage in HS deposits, the chloride content of the early sulfide-precipitating fluids does not seem to be the ultimate control of whether ores are Ag-rich (e.g., La Coipa; Table 4) or Ag-poor (e.g., Goldfield; Table 4). Rather, Ag/Au ratios may be controlled by the metal endowment of the associated magmas, as appears to be the case in the lithocap-hosted deposits of the Bolivian Sn-Ag belt (Sillitoe et al., 1998). Detailed fluid inclusion and related studies are necessary to help resolve this question.

IS ore fluids

The IS deposits show a large range in metal complement and characteristics, from Au-Ag and Ag-Au deposits (division at Ag/Au \sim 50), with <0.05 wt percent combined base-metal content, to Ag-base metal and base metal-Ag deposits (Albinson et al., 2001). The quartz veins are typically crystalline and include Mn-bearing carbonates and silicates, as well as up to 10+ vol percent sulfides (Sillitoe, 1993a). Although there is a large range of sulfide and sulfosalt minerals, these are dominated by sphalerite with low FeS content, galena, tetrahedrite-tennantite, and chalcopyrite, all consistent with an intermediate sulfidation state (Fig. 1).

Precious-metal Au-Ag and Ag-Au IS deposits occur in the western andesite arc of Nevada (e.g., Comstock Lode; Vikre, 1989b; John, 2001), as well as in the Altiplano province of southern Peru (Noble et al., 1990), and in parts of central and eastern Europe (Lexa, 1999; Grancea et al., 2002; Ciobanu et al., 2002). This style of deposit is also present in the western Pacific region, including southwestern Hokkaido in Japan (Watanabe, 2002), the Baguio district in the Philippines, deposits in Papua New Guinea (e.g., Morobe goldfield), and elsewhere (Table 5; Sillitoe, 1989). Their base metal-rich IS cousins in central and northern Peru (e.g., Colqui; Kamilli and Ohmoto, 1977), and in the Philippines and Papua New Guinea (Sillitoe, 1989) typically contain sericite (fine-grained muscovite) rather than adularia, indicating higher temperatures and limited boiling, probably the result of greater formational depths. The Mexican epithermal province discussed above is host to all of the IS variants (Albinson et al., 2001), including precious-metal deposits: small Au-Ag and small to large Ag-Au deposits (e.g., Tayoltita, Pachuca, and Guanajuato) and base metal-rich Ag deposits (e.g., Fresnillo). As a group, the Mexican deposits have the best characterized fluid inclusion compositions (Albinson et al., 2001). Therefore, these deposits are used to highlight the correlation between salinity and metal

complement, although similar deposits around the world invariably show the same relationship (e.g., Baia Mare district, Romania; Grancea et al., 2002).

There is a regular increase in salinity, particularly in the maximum salinity, with increasing base-metal content and Ag/Au ratio in the Mexican IS deposits (Albinson et al., 2001). Although most deposits, regardless of metal content, have evidence for low salinities (minimum ≤ 2 wt % NaCl), careful study of several deposits provides evidence for sharp fluctuations in salinity between stages. At Fresnillo, quartz and calcite gangue has maximum salinities of < 2 wt percent NaCl, whereas sphalerite and Ag sulfosalt deposition was associated with intermittent injection of a brine containing 10 to 12 wt percent NaCl (Simmons, 1991). Thus, the wide ranges of salinity data for these and similar deposits most likely reflect to a large degree mixing of endmember fluids (as well, perhaps, as mixing of data from different stages). With this caveat in mind, Albinson et al. (2001) noted that the maximum salinity of individual Zn-Pb-Ag deposits is 12 to 23 wt percent NaCl, in contrast to 7.5 to 12 wt percent NaCl for the Ag-rich Zn-Pb occurrences. The Ag-rich precious-metal deposits have maximum salinities of 3.5 to 7.5 wt percent NaCl, whereas most Au-rich deposits have salinities of < 3.5 wt percent NaCl, after considering the likelihood that a portion of the apparent salinity is caused by dissolved gas (Albinson et al., 2001). Some of the Ag-base metal deposits have a late Au-rich stage with distinctly lower salinities (2–4 wt% NaCl average) than the early Ag-base metal stage (average 10 wt% NaCl; Table 8).

The gas content of fluids trapped in inclusions can be determined by freezing (Hedenquist and Henley, 1985) and extraction (Albinson et al., 2001), with some characteristic gas ratios being related not only to the magmatic component, but also to the type of magmatic affiliation (Giggenbach, 1995a; see below). N₂- and H₂S-rich inclusion fluids from Fresnillo and other Mexican IS deposits (Simmons, 1995; Albinson et al., 2001) possess magmatic signatures that are consistent with their calc-alkaline arc setting (Giggenbach, 1995a), and supported by stable isotopic compositions of water that indicate a magmatic component (Albinson et al., 2001). Not surprisingly, gases extracted from sulfide-rich samples have more pronounced magmatic signatures than barren quartz samples. Barren quartz also shows a large range of gas compositions, indicating fluid circulation through shallow crust. This is consistent with lower salinities and a meteoric water-dominant isotopic signature of gangue minerals relative to the sulfide-rich material, in Mexican (Albinson et al., 2001) as well as other IS deposits, such as Creede (Foley et al., 1989). Inclusion fluids of IS deposits in the Great Basin of Nevada appear to have a higher CO₂ gas content than the LS deposits there (as summarized by John, 2001).

LS ore fluids

The gangue minerals in most LS bonanza veins are dominated by bands of botryoidal chalcedony and cryptocrystalline quartz along with adularia and typically late calcite; illite or smectite may also be present. Argillic alteration halos consist of illite or illite-smectite. Electrum in many classic LS deposits is associated with pyrite, marcasite, arsenopyrite, pyrargyrite, acanthite, and Ag selenides such as naumannite. Minor chalcopyrite, FeS-rich sphalerite, and pyrrhotite may be present, although total sulfides, dominantly pyrite and/or marcasite, typically constitute < 2 vol

percent of ore zones (Table 3; Sillitoe, 1993a; John, 2001).

Fluid inclusions associated with LS deposits are typically difficult to study, as the principal host, quartz, has an extremely fine crystallinity, and there is even some question about the validity of the existing results if the chalcedony bands were originally precipitated as opal or an amorphous gel (Sander and Black, 1988). Nevertheless, the salinity of LS deposits in one of their type localities, northern Nevada, is invariably <2 wt percent NaCl (John, 2001). This is a maximum, as most studies have not determined how much of this apparent salinity is contributed by dissolved gas such as CO₂ and subordinate H₂S (Hedenquist and Henley, 1985). Studies of LS deposits elsewhere, such as Hishikari (Etoh et al., 2002), indicate similarly low salinities, <1.5 wt percent NaCl average (Table 6). In addition, as noted above, the Au-rich IS deposits in Mexico (>10-15 g/t Au, with Ag/Au <50, and base-metal contents <100-300 ppm) have apparent salinities largely <3.5 wt percent NaCl (Albinson et al., 2001). Thus, there is a clear relationship between low salinities and low Ag/Au ratios, as well as low base-metal contents (Henley, 1985, 1990; Hedenquist and Henley, 1985), irrespective of whether deposits are LS (John, 2001) or are IS and located within Ag and base-metal provinces (Albinson et al., 2001).

The fluids responsible for vein and bulk-tonnage LS Au deposits associated with alkaline rocks contrast markedly with those in the previously discussed LS environment linked to tholeiitic and calc-alkaline magmatism, especially where compositional bimodality is apparent (Table 6). If Cripple Creek, Emperor, Ladolam, and Porgera Zone VII are representative, then 3 to 9.6 wt percent NaCl may be considered as a typical salinity range (Table 6; Richards, 1995). Notwithstanding the fact that these deposits are Au-dominant and deficient in base metals, typical for LS deposits, the ore fluids are compositionally similar to the Ag-rich IS precious-metal deposits in Mexico. This stark contrast between the ore-fluid salinities of the two main LS subtypes (Tables 3 and 6) requires explanation.

Range of fluid compositions in active hydrothermal systems

A variety of papers have reviewed the compositional variation of fluids in active systems, both magmatic-hydrothermal and geothermal (e.g., Henley and Ellis, 1983; Giggenbach, 1992, 1997). Here, we focus on the fluid characteristics that appear to have a relationship to volcanotectonic setting, both salinity and gas content, for the insight these relationships may provide on the association between epithermal ore type and tectonics, linked as they are by the composition of the hydrothermal fluid.

At one extreme are geothermal systems with near-neutral pH and relatively reduced, deeply sourced fluids that are close to or in equilibrium with the altered host rocks due to their relatively slow ascent, i.e., rock-dominated environments (Giggenbach, 1992). The geothermal liquid is typically of low salinity, <1 to 2 wt percent NaCl, and may be gas rich, dominated by CO₂ with subordinate H₂S and N₂. Where this liquid discharges at the surface in a topographically low setting, boiling neutral-pH springs deposit silica sinter, as discussed above. The fluids are reduced (Giggenbach, 1992), similar to that noted for LS deposits (Einaudi et al., this volume). Although some geothermal systems with relatively saline waters also occur, they are notably uncommon and apparently restricted to closed sedimentary basins; the high density

of the deep liquid prevents discharge at the surface. The Salton Sea system is the best known example, and although the heat engine is magmatic, the 20 wt percent NaCl brine is thought to be amagmatic in origin, derived from dissolution of anhydrite of nonmarine evaporite origin (McKibben and Hardie, 1997).

At the opposite extreme, magmatic-hydrothermal systems occur proximal to volcanic vents that focus the discharge of magmatic vapors to the surface. Their principal surface expressions are high-temperature fumaroles and related condensates of extremely acidic water, pH~1. The acidic, oxidized fluid is far from equilibrium with the host rocks, as a result of its direct magmatic derivation and reflecting a fluid-dominated environment (Giggenbach, 1997). The reactive fluids in these magmatic-hydrothermal systems leach the rock and form the lithocaps that may subsequently act as a host to HS deposits. However, the fluids that are responsible for depositing ore in HS deposits, although oxidized (Einaudi et al., this volume), are less acidic, similar to those in IS deposits, based on the associated alteration mineralogy (see above). The HS fluids have a distinctly different evolutionary path from those related to LS deposits as they ascend into the epithermal realm (Einaudi et al., this volume).

Despite the broad spectrum of volcanotectonic settings, the salinity of fluids in a wide variety of volcanic-hosted geothermal systems around the world is limited to ≤ 2 wt percent NaCl, even at drilled depths of 2 to 3 km (Hedenquist and Henley, 1985; Giggenbach, 1995b; Simmons, 1995). However, within this low-salinity range, two groups can be distinguished: 1) Andesite arc-hosted systems, e.g., in the Philippines and Japan, have actual salinities up to 1 to 2 wt percent NaCl (up to 3.5 wt% NaCl for systems on the coast). 2) Felsic rock-hosted systems, such as in the Taupo Volcanic Zone, North Island, New Zealand, as well as basalt-hosted systems in Iceland, have maximum salinities of 0.3 to 0.5 wt percent NaCl. High gas contents in some of the felsic rock-hosted systems causes the fluid to have colligative properties of a solution (i.e., as determined from freezing measurements) with an apparent salinity much higher (2-3 wt% NaCl equivalent) than its actual salinity (<1 wt% NaCl; Hedenquist and Henley, 1985).

As noted above, volcanic-hosted geothermal systems worldwide have actual salinities less than that of seawater, typically <2 wt percent NaCl, with the exception being the apparently amagmatic closed-basin systems. What is the source of the high salinities in the HS, IS, and alkaline magma-related LS epithermal environments, from 3 wt percent to in excess of 20 wt percent NaCl? Studies of some active systems in the Philippines (Reyes et al., 1993) and elsewhere report the presence of saline fluid, ~25 wt percent NaCl or higher, in fluid inclusions associated with synhydrothermal dike intrusion. Such high-salinity fluid could be produced by phase separation at the solvus (e.g., Fournier, 1999), or even by direct exsolution from magma, although some argue for a basinal brine source. Regardless of source, the widespread occurrence of such high-salinity fluids in the epithermal environment is evidence for the presence of a brine reservoir (Simmons, 1991). If present in active systems, it must lie below the depths yet drilled (Fournier, 1987). Albinson et al. (2001) noted that a 20 to 25 wt percent NaCl brine at 320°C will not ascend above 1,000 m depth due to buoyancy constraints, except in transient situations where dike

intrusion or fault motion may cause injection to shallower levels (Simmons, 1991; Fournier, 1999).

Epithermal Fluid Sources

Composition of active systems relative to tectonic setting

The Taupo Volcanic Zone consists of a typical andesitic arc to the east and a bimodal assemblage in the western backarc, albeit with basaltic products being subordinate to the dominantly rhyolitic ash-flow caldera volcanism (Cole, 1990). The geothermal systems in these arc and backarc settings are distinctly different (Giggenbach, 1995a, b), and, as a group, they are perhaps the best studied in the world. Nevertheless, the arc-related analogy is far from perfect for HS and IS deposits, as maximum salinities observed or inferred in these and other arc-hosted systems are only about 1.5 wt percent NaCl. In addition, as noted above, there are a variety of settings for extension and bimodal volcanism, not all in backarc positions during active subduction. Despite these limitations, we review the discussion of Giggenbach (1995a, b) on the linkage between the two volcanotectonic settings in New Zealand and their affiliated hydrothermal fluid compositions.

Giggenbach (1995a) noted that N₂-rich gases in the geothermal (Christenson et al., 2002) and volcanic discharges along the eastern Taupo Volcanic Zone are typical of andesitic arc settings around the world, with the N₂, as well as much of the CO₂, contributed by volatiles released from subducted marine sediments. By contrast, the systems to the west are N₂-poor but He-rich, typical of those related to basaltic magmatism with a mantle affiliation, e.g., as noted at hotspots over mantle plumes and at mid-ocean ridges. Thus subduction- vs. mantle-related systems can be distinguished on the basis of their N₂/He, as well as CO₂/He, ratios, with high values being signatures of subduction. However, there is a large variation in the CO₂/Cl ratios of arc-hosted geothermal systems around the world (Giggenbach, 1995a), which is perhaps to be expected given the wide compositional range of these components in arc magmas and the processes affecting their exsolution (see below).

Giggenbach (1995a) estimated the endmember NaCl composition of the deep magmatic component in the New Zealand rift and arc systems, prior to meteoric water dilution, based on the isotopic compositions of the shallow, 1 to 2 km depth, 260° to 320°C geothermal waters. The rift-related systems to the west have a relatively small component of magmatic water, 5 to 10 percent, in contrast to the arc-related systems that contain a 10 to 20 percent magmatic contribution (and as high as 40-70 % in Japan and the Philippines). Giggenbach's calculation indicates that New Zealand rift-related parent magmatic water has a salinity of ~5 wt percent NaCl, whereas magmatic water in the arc setting is only ~1.5 wt percent NaCl. By contrast, the CO₂ and N₂ concentrations in the parent arc water, with CO₂ as high as 4 wt percent, are about 15-times higher than those of the rift parent water. These results indicate a larger degree (~3x) of meteoric water dilution of the magmatic component in the rift compared to the arc setting, because the salinity of shallow (~1 km depth) fluids in both settings is similar at about 0.2 to 0.3 wt percent NaCl. This observation for a magmatic component of at least 10 to 20 % is consistent with the proportion of

magmatic water in arc-hosted IS epithermal deposits, as synthesized for several deposits by Simmons (1995).

Giggenbach (1995a, b) attempted to link the differences in hydrothermal fluid composition between arc and backarc of the Taupo Volcanic Zone to the different magma sources in these two environments. He argued that gas-rich andesitic magmas feed the arc systems with their high CO₂ contents, whereas Cl-rich mafic melts are responsible for generating the higher salinity magmatic fluids deep in the back-arc rift. As in many arcs, the andesites of the Taupo Volcanic Zone are thought to be derived from high-alumina basalt that underwent varying degrees of crystal fractionation and assimilation of crustal material (Graham et al., 1995). The high water contents, derived from the subducting slab, promote a calc-alkaline differentiation trend resulting in andesitic arc magmas (Sisson and Grove, 1993), as well as favoring their shallow intrusion. We next review the volatile compositions of magmas in different volcanotectonic settings, and consider their potential effects on the ore-fluid variations observed from epithermal systems.

Volatile compositions of magmas

The H₂O, CO₂, S, and Cl compositions of deep magmas have been estimated from melt-inclusion studies of phenocrysts erupted from numerous settings (see reviews by Lowenstern, 1995, 2001; Giggenbach, 1996). Several recent studies on arc magmas provide an indication of the wide range in their volatile compositions. Studies of primitive magmas in subduction-related arcs (e.g., Anderson, 1973; Sisson and Layne, 1993; Matthews et al., 1994; Roggensack et al., 1997) provide evidence for S and Cl concentrations in high-alumina basaltic melts of ~1,000 to >2,000 ppm S and 700 to 1,700 ppm Cl. High-K arc basalts in Italy can have values up to 7,000 ppm Cl (Metrich et al., 1993). The H₂O concentrations typically range from 3 to 6 wt percent, but some can be as low as <1 wt percent (Sisson and Bacon, 1999). CO₂ values show a maximum of about 1,250 ppm, but with evidence for a wide variation that is attributed to rapid loss during crystallization between eruptions (Roggensack et al., 1997), due to the propensity for CO₂ to fractionate strongly from melt to a vapor phase (Giggenbach, 1995a; Lowenstern, 2001). In comparison, data from mid-ocean ridge tholeiitic basalts indicate CO₂ and S ranges of 1,000 to 3,000 ppm and 700 to 1,300 ppm, respectively, but with H₂O typically <0.5 wt percent and very low Cl, <200 ppm (see compilation by Giggenbach, 1996).

The concentrations of H₂O, CO₂, and Cl in silicic (>70 wt% SiO₂) melts are broadly comparable to those of the arc basalts, with 3 to 7 wt percent H₂O, 600 to 2,700 ppm Cl, and highly variable CO₂ concentrations, up to 1,000 ppm but locally <25 ppm (Lowenstern, 1995). However, the S concentrations in silicic magmas, usually <200 ppm and commonly <60 ppm, are markedly lower. Likewise, many less-silicic magmas, dominated by dacite, have S concentrations in the same low range (see compilation by Giggenbach, 1996). The much higher S values in mafic magmas are not surprising, given the significantly higher solubility of S in mafic vs. silicic melts (see review by Hattori and Keith, 2001).

In contrast to the high-alumina arc basalts, Sisson and Bronto (1998) argued that the primitive Mg-rich basalt of the 1982-1983 eruption of Galunggung volcano,

Indonesia, was generated by decompression melting during upwelling of mantle peridotite. The lack of a subduction component accounts for the anomalously low water content (<0.4 wt% H₂O). Nevertheless, the melt had up to 750 ppm CO₂, 750 to 1,400 ppm Cl, and 1,000 to >2,000 ppm S (Sisson and Bronto, 1998; de Hoog et al., 2001). Some melt-inclusion, as well as matrix-glass, compositions of <100 to 200 ppm S indicate at least some pre-eruptive exsolution of S (de Hoog et al., 2001), despite evidence that the melt was undersaturated with H₂O and CO₂ at depth (Sisson and Bronto, 1998). Thus, these observations indicate that a subduction setting may not be essential as a source of S-rich fluid.

Lowenstern (1995, 2001) stressed that most silicic magmas are saturated with a volatile phase to at least 10 to 15 km depth. Indeed, despite the variability of the volatile content of different melt compositions, Wallace (2001) agreed that prior to eruption most andesites and dacites, as well as some rhyolites and basalts, are saturated with a vapor phase consisting of H₂O, CO₂, and S-species, the last mostly SO₂ in the case of oxidized melts. The source of most of the S and CO₂ observed in the silicic systems is underplating by mafic magma (cf., Hattori, 1993; Hattori and Keith, 2001; Lowenstern, 2001, and references therein). The volatile saturation of magmas contributes to the potential for eruption, and also accounts for the observation that the amount of SO₂ erupted is 10-times or more greater than the amount that could have been derived from that dissolved in the melts themselves (Wallace, 2001). Prior to or instead of eruption, a portion of this vapor phase can be lost to hydrothermal systems that develop over the magma chambers.

In summary, subduction recycling contributes much of the H₂O, CO₂, and Cl in arc magmas and, in turn, to the associated hydrothermal systems. By contrast, much of the S is provided by a mafic component from the mantle, either during magma generation over a subduction zone, or directly by mafic intrusion, even if the melt is not saturated with water. The fact that the CO₂/Cl ratio in active hydrothermal systems is highly variable can be attributed to the large difference in the vapor-melt distribution coefficients, 250 for CO₂, but only 0.1 for Cl (e.g., Giggenbach, 1995a, b). Thus, the overall salinity and gas composition of the exsolved fluid is very sensitive to the timing and extent of vapor exsolution. For this reason, the CO₂/Cl of epithermal ore fluids should be highly variable, and the preliminary data discussed above indicates this to be so. The interesting factor that these observations highlight is the potential importance of mafic magmas in contributing S to hydrothermal systems, either indirectly via intermediate-composition magma chambers, or perhaps directly in the case of LS deposits affiliated to bimodal magmatic suites. This possibility is particularly relevant to Au-rich deposits, given the dependence of Au solubility on the reduced S content of the fluid (Henley, 1990).

HS deposits

As reviewed above, there is general acceptance that the early fluid responsible for the advanced argillic alteration that hosts HS deposits is generated by absorption of magmatic volatiles, notably HCl and SO₂, into meteoric water aquifers. However, the origin of the fluid subsequently responsible for introduction of most of the Cu and Au is less certain, with similar magmatic vapor (Sillitoe, 1983; Heinrich et al., 1999), magmatic brine (White, 1991), and unseparated magmatic fluid (Hedenquist et al.,

1998) all having been proposed. The HS fluid responsible for depositing enargite has intermediate salinity, 4 to 20 wt percent NaCl (see above; Table 8), a fact that constrains any model of HS Cu ore generation. However, much of the Au precipitates later in the paragenetic sequence, associated with intermediate-sulfidation state sulfide minerals, e.g., tennantite and chalcopyrite (Hedenquist et al., 1998; Sillitoe, 1999a; Jannas et al., 1999). Although direct data are not available for the sulfides of this stage, the associated quartz indicates lower salinities of <2 wt percent NaCl (Jannas et al., 1990; Hedenquist et al., 1994). Locally, the HS fluid may even become sufficiently neutralized to enable end-stage deposition of carbonate minerals (e.g., El Indio, Chile and Masonic, California-Nevada; R. H. Sillitoe, unpublished data). Notwithstanding the evidence for lower sulfidation states and more neutral-pH fluid conditions during the later-stage IS Au introduction in these deposits (e.g., Jannas et al., 1999), input of a distinct LS fluid to supply Au to HS deposits, with acidification and metal deposition (Berger and Henley, 1989), is not supported.

The contrast between metal suites characteristic of most HS deposits (Au, Ag, Cu, As) and those related to the more reduced, ilmenite-series igneous rocks in the Bolivian Sn-Ag belt (Ag, Sn, Sb) is most easily explained in terms of the different magma chemistries involved (Sillitoe et al., 1998), as there is apparently no appreciable difference in fluid salinities (cf., Steele, 1996). Sn-concentration, for example, is known to occur during fractionation of ilmenite-series magma because the metal acts incompatibly under reduced conditions (Lehmann, 1994).

The absence of lithocaps, barren or hosting HS deposits, from most rhyolitic volcanic sequences, especially those that contribute to bimodal magmatic suites, may be primarily a function of the relatively deep levels of parental magma chambers (Fig. 3). The tops of the felsic chambers in the Taupo Volcanic Zone are at least 5 to 6 km deep, near the base of the seismogenic zone, as documented geophysically and with geochemical and geologic constraints (Rogan, 1982; Bryan et al., 1999, Wilson et al., 2002). Deep magma chambers will exsolve lesser quantities of volatiles in general, and the acidic components that do exsolve will tend to react with rock during their extended ascent paths, leading to the "primary neutralization" process of Giggenbach (1992). By contrast, John (2001) suggested that the absence of HS deposits from the rift-related rhyolites of the northern Nevada rift is attributable to the relatively reduced character of the magma, which would have inhibited production of the SO₂ required for the formation of the sulfuric acid that, along with HCl, creates lithocaps that host subsequent HS deposits. In this regard, however, it is important to stress that the reduced ilmenite-series magmas of the Bolivian Sn-Ag belt did give rise to several alunite- and APS-bearing lithocaps, notwithstanding the likelihood that the SO₂ flux was lower than that typical of hydrothermal systems above magnetite-series intrusions (Sillitoe et al., 1998).

The same argument of deep magma emplacement cannot be used, however, to explain the rarity of lithocaps and HS deposits in alkaline volcanic centers because many alkaline intrusions are emplaced shallowly and are associated with porphyry Cu-Au deposits. Furthermore, there is clear evidence of higher SO₂ contents in vapor derived from alkaline as opposed to calc-alkaline magmas (Bailey and

Hampton, 1990; Hedenquist, 1995; Hattori and Keith, 2001), and there are enhanced Cl solubilities (for HCl generation) in alkali-rich, silica-poor magmas (e.g., phonolite; Signorelli and Carroll, 2000). Therefore, Sillitoe (2002b) postulated that the highly efficient acid-buffering capacities of alkaline igneous rocks subjected to unusually intense and commonly widespread alkali and carbonate alteration may offer a viable explanation for the relatively restricted development of leaching and advanced argillic alteration. The absence of the buffering capacity inherent in lithocaps thus precludes the development of high-sulfidation state sulfide minerals (cf., Jensen and Barton, 2000; Einaudi et al., this volume).

IS deposits

The common spatial and temporal association between HS and IS deposits in and around individual volcanic centers (Margolis et al., 1991; Sillitoe, 1989, 1999a), in conjunction with the fluid-inclusion and other evidence provided by IS deposits unrelated to HS systems (e.g., Simmons, 1995; Albinson et al., 2001), strongly favors a close magmatic relationship, albeit not usually as intimate as that displayed by the HS category. The occurrence of some IS deposits fringing HS centers may be explained in either of two ways. First, the deep fluid bypasses the lithocap entirely to generate IS veins in various distal positions (Fig. 3). Second, a deep IS fluid evolves to HS conditions on entering the unbuffered lithocap (Fig. 3), as mentioned above, followed by sufficient neutralization and reduction during outward flow and water-rock reaction to return to IS stability (Margolis et al., 1991; Sillitoe, 1999a; Einaudi et al., this volume). This transition between the HS and IS types gains further support from district-scale observations, particularly in the major Julcani district of Peru, the economically unimportant Pyramid district of Nevada, and the major carbonate-hosted Colquijirca district of central Peru. At Julcani and Pyramid, HS enargite-bearing veins with alunite halos are transitional along strike to IS veins richer in Zn and bordered by sericitic alteration (Deen et al., 1994; Wallace, 1980). At Colquijirca, abrupt transitions between unaltered dolostone host rock and HS Zn ore hosted by highly sulfidic quartz-alunite-kaolinite rock are marked by minor development of sphalerite and rhodochrosite of IS affinity (Fontboté and Bendezú, 1999).

An unresolved question is the origin of saline IS fluid with salinities of 5 to 20+ wt percent NaCl responsible for forming the base metal-rich Ag deposits (Table 8). As noted above, such fluids have not been sampled in active volcanic-hosted systems, and yet we know that they were present within several hundred meters of the paleosurface in some epithermal deposits. The careful study at Fresnillo by Simmons (1991) and others (e.g., Albinson et al., 2001) provides evidence for the transient injection of brines into otherwise low-salinity, <2 to 3 wt percent NaCl systems. Such dense brines presumably are present in a stably stratified, double-diffusive reservoir beneath the low-salinity, meteoric water-dominated portions of the systems (Fournier, 1987; McKibben and Hardie, 1997), only to be injected with their base-metal complement, perhaps as the result of periodic dike intrusion or faulting. If such brines are present beneath the epithermal environment but are not injected to shallow depths, the epithermal system will be mineralized only with precious metals. This scenario can explain the range of metal suites observed in regions such as the Sierra Madre Occidental and Altiplano of Mexico (Albinson et al., 2001). Sillitoe (1977) noted in this region that only deposits with Ag/Au <200 are >500 m above the

basement, with the base-metal sulfide-rich deposits deeper, presumably reflecting the generally greater depths of the higher salinity fluids. The origin of the deep brines is likely related to exsolution from an intrusion (e.g., Simmons, 1991). However, the reason that the saline IS fluids generate mainly Zn-Pb-dominated mineralization rather than the Cu-dominated mineralization typical of HS fluids of similar overall salinities remains unclear. One may speculate that the epithermal Zn-Pb ores are zoned from deeper Cu mineralization as the result of a thermal gradient, as seen in proximal to distal porphyry environments (Einaudi et al., this volume) or, alternatively, that the magmatic metal inventory was radically different.

LS deposits

LS epithermal deposits formed in association with alkaline igneous rocks show clear-cut relations to magmatism and, in places, to subjacent mineralization of porphyry or related types (Bonham, 1988; Richards, 1995; Jensen and Barton, 2000). This relationship is documented at Porgera where the Zone VII LS vein-breccia zone overprints an intrusion-related Au-Ag-(Zn-Pb) stockwork (Richards and Kerrich, 1993) and at Ladolam where a grossly subhorizontal breccia zone overprints low-grade porphyry Au mineralization (Moyle et al., 1990; Carman, this volume). These relationships suggest that deposit-scale rock buffering (Giggenbach, 1997) in alkaline rock-hosted systems, in contrast to calc-alkaline systems, is commonly efficient enough (Sillitoe, 2002b) to generate LS rather than HS or IS mineralization. The abundance of Au telluride minerals and of roscoelite, the vanadinite mica, in many LS deposits related to alkaline igneous rocks, such as Cripple Creek, Emperor, and Porgera Zone VII, may indicate a distinctive character of the magmatic input. Although Ag-, Au-, Bi-, and Hg-bearing tellurides are present in minor amounts in many HS, IS, and LS deposits, it is generally only in some of the alkaline rock-related systems that Te is sufficiently abundant for most of the hypogene Au to occur as tellurides. However, the IS vein deposit at Sacarimb, Romania (Table 5) and the complex HS-IS Au deposit at Kochbulak, Uzbekistan, both with Au largely present as tellurides (Alderton and Fallick, 2000; Kovalenker et al., 1997), are obvious exceptions to this rule. This intimacy between alkaline rock-related LS deposits and magmatic centers may also readily explain why their fluid salinities are several-times higher than those typical of all other LS deposits. Notwithstanding these higher salinities, comparable to those in IS Ag-Au deposits (Tables 5 and 8), the Ag/Au ratios of ≤ 1 are exceptionally low (Table 6). This observation may be taken to further emphasize the distinctive character of the magmatic input and its influence on metal budgets.

In marked contrast to all HS and many IS deposits, and to LS deposits associated with alkaline rocks, other LS deposits lack such clear-cut connections to magmatism (Heald et al., 1987; Hedenquist, 1987; Cooke and Simmons, 2000). However, a contribution from deep-seated magma chambers has been increasingly proposed (Matsuhisa and Aoki, 1994; Simmons, 1995) since the concept of wallrock flushing as a precious-metal source has become less generally accepted (e.g., Hedenquist and Lowenstern, 1994). The genetic uncertainty is particularly acute in the case of LS deposits associated with bimodal volcanism in rift settings, as discussed by John (2001) in the case of the Au deposits within the northern Nevada rift. Unfortunately, characterizing the LS ore fluid by means of stable isotope studies is hampered by

recrystallization of the amorphous silica to quartz, although associated clays do provide information that supports a magmatic fluid input (Faure et al., 2002).

As noted above, studies on the volatile contents of high-temperature, meteoric water-dominated geothermal systems in New Zealand by Giggenbach (1995a, b) revealed the existence of two distinct magmatic fluids that he correlated with volcanotectonic setting. Giggenbach's preliminary observations on the magmatic volatile contributions to systems of the Taupo Volcanic Zone, the site of minor LS epithermal mineralization (Barker, 1993), may have profound significance for the origin of epithermal precious-metal deposits, in particular those occupying the spectrum of bimodal rift settings that includes the Taupo backarc rift. It is tempting to suggest that the low sulfidation state, deficiency of base metals, Mn-carbonates, and Mn-silicates, and other characteristics of these rift-related LS deposits may be imposed at source by the distinctive nature of a magmatic contribution to the ore fluid. Although rhyolitic rocks, commonly in the form of domes, are closely related to many of the rift-related LS deposits, the basaltic component of the bimodal magmatic suites may play the fundamental role in the provision of the S, chloride, and even metals (Noble et al., 1988; Hattori and Keith, 2001). In this regard, it should be recalled that some LS deposits in alkaline magmatic centers display a close relationship to mafic rocks (e.g., Cripple Creek, Porgera). This suggestion of a direct relationship to bimodal magmatism is supported by the correlation of the reduced oxidation state of the associated magmas and that of the LS deposits themselves, at least in the northern Nevada rift (John, 2001). Nevertheless, the petrochemical evidence from many extension-related bimodal suites reveals a subduction-zone input to the mantle-derived basalts (e.g., Patagonian backarc; Riley et al., 2001), either from previous modification of the mantle by a subduction event or from directly contributed slab volatiles.

The outstanding characteristic of LS deposits in bimodal rift settings is the relative abundance of bonanza Au shoots (Sillitoe, 2002a). A bonanza Au vein may be informally defined as one containing at least 1 million metric tons averaging roughly 1 oz/t (30 g/t) Au or more. Such high-grade Au veins occur sparingly in a number of different epithermal provinces, although they constitute appreciably less than one percent of auriferous epithermal veins worldwide. Bonanza HS, IS, and LS veins are all well known, but 75 percent of all bonanza deposits are of the LS type (Table 9). Furthermore, almost 60 percent of all bonanza deposits occur in association with bimodal volcanism in rift settings (Table 9).

Although both structural and lithologic controls, or combinations of both, are commonly used to explain the localization of exceptionally high Au contents in epithermal veins (e.g., McKinstry, 1955), such reasons would seem unable to account for the clear dominance of bonanza veins in rift settings. Perhaps a more fundamental explanation is required. This might involve the ability of mantle-derived mafic melts to provide fluid intrinsically richer in Au or somehow able to generate higher Au concentrations. If the fluid were exceptionally rich in reduced S (H_2S) it would be able to carry and, during rapid boiling, precipitate more Au (e.g., Henley, 1990). As noted above, mafic magmas in general are S-rich (Hattori and Keith, 2001), and tend to be reduced, relative to other magma types, where in rift settings

(John, 2001; Einaudi et al., this volume). If the exsolved fluid ascended rapidly – as it might from a mafic magma chamber connected to the surface by rift-bounding faults, with pulses triggered by dike intrusion – the flashing and sharp cooling would lead to supersaturation of several components. This would result in the observed formation of silica and Au colloids in the epithermal environment, to be deposited as silica gel and bands of dendrites, respectively (Saunders, 1994), within only 50 to 300 m of the surface (Hedenquist et al., 2000). The common occurrence of bonanza Au veins within or near outcropping rhyolite bodies rather than basalts may reflect retention of much of the basaltic magma at depth as a result of density constraints. This suggestion is supported by the broad spatial coincidence of linear magnetic anomalies, ascribed to subsurface mafic dikes, and LS Au deposits in the northern Nevada rift (Ponce and Glen, 2002).

Notwithstanding the perceived similarity in volcanotectonic settings between LS epithermal and VMS deposits, the base-metal contents of the two deposit types are notably different. The elevated base-metal contents of VMS deposits, however, are readily explained (Henley, 1985) as a direct result of higher fluid salinities consequent upon the seawater presence in VMS hydrothermal systems. In contrast, base-metal contents are minimal in LS epithermal environments where the salinity of the mineralizing fluid is an order of magnitude lower than seawater concentrations. It is interesting to note, however, that the rift settings characterized by LS epithermal deposits, in contradistinction to the arc terranes that give rise to the IS and HS types, are not associated with porphyry Cu systems (e.g., John, 2001), a characteristic that seems to be shared by the similarly rift-related VMS deposits (Sillitoe, 1999b).

Conclusions

Although considerable uncertainty still surrounds the nature of magmatic contributions to epithermal precious-metal deposits, the broad correlations that are evident between epithermal types and subtypes and their volcanotectonic settings strongly suggest that fundamental epithermal characteristics are mainly dictated by the nature of the associated magmatism. The control of metal associations in lithocap-hosted deposits by magma type (oxidized vs. reduced), the compositional contrasts between IS deposits in arc terranes and LS deposits in rifts, including in some cases the inheritance of magmatic oxidation state, and the distinctive compositional features of LS deposits associated with alkaline magmas vis-à-vis those in bimodal suites, provide particularly compelling lines of evidence. An actualistic framework underpinning differences between epithermal deposits in subduction-related arcs (HS and IS) and rifts (LS) is provided by Giggenbach's (1995a, b) work on the Taupo Volcanic Zone, where he identified a clear distinction in hydrothermal character between the two volcanotectonic environments.

Exploration implications

Data and discussions presented herein suggest several conclusions that have implications for Au and Ag exploration in the epithermal environment.

(1) Major auriferous HS deposits occur in andesitic-dacitic arcs characterized by neutral to weakly extensional stress states, although compressive arcs characterized

by extremely limited coeval volcanic products are also capable of generating them. In such compressive settings, much older, nonvolcanic units may act as the immediate host rocks. IS deposits, typically of small size but locally economically important, such as the Victoria deposit at Lepanto (Claveria, 2001; Table 5), share the same spectrum of arc settings. Modestly sized porphyry Cu-Au or Cu-Mo deposits commonly occur in the same arc terranes as the HS deposits, as observed in the Cajamarca belt of northern Peru, the Maricunga belt of northern Chile, Nevados del Famatina in northwestern Argentina, Recsk-Lahóca in the Carpathians of Hungary, and Zhijinshan, China (as compiled by Sillitoe, 1999a). In contrast, the largest and highest-grade porphyry Cu deposits are favored by regionally compressive arc regimes (Sillitoe, 1998), albeit with stress relaxation coinciding with deposit emplacement (e.g., Tosdal and Richards, 2001). Such tectonic settings generally appear to lack major HS deposits, although they contain widespread barren advanced argillic lithocaps. Lithocaps are an integral part of porphyry development (e.g., Hedenquist et al., 1998). However, subsequent HS mineralization depends on the hydrothermal system hydrology, which is apparently not favorable in many compressive regimes. Where reduced, ilmenite-series volcanic sequences occur, Ag rather than Au is the dominant precious metal deposited in lithocaps, but to date such systems appear to be preserved only in the Sn-Ag belt of Bolivia.

(2) Alkaline volcanic centers, mostly in postarc, backarc, and postcollisional settings, offer potential only for LS deposits, albeit possibly of giant size, with IS and HS deposits being of only very minor importance (Sillitoe, 2002b). These alkaline rock-related LS deposits are typified by moderate fluid salinities. If unweathered, alkaline rock-related LS deposits, especially those of vein type, are likely to possess much of the Au in telluride minerals and, hence, pose metallurgical difficulties. Although many alkaline rock-related LS deposits are accompanied by intense and widespread alkali alteration, the low quartz contents of some of them, as exemplified by Cripple Creek (Lindgren and Ransome, 1906; Table 6), makes visual recognition difficult in the weathered environment, meaning that some outcropping examples may still await discovery (Sillitoe, 2002b).

(3) Most LS deposits are confined to bimodal magmatic suites in and around rifts generated during intra-, near-, and backarc extension, as well as in postcollisional settings. Although LS deposits are uncommon in andesitic-dacitic volcanic arcs, a few are formed with such rock suites where regional extension is prevalent. Some of these bimodal volcanic provinces are distant from convergent plate boundaries, the usual focus for epithermal exploration.

The greater chance of encountering bonanza grade Au ore shoots in low-salinity LS deposits provides a strong inducement to concentrate the search for bonanza veins in bimodal volcanic suites related to rifting. Nevertheless, the bimodal character of some volcanic provinces may not always be readily apparent from published geologic maps or during prospecting and reconnaissance geologic work for one of several reasons. First, the bimodal contribution may be volumetrically minor and visually overwhelmed by more voluminous preore andesitic rocks, as exemplified by the Esquel deposit (Sillitoe et al., 2002; Table 6). Second, the volumetrically and areally dominant rhyolitic products dwarf the mafic component in some extensional

provinces, such as the Patagonian backarc (Pankhurst et al., 1998). Third, bimodal volcanic products may be subjugated to fluviolacustrine sedimentary rocks, as reported for the Trans-Baikal rift zones of Russia and neighboring Mongolia (Zorin et al., 1999, 2001). Even where the bimodal character of volcanic provinces is correctly determined, LS deposits are commonly not easy to pinpoint visually because of the deficiency of alteration and pyritization in the immediate host rocks, a problem emphasized by John (2001) in the case of the northern Nevada rift. Nevertheless, as in the case of all epithermal deposits, alteration at very shallow levels in rift-hosted LS systems is generally more prominent because of the likelihood of preservation of blankets of advanced argillic alteration caused by acid leaching in the steam-heated environment above paleowater tables (e.g., Sillitoe, 1993a).

(4) The common volcanotectonic settings of LS epithermal veins and VMS deposits imply that LS veins may be found stratigraphically beneath or along strike of VMS provinces as parts of rift sequences that accumulated subaerially in response to more limited subsidence. An example is provided by the rhyolite-dominated extensional province of Patagonia, in which the dominantly subaerial parts contain LS deposits (including Cerro Vanguardia), whereas the restricted submarine part in the extreme south is characterized by polymetallic VMS prospects (Ametrano et al., 1999), both deposit types being of Late Jurassic age. Furthermore, rifted primitive arcs characterized by bimodal volcanism were, until the recent discovery of submarine VMS deposits in the Izu-Bonin arc (Iizasa et al., 1999), thought to be metallogenically infertile. Based on this discovery, could the subaerial (and, indeed, shallow submarine) parts of such primitive arcs, like Izu-Mariana, Tonga-Kermadec, and ancient analogues, be prospective for LS veins deposits?

(5) As outlined above, there are many types of extensional settings, and some are unrelated to subduction and the generation of high-alumina basalt. Tholeiitic basalts from mid-ocean ridges (e.g., East Pacific, Mid-Atlantic) and hot spots (e.g., Canary Islands, Galapagos, Hawaii) contain an order of magnitude less water and Cl than calc-alkaline andesites and rhyolites (Lowenstern, 1995; Giggenbach, 1996). However, they and their high-alumina arc basalt counterparts contain an order of magnitude more S and generally higher CO₂. This is not surprising given the high S solubility of mafic melts, >1,000 ppm, and even higher for alkaline mafic magmas, >3,000 ppm (Hattori and Keith, 2001). This observation raises the question of the importance of such mantle melts, some with a component from the subducted slab, in contributing S, as well as CO₂ and metals, not only to porphyry-forming systems (Hattori and Keith, 2001), but also to deep fluids responsible for forming LS (and indeed all) epithermal deposits. Are there regions of oceanic spreading (e.g., Iceland), or more likely, plume-related volcanism with unrecognized potential for high-grade LS vein Au deposits?

(6) Finally, many – but certainly not all – giant epithermal deposits of all three types, but especially LS deposits related to alkaline rocks, appear to be favored by unusual volcanotectonic settings that appear to reflect a variety of plate-boundary reorganizations (Sillitoe, 1993b; Barley et al., 2002). Therefore any special tectonic niche characterized by uncommon volcanic rocks might be emphasized above more normal arc and backarc provinces during grassroots exploration.

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Figures

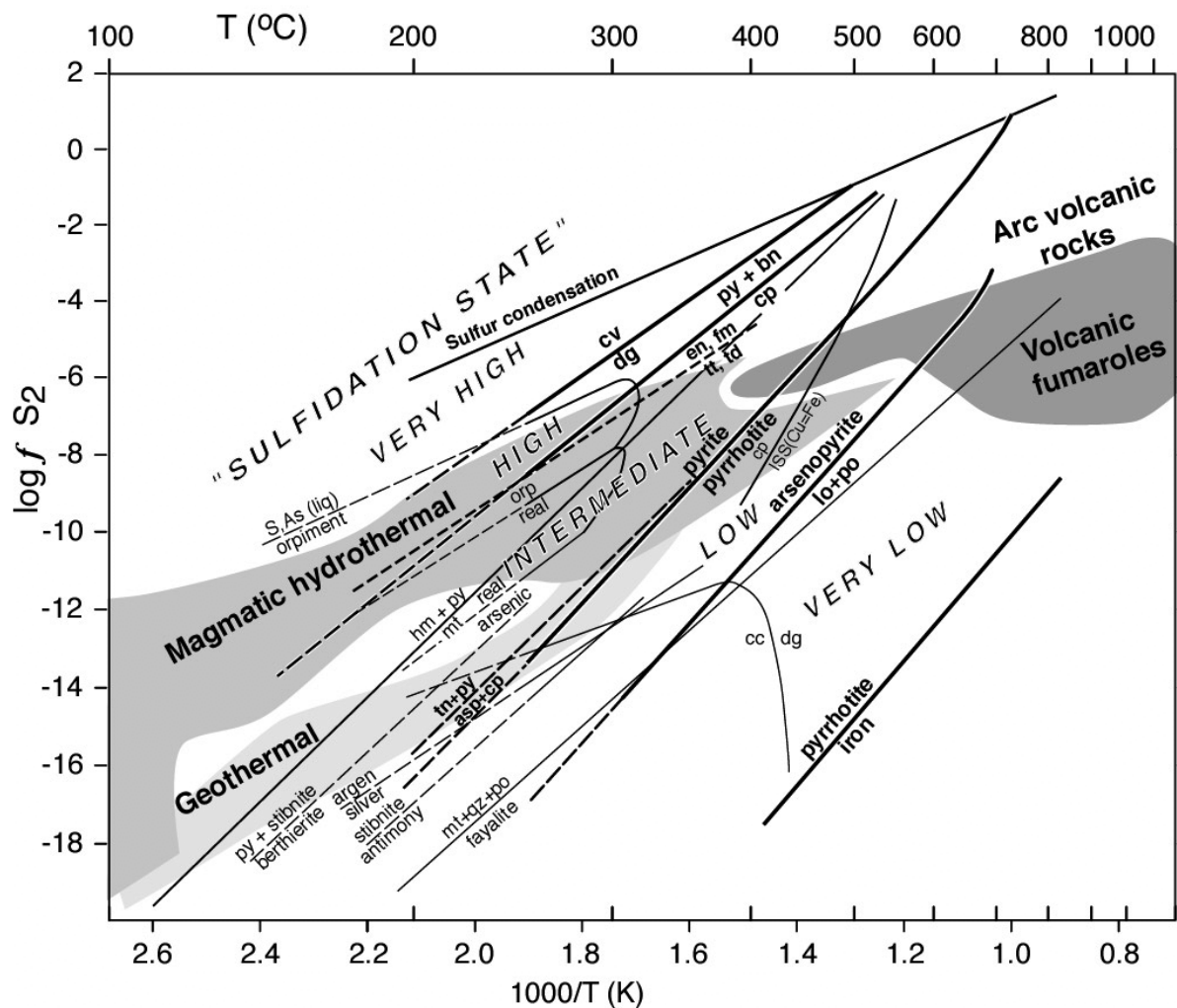


Fig. 1

Fig. 1: f_{S_2} - T diagram showing the variety of sulfide assemblages in epithermal deposits that reflect sulfidation state, from very low and low through intermediate to high and very high. Compositional fields of arc volcanic rocks, high-temperature volcanic fumaroles, magmatic-hydrothermal fluids, and geothermal fluids shown, as discussed by Einaudi et al. (this volume). Simplified from Einaudi et al. (this volume; see their figure 4 for mineral abbreviations).

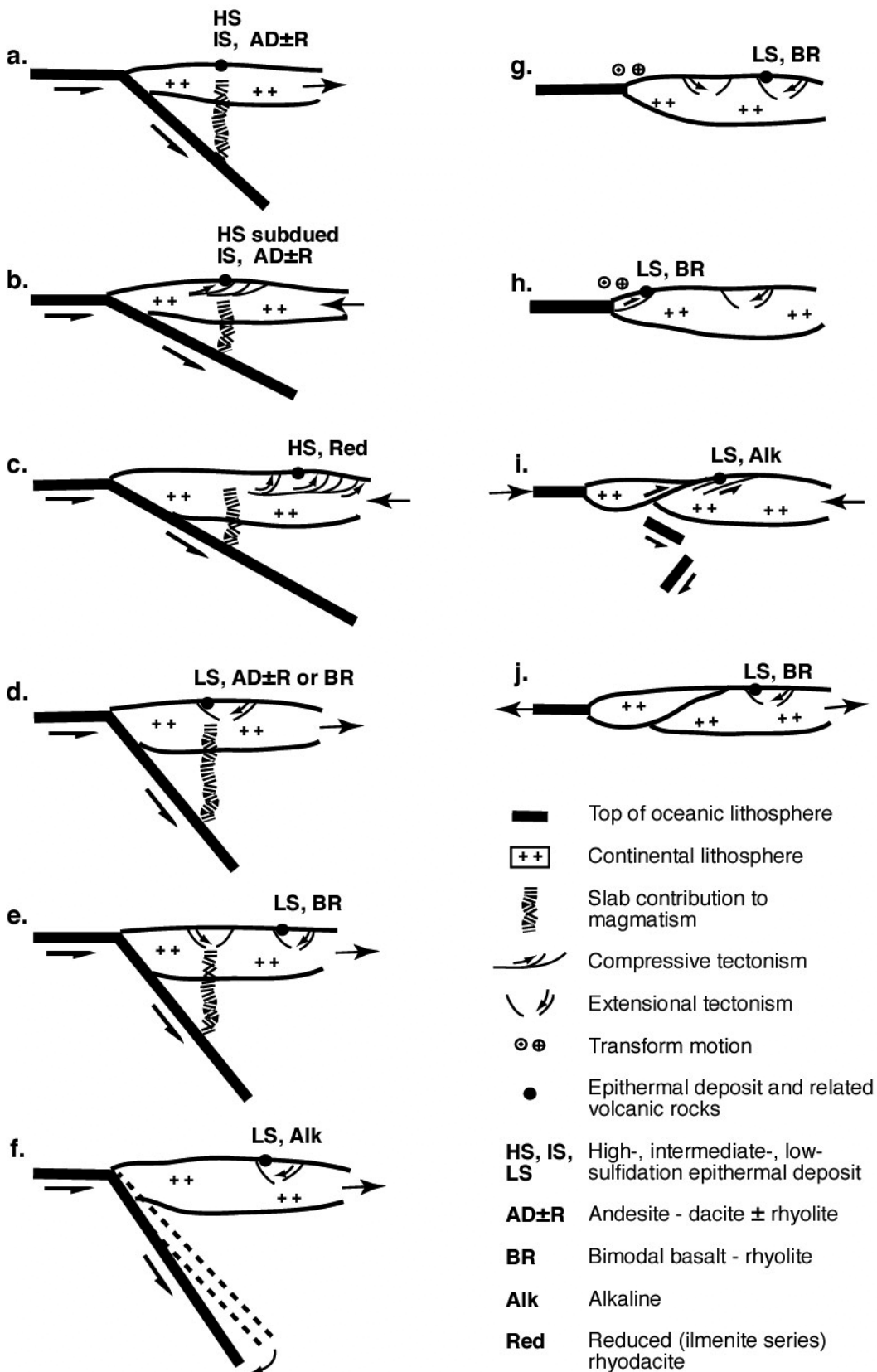


Fig. 2 Schematic continental margin-scale sections to illustrate selected volcanotectonic settings for HS, IS, and LS epithermal deposits: a. neutral-stress to mildly extensional arc (e.g., Yanacocha, Peru); b. compressive backarc during arc volcanism (e.g., Potosí, Bolivia); c. compressive arc with subducted volcanism (e.g., El Indio); d. extensional arc (e.g., Bodie, California or El Peñón, Chile); e.

extensional backarc during arc volcanism (e.g., Cerro Vanguardia, Argentina); f. extensional backarc during transition from subduction-related to rift-related bimodal magmatism (e.g., Cripple Creek, Colorado); g. extensional continental margin following cessation of subduction (and advent of transform faulting; e.g., Bullfrog, Nevada); h. compressive tectonism linked to transform fault boundary (e.g., McLaughlin, California); i. restricted postcollisional magmatism during collision-induced slab breakoff and compressive tectonism (e.g., Porgera, Papua New Guinea); and j. extension due to tectonic collapse following continental collision (e.g., Baley, Russia).

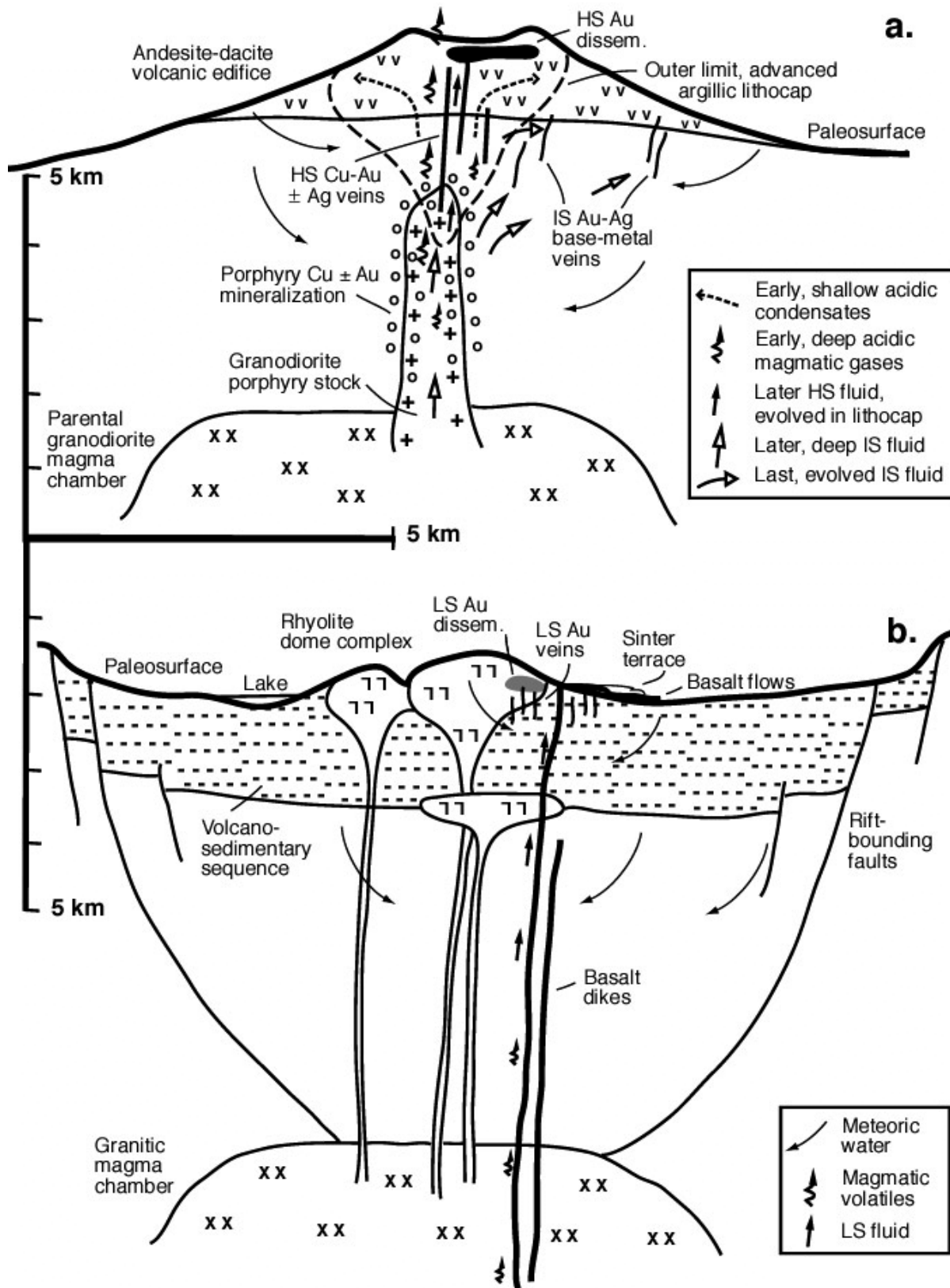


Fig. 3 Schematic sections of endmember volcanotectonic settings and associated epithermal and related mineralization types: a. Calc-alkaline volcanic arc with neutral to mildly extensional stress state showing relations between HS and IS epithermal and porphyry deposits (note that the complete spectrum need not be present everywhere). Early magmatic volatiles are absorbed into groundwater within the volcanic edifice (shown here as a stratovolcano, but it may also be a dome setting) to produce acidic fluid for lithocap generation, over and/or supra-adjacent to the causative intrusion.

Later, less-acidic IS fluid gives rise to IS mineralization, both adjacent to and distal from the advanced argillic lithocap. Where the IS fluid flows through the leached lithocap environment, it evolves to an HS fluid (Einaudi et al., this volume) to produce HS veins or disseminated mineralization, depending on the nature of the structural and lithologic permeability. The HS fluid may evolve back to IS stability during late stages, supported by paragenetic relationships and lateral transitions of HS to IS mineralogy. See text for details. b. Rift with bimodal volcanism and LS deposits. Deep neutralization of magmatic volatiles, typically reduced, results in an LS fluid for shallow LS vein and/or disseminated mineralization and related sinter formation (cf., Henley and Ellis, 1983; Giggenbach, 1992; John, 2001).

Tables

Table 1 Modern nomenclature for epithermal deposit types

Acid	Alkaline		Sillitoe (1977)
	Epithermal		Buchanan (1981)
Enargite-gold		Hot-spring type	Ashley (1982)
High sulfur	Low sulfur		Giles and Nelson (1982)
Acid sulfate	Adularia-sericite		Bonham (1986, 1988)
High sulfidation	Low sulfidation		Hayba et al. (1985), Heald et al. (1987)
Alunite-kaolinite	Adularia-sericite		Hedenquist (1987)
	Type 1 adularia-sericite	Type 2 adularia-sericite	Berger and Henley (1989)
High sulfidation	High sulfide + base metal, low sulfidation	Low sulfide + base metal, low sulfidation	Albino and Margolis (1991)
HIGH SULFIDATION	INTERMEDIATE SULFIDATION	LOW SULFIDATION	Sillitoe (1989, 1993a)
			Hedenquist et al. (2000)

Note: CAPITALIZED names used in this paper

Table 2 Reassignment of early epithermal classification schemes

	High sulfidation	Intermediate sulfidation	Low sulfidation
Ransome (1907)	Goldfield type		
Emmons (1918)	Alunitic kaolinic gold veins	Sericitic zinc-silver veins	Gold-silver-adularia veins Fluoritic tellurium-adularia gold veins
Lindgren (1933)	Gold-alunite deposits	Gold-quartz veins in andesite	
		Argentite-gold quartz veins Argentite veins Base metal veins	Gold-quartz veins in rhyolite Gold telluride veins Gold selenide veins

Table 3 Principal field-oriented characteristics of epithermal types and subtypes

	High sulfidation		Intermediate sulfidation	Low sulfidation	
	Oxidized magma	(Reduced magma) ¹		Subalkaline magma	Alkaline magma
Type example	El Indio, Chile (vein); Yanacocha, Peru (disseminated)	Potosí, Bolivia	Baguio, Philippines (Au-rich); Fresnillo, Mexico (Ag-rich)	Midas, Nevada	Emperor, Fiji
Genetically related volcanic rocks	Mainly andesite to rhyodacite	Rhyodacite	Principally andesite to rhyodacite, but locally rhyolite	Basalt to rhyolite	Alkali basalt to trachyte
Key proximal alteration minerals	Quartz-alunite/APS; quartz-pyrophyllite/dickite at depth	Quartz-alunite/APS; quartz-dickite at depth	Sericite; adularia generally uncommon	Illite/smectite-adularia	Roscoelite-illite-adularia
Silica gangue	Massive fine-grained silicification and vuggy residual quartz		Vein-filling crustiform and comb quartz	Vein-filling crustiform and colloform chalcedony and quartz; carbonate-replacement texture	Vein-filling crustiform and colloform chalcedony and quartz; quartz deficiency common in early stages
Carbonate gangue	Absent		Common, typically including manganiferous varieties	Present, but typically minor and late	Abundant, but not manganiferous
Other gangue	Barite common, typically late		Barite and manganiferous silicates present locally	Barite uncommon; fluorite present locally	Barite, celestite, and/or fluorite common locally
Sulfide abundance	10-90 vol %		5->20 vol. %	Typically <1-2 vol % (but up to 20 vol % where hosted by basalt)	2-10 vol %
Key sulfide species	Enargite, luzonite, famatinite, covellite	Acanthite, stibnite	Sphalerite, galena, tetrahedrite-tennantite, chalcopyrite	Minor to very minor arsenopyrite ± pyrrhotite; minor sphalerite, galena, tetrahedrite-tennantite, chalcopyrite	
Main metals	Au-Ag, Cu, As-Sb	Ag, Sb, Sn	Ag-Au, Zn, Pb, Cu	Au±Ag	
Minor metals	Zn, Pb, Bi, W, Mo, Sn, Hg	Bi, W	Mo, As, Sb	Zn, Pb, Cu, Mo, As, Sb, Hg	
Te and Se species	Tellurides common; selenides present locally	None known, but few data	Tellurides common locally; selenides uncommon	Selenides common; tellurides present locally	Tellurides abundant; selenides uncommon

APS, aluminum-phosphate-sulfate minerals

¹ See text for elaboration of this lithocap-hosted style of mineralization

Table 4 Selected features of HS epithermal deposits referred to in text

Deposit location	Au or (Ag) content, metric tonnes	Ag/Au Ratio	Age, Ma	Deposit style	Metal signature	Fluid-inclusion salinities, wt% NaCl equiv	Genetically related igneous rock	Tectonic setting	Reference (salinities)
Yanacocha, Peru	868	6-20	11.5-10.9	Disseminated bodies	Au, Cu, As	n.d.	Andesite-dacite tuffs and domes	Extensional continental-margin arc	Harvey et al. (1999)
Goldfield, Nevada	160	<1	21	Fault-controlled ledges	Au, Cu, Sb, As, Te, Se, Bi, Sn	Barite + quartz: 0.2-7.9	Rhyodacite domes	Extensional continental-margin Arc	Ashley (1974) (Vikre, 1989a)
Bor, Serbia	150	7	~85	Massive sulfide bodies and stockworks	Cu, Au, Ag, As, Zn, Pb	Enargite: 4.6-8.4	Andesite lavas and tuffs	Extensional continental-margin arc	Jankovic et al. (2002) (Bailly et al., 2002)
Chelopech, Bulgaria	195 (108.5 Rr, 34.5 P)	Low	78-74	Massive sulfide and breccia bodies, pipes, and stockworks	Au, Cu, As, Sb, Bi, Se, Te, Ag, Zn, Pb	Enargite: 6.3-20.4	Andesite lava and tuff + andesite-dacite minor intrusions	Extensional continental-margin arc	Bonev et al. (2002) (Moritz et al., 2002)
La Coipa, Chile	96	60	24-20	Vuggy quartz bodies and breccias	Cu, Au, Ag	n.d.	Dacite pyroclastics and dome	Neutral-stress continental-margin arc	Oviedo et al. (1991)
Pascua-Lama, Chile-Argentina	509	~30	9.4-8.1	Stockwork-disseminated bodies	Au, Ag, Cu, As	n.d.	None known	Compressive continental-margin arc	Bissig et al. (2002)
El Indio, Chile	>300	~10	7.6-6.2	Veins	Au, Ag, Cu, Te, As, Bi, W	Enargite-stage quartz: 2.9-4.6; Gold-stage quartz: <1.7	None known	Compressive continental-margin arc	Jannas et al. (1990)
Lepanto, Philippines	123	3.2	1.5-1.2	Fault-controlled vuggy quartz, massive sulfide, and breccia bodies	Cu, Au, Ag, Te, Se, Zn	Enargite: 2-4	Dacite diatreme and plug	Compressive island arc	Hedenquist et al. (1998) (Mancano and Campbell, 1995)
Pueblo Viejo, Dominican Republic	1364	5.5	~130 or ~77	Stockwork-disseminated bodies	Au, Ag, Zn, Cu, Pb, As, Te, Hg	n.d.	Dacite porphyry plug	Primitive island arc or normal continental-margin neutral-stress arc	Russell and Kesler (1991)
Potosí, Bolivia	(86,000)	>1,000	13.8	Veins and stockworks in massive and vuggy quartz	Ag, Sn, Sb	Barite: ~6.6	Rhyodacite porphyry dome	Compressive continental-margin backarc	Sillitoe et al. (1998) (Steele, 1986)

Note: Equivalent salinities have not been corrected for the possible presence of dissolved CO₂ (Hedenquist and Henley, 1985). n.d., no data.

Table 5 Selected features of IS epithermal deposits referred to in text

Deposit location	Au or (Ag) content, metric tonnes	Ag/Au ratio	Age, Ma	Deposit style	Metal signature	Fluid-inclusion salinities, wt% NaCl equiv	Genetically related igneous rocks	Tectonic setting	Reference (salinities)
Victoria, Philippines	80	~10	1.15	Sheeted veins	Au, Ag, Cu, Zn, Pb, As, Sb, Mn	Quartz and carbonate: <2.4; Sphalerite: <4	Dacite diatreme and plug	Compressive island arc	Claveria (2001) (Sajona et al., 2001)
Comstock Lode, Nevada	260 (5,950)	23	14	Veins and stockworks	Au, Ag, Zn, Pb, Cu	Quartz and Sphalerite: <1->6	Andesite-dacite lavas and domes	Neutral-stress arc	Vikre (1989b)
Tonopah, Nevada	(5,416)	94	20-18	Veins	Ag, Au, Zn, Pb, Cu, As, Sb, W, Mn	Quartz: 0.8-2.0	Andesite-dacite lavas and subjacent intrusion(s)	Extensional continental-margin arc	Bonham and Garside (1979) (Fahley, 1979)
Creede, Colorado	(2,493)	400	24.6	Veins and disseminated bodies	Ag, Au, Zn, Pb, Cu, Mn	Quartz, carbonate, sphalerite: 4-12	None known	Extensional continental-margin arc	Hayba et al. (1985) (Barton et al., 1977)
Fresnillo, Mexico	(>5,000)	1,393	~29	Veins, stockworks, mantos, and chimneys	Ag, Au, Zn, Pb, Cu, As, Mn	Quartz and carbonate: ~2; Sphalerite: 8.5-12.0	Quartz monzonite stock	Extensional continental-margin arc	Simmons et al. (1988) (Simmons, 1991)
Pachuca-Real del Monte, Mexico	(40,000)	216	21-20	Veins	Ag, Au, Zn, Pb, Cu, Mn	0-5.7	Rhyolite domes and dikes	Extensional continental-margin arc	Geyne et al. (1963) (Drier, 1976)
Tayoltita, Mexico	>150	62	38.6-31.9	Veins	Au, Ag, Zn, Pb, Cu, Mn	0-7.5	Andesite stock	Extensional continental-margin arc	Smith et al. (1982)
Arcata, Peru	10.5 (3450)	420	5.4	Veins	Ag, Au, Zn, Pb, Mn	Quartz: <2.6; Sph: >2.9	Andesite-dacite lavas and domes	Extensional continental-margin arc	Candiotti et al. (1990)
Sacarimb, Romania	~84	1-10	13-14	Veins	Au, Ag, Te, Zn, Pb, Cu, Sb, Hg, Mn	Quartz: 0-6	Andesite stock, lava, and pyroclastics	Extensional continental-margin arc	Alderton and Fallick (2000)
Rosia Montana, Romania	414	4.3	13.5	Breccia, stockworks, and veins	Au, Ag, Te, Zn, Pb, Cu	n.d.	Dacite lavas and tuffs + lacustrine sediments	Extensional continental-margin arc	Lexa (1999)
Baguio, Philippines	~400	~1	0.65	Veins, breccia pipes, and stockworks	Au, Ag, Zn, Pb, Te, Mn	0-5.1	Diatreme and dacite porphyry plug	Compressive island arc	Cooke et al. (1996)
San Cristóbal, Bolivia	(19,125)	>1,000	Late Miocene	Stockworks, disseminations, breccias, and veins	Ag, Zn, Pb, Cu, As, Sb	<5	Andesite-dacite dome complex	Compressive continental-margin backarc	Buchanan (2000) (Phillipson, 2000)
Kelian, Indonesia	~240	2	~20	Veins, stockworks, and breccias	Au, Ag, Zn, Pb, Cu, As, Mn	Quartz: 0.5-4.2; carbonate: 3-10.5; sphalerite: 4.5-6.8	Diatreme and rhyolite plugs and dikes	Probable neutral-stress or mildly extensional arc	van Leeuwen et al. (1990)

Note: Equivalent salinities have not been corrected for the possible presence of dissolved CO₂ (Hedenquist and Henley, 1985). n.d., no data.

Table 6 Selected features of LS epithermal deposits referred to in text

Deposit location	Au content, metric tonnes	Ag/Au ratio	Age, Ma	Deposit style	Metal signature	Fluid-inclusion salinities, wt% NaCl equiv	Genetically related igneous rock	Tectonic setting	Reference
Round Mountain, Nevada	380	~2	26	Disseminated body	Au, Ag	Quartz and adularia: <0.2	Silicic ashflow tuff	Extensional continental-margin arc	Henry et al. (1996) (Sander and Einaudi, 1990)
Bodie, California	45	5	8-8.5	Veins	Au, Ag, Zn	Quartz: <0.5	Dacite lava, tuff breccia, domes	Extensional continental-margin arc	Silberman and Chesterman (1991)
Waihi, New Zealand	199	10	7.2-6.6	Veins and stockworks	Ag, Au, Zn, Pb, Cu, Mn	Quartz and carbonate: <1.8	Andesite lavas, or possibly dacite or rhyolite	Neutral-stress to mildly extensional continental-margin arc	Brathwaite and Faure (2002)
El Peñón, Chile	103.8	17.7	59.4	Veins	Au, Ag	Quartz: <2	Rhyolite domes	Extensional continental-margin arc	Robbins (2000)
Esquel, Argentina	118	1.8	160	Veins	Au, Ag, As	n.d.	Rhyolite and basaltic andesite dikes	Extensional continental-margin arc	Sillitoe et al. (2002)
Hishikari, Japan	260	~2	1.15-0.6	Veins	Au, Ag, Se, Sb, Cu	Quartz: 0-1.4	Basaltic andesite and dacite lavas	Extensional island arc	Izawa et al. (1990)
Midas, Nevada	86	12	15.2-15.1	Veins	Au, Ag, Se	Quartz: 0.5-0.6	Rhyolite and basaltic andesite minor intrusions	Continental-margin extension: northern Nevada rift	Goldstrand and Schmidt (2000)
Sleeper, Nevada	52	1.3	16.1-15.5	Veins, vein breccias, and stockworks	Au, Ag, Se	Quartz: 0.1-0.6	Rhyolite dome	Continental-margin extension: northern Nevada rift	Nash and Trudel (1996)
Bullfrog, Nevada	83	1.5	10	Veins, vein breccias, and stockworks	Au, Ag	Quartz: <2	Rhyolite domes, plugs, and dikes	Continental-margin postsubduction (Basin and Range) extension	Eng et al. (1996)
Cerro Vanguardia, Argentina	143	14.2	~157	Veins	Au, Ag	Quartz: 1.5-3	Rhyolite ignimbrite	Extensional continental-margin backarc	Zubia et al. (1999)
Konomai, Japan	73	17	12.9	Veins	Au, Ag	n.d.	Rhyolite lava	Extensional backarc	Watanabe (1995)
Republic, Washington	93	5.9	50.3-50.1	Veins, stockworks, and breccias	Au, Ag, Se, As, Sb, Te, Hg	n.d.	Andesite and dacite lavas and minor intrusions	Extensional continental-margin backarc	Fifarak et al. (1996)
Baley, Russia	>1,000	0.5-3	120-114	Veins and stockworks	Au, Ag	Quartz: <8	None known; but alkaline basalt and rhyolite in area	Postcollisional rift	Zorin et al. (2001)
Cripple Creek, Colorado	834	0.1	31-28	Veins and disseminated bodies	Au, Ag, Te, Zn, Pb, Mo	Quartz: 3-9.6	Phonolite to alkali basalt diatreme complex	Extensional continental-margin backarc (pre-Rio Grande rift)	Kelley et al. (1998)
Emperor, Fiji	>150	<1	3.7	Veins and shatter zones	Au, Ag, Te, Cu, Mo, As, Hg	Quartz: 4.5-7	Banakititic volcanic rocks	Postsubduction island-arc extension	Anderson and Eaton (1990)

Ladolam, Papua New Guinea	1190	~1	0.35-0.1	Breccias and stockworks	Au, As, Sb, Te, Ag, W	Quartz: 3-7	Trachybasalt, trachyandesite and latite lava and tuff + monzodiorite	Extensional faults transverse to backarc	Carman (1995)
Porgera (Zone VII), Papua New Guinea	68	1	6	Vein and vein breccia	Au, Ag, Te, Cu, As	Quartz: ~4.2 and ~7.8	Alkali gabbro and mafic porphyry stocks and dikes	Postcollisional extension	Richards and Kerrich (1993)
McLaughlin California	109	3.5	2.2-0.75	Sheeted veins	Au, Ag, As, Sb, Hg	Quartz: 0.3-7.3 (av. 2.4)	Basaltic andesite lava and minor intrusions	Transpression linked to San Andreas fault system	Sherlock et al. (1995)

Note: Equivalent salinities have not been corrected for the possible presence of dissolved CO₂ (Hedenquist and Henley, 1985). n.d., no data.

Table 7 Epithermal Au provinces containing hot-spring sinter

Epithermal Province	Epithermal type	Magmatic suite	Typical sinter (lacking known mineralization)	Features	Reference
Western andesitic arc, Nevada-California	LS	Andesite-dacite	Bodie Bluff	Sinter terrace	Herrera et al. (1993)
Eocene grabens, Washington	LS	Andesite-dacite to bimodal transition	Golden Promise, Republic	Sinter clasts in hydrothermal eruption breccia	Fifarak et al. (1996)
Northern Nevada rift	LS	Bimodal	Buckskin Mountain, National district	Sinter terrace	Vikre (1985)
Lake Owyhee volcanic field, Oregon-Idaho	LS	Bimodal	Milestone prospect, DeLamar	Sinter horizons interbedded with hydrothermal eruption breccia	Rytuba et al. (1991)
West-central Great Basin, Nevada	LS	Bimodal (?)	Hasbrouck Mountain	Sinter terraces	Graney (1987)
Clear Lake volcanic field, California	LS	Bimodal	McLaughlin	Sinter terrace	Lehrman (1996)
Deseado massif, Patagonia, Argentina	LS	Bimodal	(Bocha)	Sinter terrace	R. H. Sillitoe, unpub. rept., 1998
Drummond Basin, Queensland, Australia	LS	Bimodal	(Verbena)	Sinter terrace (Late Paleozoic)	Cunneen and Sillitoe (1989)
Taupo Volcanic Zone, North Island, New Zealand	LS	Bimodal	(Waiotapu)	Sinter mounds and terraces (active vents and outflow)	Weissberg (1969)
Kitami region, northern Hokkaido, Japan	LS	Bimodal	Seta	Several sinter horizons	Yajima et al. (1997)
Hokusatsu district, southern Kyushu, Japan	LS	Bimodal	Hishikari	Sinter horizon	Nakanishi et al. (2001)

Table 8: Fluid compositions in epithermal and related environments (wt% NaCl equivalent)

Deposit type	Pre-main ore stage	Main ore stage I	Main ore stage II	Late (or post) ore stage	General references
Porphyry Cu-(Au)	Coexisting brine (35-70%) and vapor (<1%)	Potassic, Cu: brine + vapor	Sericitic (Cu): 5-20%	<5%	Roedder (1984); Reynolds and Beane (1985)
HS Cu-Au-As	Condensed vapor in lithocap; brine below or early ($\leq 45\%$)	Enargite: 4-20% max; quartz: 1-18%	Au stage: <2% ?; quartz: 1-18%	Barite, <5%	Jannas et al. (1990); Arribas (1995)
HS Sn-Ag-Sb	Condensed vapor (lithocap); brine below	Quartz: 10-13%		Barite (+ Ag): 6.6%	Steele (1996)
IS Zn-Pb-Ag		12-23% max			Albinson et al. (2001)
IS Ag-Zn-Pb	Quartz-calcite gangue: <2%	7.5-12% max in ore minerals		Quartz-calcite gangue: <2%	Albinson et al. (2001); Simmons (1991)
IS Ag-(Au)-Pb-Zn		Pb-Zn: 4-14 (av 10)%	Ag-Au stage, 1-9 (av 2-4)%		Albinson et al. (2001)
IS Ag-Au		3.5-7.5% max			Albinson et al. (2001)
IS Au-Ag		<3.5% (incl. gases)			Albinson et al. (2001)
LS Au-Ag (bimodal)		<1-2% (incl. gases)			Hedenquist and Henley (1985); John et al. (1999)
LS Au-Ag (alkaline)		3-9.6%			Richards (1995)

Note: Equivalent salinities have not been corrected for the possible presence of dissolved CO₂ (Hedenquist and Henley, 1985).

Table 9 Epithermal Au deposits containing major bonanza ore shoots (>30 metric tonnes Au, >30 g/t Au)

Deposit Location	Epithermal type	Magmatic suite
Republic, Washington	LS	Calc-alkaline andesite-dacite to mafic transition
Comstock Lode, Nevada	IS	Calc-alkaline andesite-dacite
Sleeper, Nevada	LS	Rhyolite (bimodal)
Midas, Nevada	LS	Rhyolite (bimodal)
Goldfield, Nevada	HS	Calc-alkaline andesite-dacite
Cripple Creek, Colorado	LS	Alkaline
El Peñón, Chile	LS	Rhyolite (bimodal)
El Indio, Chile	HS	Cessation of arc magmatism
Cerro Vanguardia, Argentina	LS	Rhyolite (bimodal)
Porgera (Zone VII), Papua New Guinea	LS	Alkaline
Hishikari, Japan	LS	Bimodal
Baley, Russia	LS	Bimodal

Note: References available in Tables 4, 5, and 6