Porphyry Deposits: Characteristics and Origin of Hypogene Features

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Abstract

Porphyry deposits arguably represent the most economically important class of nonferrous metallic mineral resources. These magmatic-hydrothermal deposits are characterized by sulfide and oxide ore minerals in veins and disseminations in large volumes of hydrothermally altered rock (up to 4 km³). Porphyry deposits occur within magmatic belts worldwide and are spatially, temporally, and genetically related to hypabyssal diorite to granitic intrusions that are porphyritic and that commonly have an aplite groundmass. The preponderance of Phanerozoic and most typically Cenozoic in age, which reflects the dominance of magmatism related to subduction tectonics and preservation in young rocks.

Porphyry deposits are here grouped into five classes based on the economically dominant metal in the deposits: Au, Cu, Mo, W, and Sn. For each porphyry class, the major metal concentration is enriched by a factor of 100 to 1,000 relative to unmineralized rocks of a similar composition. The mass of porphyry deposits ranges over four orders of magnitude, with the mean size of a deposit ordered Cu > Mo > Au > Sn > W. Hydrothermal alteration is a guide to ore because it produces a series of mineral assemblages both within the ore zones and extending into a larger volume (> 10 km³) of adjacent rock. The typically observed temporal evolution in porphyry ores is from early, high-temperature biotite + K-feldspar assemblages (potassic alteration) to muscovite + chloride assemblages (sericitic alteration) to low-temperature, clay-bearing assemblages (advanced argillization and intermediate argillic alteration), which is consistent with progressive greater acidity and higher fluid-to-rock ratios of fluids, prior to their eventual neutralization. Although advanced argillic alteration is relatively late in the deposits where it is superimposed on ore and potassic alteration, in the deposits where advanced argillic alteration (especially as quartz + alunite) is preserved spatially above ore and commonly extending to the paleosurface, it can form early, broadly contemporaneous with potassic alteration. In contrast, assemblages of Na plagioclase-actinolite (sodic-calcic alteration) and albite-epidote-chlorite-carbonate (propylitic alteration) form from a fluid with low acidity and commonly lack ore minerals. Geologic, fluid inclusion, and isotopic tracer evidence indicate magmatic fluids dominate acidic alteration associated with ore and nonmagmatic fluids dominate sodic-calcic and propylitic alteration. Veins contain a large percentage of ore minerals in porphyry deposits and include high-temperature sugary-textured quartz veins associated with ore minerals and biotite-feldspar alteration and moderate-temperature pyritic veins with sericitic envelopes.

The compositions of igneous rocks related to porphyry deposits can cover virtually the entire range observed for present-day volcanic rocks. Mineralizing porphyries are intermediate to silicic (>56 wt % SiO₂) and their aplite-textured groundmass represents crystallization as a result of abrupt depressurization of water-rich magma; however, small volumes of ultramafic to intermediate rocks, including lamprophyres, exhibit a close spatial and temporal relationship to porphyry ore formation in some deposits.

The understanding of porphyry systems depends critically on determination of the relative ages of events and correlation of ages of events in different locations, which in part depends on exposure. Systems with the greatest degree and continuity of exposure generally have been tilted and dismembered by postmineralization deformation. Most porphyry intrusions associated with ore are small-volume (<0.5 km³) dikes and plugs that were emplaced at depths of 1 to 6 km, though some were emplaced deeper. Deposits commonly occur in clusters above one or more cupolas on the roof of an underlying intermediate to silicic intrusion. Altered rocks extend upward toward the paleosurface, downward into the granitoid intrusion from which the porphyry magma and aqueous fluids were generated, and laterally for several kilometers on either side of a deposit. The underlying...

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magma chambers operated as open systems via magmatic recharge, wall-rock assimilation, crystallization, and intrusion. But mineralizing intrusions did not erupt.

Present-day distributions of hydrothermally altered rock and sulfide-oxide ore minerals are time-integrated products of fracture-guided fluid flow. We distinguish three spatial configurations characteristic of all five classes of porphyry deposits, the first of which has two variants: (1a) sericite alteration largely lies above and beside potassic alteration in a bell- or hood-shaped volume that narrows upward, as at Cherrapu, Henderson, and San Manuel-Kalama:zoon; (1b) sericite alteration is present with advanced argillic alteration, and the latter in some cases forms a broader zone at higher levels in the system, as at Batu Hijau, Cerro Rico, and El Salvador; (2) intense sericite and local advanced argillic alteration cuts through enclosing potassic alteration near ore but also extends above potassic alteration in an upwardly expanding zone with an overall geometry of a funnel, as at Butte, Chupic, and Resolution; (3) sodic-calcic, in addition to potassic, alteration is widespread in the center of the system and has an inverted cup-shaped volume under potassic alteration, with fingerlike projections of sodic alteration extending upward through the overlying orebody, as at Yerington. Metal grades are directly related to where ore minerals originally precipitate and the degree of subsequent remobilization. Precipitation of metals is a function of multiple variables, typically including temperature, acidity, and iron and sulfide availability. Hence, the shape of an orebody depends on the number and positions of mineralizing veins and barren intrusions; the proportions, shapes, and orientations of veins, lodes, or breccia; and pressure-temperature changes and wall-rock reactions that govern ore mineral stability.

Geochronology and thermal models suggest that durations of hydrothermal activity of 50,000 to 500,000 yr are common, but several large porphyry Cu deposits include multiple events spanning several million years. Crosscutting relationships, including offshoot veins, provide definitive evidence for the relative ages of hydrothermal events at a particular spatial location. Intrusive contacts that cut off older veins and are in turn cut by younger veins provide time lines that permit correlation of spatially separated events. Most porphyry deposits exhibit multiple intrusions, each associated with a series of hydrothermal veins formed over a declining temperature interval. The high-temperature starting point of hydrothermal fluid compositions varies systematically between porphyry classes and must reflect magma composition and chemical partitioning between melt, mineral, and aqueous fluid. Although the data are sparse, the magmas and associated high-temperature ore fluids vary such that oxidation state, sulfidation state, and total sulfide content are highest for porphyry Cu and Au classes, slightly lower for Mo, lower yet for Sn, and lowest for W. Nearly all classes and subclasses, however, have examples that diverge to low ax, ax, and high sulfide fugacity at lower temperature to produce advanced argillic alteration and high-sulfidation state ore minerals.

Just as with the spectrum of global magnetism, the breath of porphyry mineralization illustrates fundamental processes yet maintains distinctive geologic characteristics. In spite of a century of study and economic impact, many questions remain unanswered.

Introduction

The miner and the engineer naturally have little regard for a classification that does not help them to find ore.

—G. F. Loughlin and C. H. Behre, Jr., 1933

Porphyry deposits have played important roles in the history, science, and economics of modern society. The birth of the Journal of Economic Geology in 1905 coincided with the inception of large-scale open-pit mining at Bingham, Utah, several metallurgical innovations, and a long-term peak in the real Cu price. The Cu-rich portion of the spectrum of porphyry deposits was recognized first and in all respects remains the most important today, but Mo, Au, Sn, and W are important by-products and in certain deposits can be the principal commodity. The Economic Geology Seventy-Fifth Anniversary Volume reviewed porphyry Cu deposits (Titley and Bean, 1981) and porphyry Mo deposits (White et al., 1981). Porphyry Sn and W deposits had only recently been identified, and porphyry Au deposits had not yet been recognized.

Porphyry deposits are magmatic-hydrothermal deposits in which sulfide and oxide ore minerals are precipitated from aqueous solution at elevated temperatures. They draw their name from the porphyritic texture of dike- and pluglike intrusions commonly associated spatially and genetically to ore. The intrusive rocks associated with mineralization, porphyries, can be distinguished from other porphyritic rocks by their fine-grained aplitic (sugary) groundmass texture, which is closely linked to their genesis.

Porphyry deposits are characterized by their large tonnages and relatively low ore grades and geologically for the disseminated character of their ore minerals that occur both in narrow, closely spaced veins and within hydrothermally altered rock (e.g., Titley, 1966; Lowell and Guilbert, 1970). Many variations on the porphyry theme reflect differences in magmatic compositions, structural styles, host rocks, and other factors (Gustafson and Hunt, 1975; Gustafson, 1978; Einemild, 1982a). The most homogeneous porphyry deposits are uniform on the scale of hundreds of meters, yet the characteristic veins and their alteration envelopes have widths of millimeters and centimeters (e.g., Titley, 1982b). The system and deposit-scale patterns and processes are of practical and scientific interest, but a key geologic attraction of porphyry deposits is that many features are visible at the hand specimen scale and are amenable to rigorous, continuous documentation of paragenetic relationships during mapping and core logging.

The origin of porphyry deposits encompasses fundamental questions about tectonic, magmatic, geochemical, and metallogenic processes of the Earth. The geologic attributes of porphyry deposits, coupled with their persistent economic importance, account for why this deposit type has been at the forefront for developing geologic mapping techniques and applying new laboratory techniques and theoretical methods (Hunt, 1991; Hedenqrist and Richards, 1998). In turn, these approaches have led to many conceptual advances in
economic geology as a whole and influenced the understanding of crustal-scale mass transfer and fluid circulation (Skinner, 1967; Barton and Hanson, 1989; Burton et al., 1991b). The use of temporal constraints will be critical to advancing the understanding of these deposits and solving the larger scientific questions.

Scope and approach

Porphyry deposits display a remarkable diversity of characteristics that have several common genetic themes, as well as practical implications. Although porphyry Cu deposits and porphyry Mo deposits have received considerable attention, the geologic features and origins have not been well integrated across the spectrum of porphyry deposits. Indeed, the geologic literature on the Cu and Mo end members had virtually no reference to one another before 1980 and minimal overlap since then. Hence, we address the spectrum of porphyry deposits in order to examine whether genetic hypotheses developed in one portion of the spectrum are more broadly applicable. In this review of porphyry deposits, we highlight information commonly accessible to exploration and mining geologists, including geologic maps, hydrothermal mineral assemblages, chemical compositions of associated igneous rocks, metal ratios, and relative ages based on cross-cutting relationships. We emphasize how porphyry magmatic-hydrothermal systems evolve over spatial scales from <1 mm to >10 km and time scales of up to 50 m.y. duration, which we refer to as space-time evolution. Sillitoe and Perelló (2005) review the tectonic and magmatic evolution of porphyry deposits in the South American Cordillera over greater distances and longer time scales. During the past 25 yr new analytical chemical methodology and physical modeling techniques have expanded the scientific approaches to understanding chemical and physical processes that produce porphyry ore deposits. We take a more traditional approach, and due to space limitations only briefly summarize the key conclusions from studies of stable isotopes, fluid inclusions, radiometric ages, and the trace element and radiogenic isotope studies (see review by Hedenquist and Richards, 1998). The overall purpose of this review is to summarize the current data and the understanding of processes that form porphyry ore deposits, with an emphasis on diversity and complexity of these systems.

Geologic Framework and Context

In the early 1960s, the only known porphyry deposits were Cu rich, and these were mostly described as disseminated Cu deposits. According to Titley (1997), the first formal use of the term porphyry Cu deposit may have been by Emsen (1938). A few Mo-rich members became recognized about the same time.

Porphyry deposits are intimately associated with magmatic arcs formed along consuming plate margins (Sillitoe, 1972, 1976), although the association with arcs breaks down for certain deposits (Sillitoe, 1980, 2000; Westra and Keith, 1981). Because of their broad distribution in space and time, porphyries can be used to monitor Earth evolution (Meyer, 1981).

Classification scheme

The porphyry deposit type is divided into five classes based on the principal contained metals: porphyry Au, porphyry Cu, porphyry Mo, porphyry W, and porphyry Sn. The classes are further divided into subclasses that are based on compositions of intrusive rocks, major and by-product metals, and other distinctive features (Table 1), as fully discussed in a later section.

Global-scale distribution in space and time

The global spatial distribution of porphyry deposits generally outlines Phanerozoic orogenic belts (Fig. 1). Since the appearance of the Economic Geology Seventy-Fifth Anniversary Volume, many new deposits have been discovered; these include discoveries in the Andes (e.g., Escondida Norte,
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Abbreviations: AK = Alaska; AZ = Arizona; BC = British Columbia; dist. = district; CA = California; CO = Colorado; ID = Idaho; Indio. = Indonesia; Kazakh. = Kazakhstan; MT = Montana; NB = New Brunswick; NM = New Mexico; NSW = New South Wales; NV = Nevada; ON = Ontario; Phil. = Philippines; PNG = Papua New Guinea; TX = Texas; UT = Utah; Uzbek. = Uzbekistan; WA = Washington
Ujina, Gaby, Spence, and Toki, Chile; Antapaccay, Peru), Southwest Pacific (e.g., Grasberg and Batu Hijau, Indonesia; Boyongan, and Tampakan, Philippines), Australia (Cadia, New South Wales), and North America (e.g., Resolution or Magna Porphyry, Arizona; Mount Milligan, British Columbia; Pebble Copper, Alaska). Especially since the mid-1990s, central Asia has become the latest new frontier of porphyry exploration with the discovery of Oyu Tolgoi, Mongolia.

The distribution of ages of Phanerozoic deposits (Fig. 2) is consistent with earlier recognition of the dominance of Cenozoic and Mesozoic deposits (Meyer, 1981; Titly and Beane, 1981; Hunt, 1991). More deposits of Paleozoic age, however, are being discovered and developed, for example in the Altai belt of Eurasia (Watanabe and Stein, 2000; Perello et al., 2001; Yakubchuk et al., 2001, 2005), and the Tasman belt of Australia (Perkins et al., 1995).

Archean and Proterozoic examples of porphyry deposits occur in Fennoscandia and Canada (Gaid and Isokann, 1979; Fraser, 1993; Wanhairen et al., 2003), although most have small to moderate quantities of contained metal. In addition, there are probable porphyry Cu occurrences in the Pan-African orogen in Egypt, Jordan, and Saudi Arabia (e.g., Burgath et al., 1984; Jaques et al., 1999; Bekhit et al., 2001). Possible Precambrian porphyry Mo deposits include those in Ontario and Finland (Gaid and Isokann, 1979; Nunes and Ayres, 1992).

The global distribution of porphyry deposits is a complex function of numerous factors, including the uneven distribution of magmatism through time that is related to changes in plate configurations (e.g., Burke and Kidd, 1980). The Phanerozoic distribution, however, is in large part explained in terms of preservation and exposure of deposits. Porphyry deposits form relatively near the surface, typically from 1- to 6-km depth and some deeper than 6 km, and are subject to subsequent tectonism, erosion, and burial (e.g., Staude and Barton, 2001).

Mass-balance perspectives

Estimates of ore reserves and mineral resources, though critically dependent on the availability of data, mining method, and economic and social inputs, are useful for first-order comparison of deposits (Fig. 3). For any one class of deposits, the sizes vary by two to three orders of magnitude. The average grade of the major metal in all five classes of porphyry deposits is enriched by 100- to 1,000-fold compared to the average concentration in unmineralized rocks of a similar composition.

The sizes of porphyry deposits of all classes range over about four orders of magnitude, and both the size of the largest deposits and the mean size of a deposit in a class typically vary in the order Cu > Mo > Au > Sn > W (Fig. 3). The largest porphyry Cu deposits have two orders of magnitude more Cu than deposits of other classes. In contrast, the largest porphyry Mo deposits have contain less Mo than several porphyry Cu deposits, and the largest porphyry Au deposits contain less Au than many porphyry Cu deposits (Fig. 3).

Many dozens of other major and trace elements are concentrated, redistributed, or depleted within the much larger volume of rock (~10–100 times) affected by hydrothermal fluids. The total mass transfer in porphyry systems is, therefore, enormous. After magmatism and sedimentation, porphyry hydrothermal systems play a key role in redistributing elements in continental crust within the Phanerozoic.

Economic and historical perspectives

Porphyry deposits currently annually contribute in excess of US$40 billion globally, principally as Cu, Mo, Au, and Ag (U.S. Geological Survey, 2005); the economic impact of porphyry deposits is similar in magnitude to aluminum and is only exceeded by iron and steel. Porphyries are the most important source of Cu and are the source of nearly all Mo. Certain porphyry Cu deposits are important producers of Au or Ag but not Sn or W. Porphyry Cu and Mo deposits played leading roles in economic growth of the American West in the first half of the twentieth century (Hyde, 1998). More recently, they have been important in establishing industrial economies in Chile and Peru and could have a similar impact in countries such as Mongolia.

![Figure 2: Histogram of ages by classes of porphyry, keyed to porphyry belts.](image-url)
Exploitation of porphyry deposits benefited from key technological developments (Fig. 4), including the success of Daniel Jackling's bulk tonnage open-pit mining method in 1905, coupled with advances in milling Cu ores and in smelting and refining (Nadin, 1978). The economic and mining trends also strongly influence scientific studies of porphyry deposits as illustrated in the timeline (Fig. 4) by special issues of Economic Geology, a few key books, and collections of papers on porphyry deposits.

**Hydrothermal Alteration and Mineralization**

Ores and structure have been dealt with in detail, but the important changes which adjacent rocks of known composition have suffered are too often briefly dismissed, or even incorrectly indicated.

—W. Lindgren, 1902

The host rocks of porphyry deposits are hydrothermally altered to form characteristic types of gangue mineral wall-rock alteration, and they are mineralized by sulfide and oxide minerals. Together these constitute specific processes and effects that characterize and divide deposits of the porphyry type and that we refer to as alteration and mineralization. Wall-rock alteration and mineralization studies are relevant because they provide limits on the pressure-temperature conditions and composition of ore fluids and understanding of the controls on ore deposition. Furthermore, they represent the most easily recognized and broadly spatially distributed guides used in the exploration for new ore deposits. About the time of the golden anniversary of Economic Geology in 1955, economic geologists benefited from deeper open-pit exposures, where effects of hypogene processes could be more easily understood. During this period, more attention was paid to alteration and mineralization and the first syntheses on the topic were written (Schwartz, 1947; Burham, 1962; Creasey, 1966; Nakovnik, 1968). The physical-chemical foundation of economic geology was strengthened by experimental and theoretical studies in hydrothermal fluids culminating in several seminal review papers (Barton and Skinner, 1967; Meyer and Henley, 1967; Barton, 1970) that subsequently were updated and supplemented (e.g., Barton and Skinner, 1979; Barton et al., 1991a; Erni et al., 2003).

**Terminology**

Veins and their alteration envelopes: We use the term vein sensu lato but informally adopt names based on the vein-filling widths of >1 m for lodes, 1 m to 1 cm for veins, and <1 cm for veinlets. Typical veins in porphyry deposits have widths of less than a few centimeters and most are only a few millimeters wide.
Fig. 4. Graphical summary of the historical context of porphyry deposits in the 20th century. Major categories are world political and economic conditions, copper and molybdenum metal prices and production, mining and processing technologies, discovery and development of deposits, trends in industry, scientific approaches, and notable books and collections of papers. Principal sources are Mustard (1978), U.S. Bureau of Mines (1976), Saegert and Lewis (1977), Sanneh and Puzicha (1977), Nadin (1975), Layr (1991), Sillitoe (1995a), Taylor (1993), Tilley (1997), and Hyde (1999). Near the bottom of the diagram, “IGC (1933)” refers to the International Geological Congress that was held in Washington D.C. in 1933; the proceedings that were published later contained papers such as Eninn (1936); “Ore Deposits W States (1934)” is the Lindgren Volume (Finch et al., 1933).
Veins may have distinct, mesoscopically visible wall-rock alteration envelopes (also known as selvages or halos) that are generally developed symmetrically on either side of a vein as a result of diffusion of reactive components from vein into wall rock in pore fluids. Vein envelopes in porphyry systems can have a single mineral assemblage or can develop zoned envelopes (Meyer and Henley, 1987). Zoned envelopes are well described from the Main Stage lodes (Sales and Meyer, 1948) and porphyry Cu-(Mo) pre Main Stage veins (Meyer, 1965; Brimhall, 1977) at Butte, Montana, from sericitic, sodic-calcic, and endolomitic assemblages at Yerington, Nevada (Dilles and Einaudi, 1992), and in a range of settings at Henderson, Colorado (Carter et al., 1988b; Seedorff and Einaudi, 2004a).

**Textures:** Hydrothermal alteration produces textures that are distinct from replacement textures developed in the magmatic environment. For example, secondary (hydrothermal) biotite after magmatic hornblende commonly consists of randomly oriented aggregates of fine-grained biotite (shreddy biotite) intermixed with rutile, magnetite, and anhydrite (Carson and Jambor, 1974; Brimhall et al., 1985). Destruction of hornblende can range from partial to complete, but the outline of the former amphibole site may be well preserved. In contrast, magmatic reaction textures produced by a shift from stability of hornblende to biotite in the presence of melt produces a single, optically continuous crystal of biotite that overgrows and variably replaces hornblende.

The degree of textural preservation depends on the environment of formation and the intensity of alteration. For example, if hydrothermal biotite forms readily in both igneous feldspar and mafic mineral sites, or if hydrothermal feldspar forms readily in both igneous mafic and feldspar sites, then the igneous texture will be obscured or destroyed. In some cases, textural destruction may be characteristic of envelopes on certain types of veins; in other cases, it may develop in areas where alteration is so intense that adjacent alteration envelopes overlap about closely spaced veins or due to flow of fluids along fractures that are not resolved mesoscopically (Gustafson and Hunt, 1975; Titley, 1982b).

**Distinction between mineral assemblages and mineral associations:** A hydrothermal mineral assemblage is a group of minerals that appears to be stable together at the mesosopic scale and to have formed contemporaneously. Assemblages are geochemically significant because they can be used to estimate restricted physico-chemical conditions of formation assuming local equilibrium thermodynamics. The concept of assemblages is equally applicable to vein fillings and alteration envelopes. The complete mineral assemblage includes all minerals: silicate, sulfide, oxide, and other gangue phases such as anhydrite or titinite.

A hydrothermal mineral association is a group of minerals that occurs together, regardless of whether the minerals formed at the same time or in local equilibrium. Mineral associations are commonly reported in the geologic literature (and sometimes are mistakenly used as assemblages). For example, alteration zones (e.g., potassic zones) and sulfide zones (e.g., pyrite zone) are based on individual minerals or on mineral associations. Although these associations have value in exploration and science, by themselves they have little utility for understanding of dynamics because they are the result of time-integration of a number of events.

**Principal variables**

**Temperature:** Thermal decline drives hydrothermal fluids toward supersaturation with respect to many ore and alteration minerals (Henley et al., 1992; Redmond et al., 2004).

Hydrothermal systems operate from above magmatic solidus temperatures (750–800°C, depending on magmatic composition) to low temperatures (<200°C). The exposed portions of many porphyry systems formed over at least a 400°C range of temperatures (e.g., ~600–200°C; e.g., Wilson et al., 1980; Ulrich et al., 2001). One of the persistent controversies in porphyry deposits is the temperature of hypogene ore deposition, which may take place at the higher end of the temperature range in some deposits (e.g., Gustafson and Hunt, 1975; Eastoe, 1978; Carter et al., 1988b; Dilles et al., 2000a) or be confined to the lower end in others (e.g., Titley and Bean, 1981; Hedenquist et al., 1998; Redmond et al., 2004).

**Pressure:** Fluid pressure is a key variable because at <1.5 kbars low-salinity aqueous fluids undergo brine-vapor phase separation (Henley and McNabb, 1978) due to fluid immiscibility below a solvus, which in turn can have a pronounced effect on metal partitioning and solubility (Henley and Hunt, 1992; Henley et al., 1992, Heinrich et al., 1996). Pressure is a function not only of depth but also of the hydrologic regime. Overpressured and lithostatic conditions occur near magma, but hydrostatic conditions dominate and characterize all temperatures <350°C (e.g., Fournier, 1999). Silica solubility also is a strong function of pressure at >350°C (Fournier and Potter, 1982), as well as salinity (Xie and Walther, 1993). Quartz exhibits retrograde solubility at pressures of <800 bars, and the change from lithostatic to hydrostatic pressure is a likely cause for formation of abundant quartz veins (Rusk and Reed, 2002; Redmond et al., 2004). Pressure, however, is commonly poorly known. Geologic reconstructions, both stratigraphic and structural, are generally used to determine depth, and these lead to pressure estimates when coupled with interpretation of rock fabrics and vein characteristics.

Geochronometric methods for estimating pressure include fluid inclusion petrology (e.g., Roedder and Bodnar, 1980; Hedenquist et al., 1998) and mineral barometry (e.g., Anderson and Smith, 1995). Horizons of immiscible phase separation ("boiling") have been documented in porphyry deposits as well, but they do not have a direct temporal and spatial link to ore deposition (e.g., John, 1989a; Gustafson and Quiroga, 1995; Redmond et al., 2001).

**Chemical composition:** The bulk chemical composition pertinent to wall-rock alteration is dictated by the magmatic fluid composition (see below), wall-rock composition, and the proportions of these two which is also known as the fluid-to-rock ratio. Hence, wall-rock composition partly dictates alteration and ore mineral assemblages, particularly at low fluid-to-rock ratios typical of weak alteration. Wall-rock composition can buffer pH through reactions with carbonate or feldspar or oxidation state through reaction with iron-bearing minerals, organic matter, or reduced pore fluids. Because neutralization and reduction of sulfate to sulfide directly promotes sulfide mineral deposition (Barton, 1970), wall-rock reaction may be coupled with ore deposition. Wall rocks also can supply key
chemical constituents necessary to form hydrothermal minerals, such as mafic or oxide minerals supplying iron for Cu-Fe sulfides or feldspars supplying aluminum for epidote or garnet (e.g., Carten et al., 1988b; Seedorff, 1986; Zweng and Clark, 1995; Maher, 1986).

Wall-rock effects commonly are visible at the scale of disseminated grains to veins but also at the scale of ore zones and entire deposits, such as transitions between silicic and potassic alteration (Carten et al., 1988b) and contacts between silice and mafic rocks (e.g., Metz and Rose, 1986; Manske and Paul, 2002). Considering the differing behavior of metals like Cu and Mo, contrasting wall-rock compositions affect metal deposition, zoning, and ratios in otherwise similar deposits. Therefore, wall-rock effects have practical consequences for exploration targeting, ore reserve calculation, and grade control.

Composition of external fluids: The types of external fluids that participate in porphyry hydrothermal systems include saline formation waters related to coeval or older evaporite basins (e.g., Dilles et al., 1992, 1995; Battles and Barton, 1995), dilute meteoric waters (e.g., Sheppard et al., 1971; Selby et al., 2000), and seawater (e.g., Ostenko and Jones, 1976; Chivas et al., 1984). The paleohydrologic environment may be deduced by integrating evidence from regional geologic relationships, system-scale alteration map patterns, and the mineralogy and orientations of fracture patterns, mineral compositions, fluid inclusion compositions, stable isotope analyses, and studies of chemical gains and losses (e.g., Bowman et al., 1987; Dilles et al., 2000a).

Structural style: The three most common end members of structural styles of porphyry deposits are (1) the disseminated style, in which mineralization occurs predominantly in thin veins (Tittley, 1982b); (2) the lode style (Einaudi, 1977b; 1982a); and (3) the breccia style (Sillitoe, 1985; Skewes and Stern, 1996). The structural styles in part control the degree of involvement and timing of ingress of external fluids.

We use lode style loosely to refer to any of the major characteristics of zoned base metal veins such as the Main Stage at Butte (Meyer et al., 1988). Lodes are characterized by the presence of major through-going Cu-rich veins, minerals that are characteristic of high- and very high sulfidation states, and intense sericitic alteration with silification and/or advanced argillic alteration (regardless of geometry). Other types of late veins and lodes, e.g., those with weak seritic or intermediate argillic envelopes, are specifically excluded from the lode designation herein.

Wall-rock alteration-mineralization assemblages

Alteration types are groups of mineral assemblages that formed in geochemically similar environments (Meyer and Henley, 1967). Alteration types can be divided into those hosted by aluminosilicate rocks (most igneous, meta-igneous, metavolcanic, and clastic sedimentary rocks) and those formed in carbonate and ultramafic rocks (Barton et al., 1991a). Here we concentrate on the aluminosilicate rocks but address the consequences of encountering mafic host rocks. Four broad processes or types of chemical reactions are responsible for the principal alteration types in porphyry environments: (1) volatile addition (propylitic alteration); (2) hydrolysis (sericitic, advanced argillic, and intermediate argillic alteration); (3) alkali exchange (e.g., potassic and sodic-calcic alteration); and (4) addition of silica (silicic alteration).

We outline a geologic and geochemical framework for alteration-mineralization assemblages and alteration types that can be used descriptively by geologists (Tables 2, 3), without

| ab | albite               | cv | covellite          | plag | plagioclase          |
| act | actinolite          | das | diasporite         | po   | pyrophyllite         |
| AFS | aluminum phosphate-sulfate minerals (e.g., scanlonite, woodhouseite) | dick | dickite | py | pyrite |
| alun | alunite             | dag | digenite          | pyroph | pyrophylite |
| amphi | amphibole          | diop | diopside         | qtz | quartz |
| anat | anatase             | dun | dumortierite      | rhod | rhodonite |
| andal | andalusite         | en | enargite          | ri | rutile |
| anh | anhydrite           | fl | fluorite          | seh | scheelite |
| ank | ankrite             | flrkt | franklinite PbSbSnSb5 S14 | scree | screechite |
| ap | apatite             | ft | fannitite         | ser | sericite (line grained) |
| arg | argentine           | ga | galena            | sfl | sphalerite |
| As | native arsenic      | gt | garnet            | sn | scheelite |
| arp | arsenopyrite FeAsS | hbl | hornblende        | stan | stannite CuFeSnS |
| ba | barite              | il | ilinite           | td | titanite |
| Bi | native bismuth    | ilm | ilmenite          | titan | titanite (sphene) |
| bino | biotite (including phlogopite) | illm | ilmenomultite | tour | tourmaline |
| bi | biorite             | ill | ilmenomultite     | tn | tennantite |
| brkt | brookite           | iss | intermediate solid solution | tpz | tospaz |
| Bk | bismuthinite Bi2SiO5 | kaol | kaolinite         | ve | vernieculite |
| ca | calcite or carbonate of | ksp | K-feldspar | wol | woldramite |
| cass | cassiterite SnO2 | laz | lazulite MgAl2[PO4]2(OH)2 | zeco | zeolite, different |
| cc | chalcopyrite       | lael | laumontite        | zir | zircon |
| chl | chlorite           | mnc | marcasite         | zoi | zoisite |
| clz | clinopyroxite      | mno | molybdenite       | zun | zunyite |
| co | corundum           | mns | muntmorillonite   |     |     |
| cord | cordierite         | mt | magnetite         |     |     |
| cp | chalcopyrite       | rnc | nussite (coarse grained) |     |     |
chemical constituents necessary to form hydrothermal minerals, such as mafic or oxide minerals supplying iron for Cu-Fe sulfides or feldspars supplying aluminum for epidote or garnet (e.g., Carten et al., 1988b; Seedorff, 1998; Zweng and Clark, 1995; Muir, 1986).

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**Table 2. List of Mineral Abbreviations**

<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>Full Name</th>
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<tbody>
<tr>
<td>ab</td>
<td>albite</td>
</tr>
<tr>
<td>act</td>
<td>actinolite</td>
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<tr>
<td>APS</td>
<td>aluminum phosphate-sulfate minerals (e.g., scarnbergeite, woodhouseite)</td>
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<td>alunite</td>
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<td>zno</td>
<td>zonedite</td>
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<tr>
<td>zir</td>
<td>zircon</td>
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<tr>
<td>zoisite</td>
<td>zoisite</td>
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<td>Potassic (K-silicate)</td>
<td>Ksp field on T vs. K/H diagram; alteration of hornblende to aggregates of fine-grained (shredded) biotite and of plag to ksp with ca, fl, or anh. Bio and ab more common in wall rocks with higher mafic or Na content</td>
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<td>Transitional potassic-sericite</td>
<td>Geochemical environment near ksp-muscovite phase boundary on T vs. K/H diagram, i.e., T of ~355°C to ~365°C</td>
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<td>Muscovite (ser) field on T vs. K/H diagram; precursor ksp is altered to sericite + quartz; T (~355°C to ~365°C)</td>
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<td>Transitional potassic-advanced argillite</td>
<td>Forms near ksp-analal boundary on T vs. K/H diagram, with T &gt; 355°C.</td>
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<td>Transitional potassic-sericite-advanced argillite</td>
<td>Forms near intersection of ksp, musc, and analal fields on T vs. K/H diagram at T &gt; 355°C</td>
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| Transitional sericite-advanced argillite | Along boundary between muscovite and analal, pyrophyllite, or kaolinite on T vs. K/H diagram at T > 355°C | Ser, analal, pyrophyllite, kaolinite, (co), dias, analal, (ab) ± tpz, tour, dumort | Py + bn, cp, py, en, rt | APS?, scorz-laz?
<p>| Advanced argillite (high T) | Geochemical environment projects into analalite field on T vs. K/H diagram, i.e., T greater than ~350°C | Andal, (co), dias, (ab) ± tpz, dumort, cnd | Py, rt | APS?, scorz-laz? |
| Advanced argillite (moderate T) | Pyrophyllite field on T vs. K/H diagram, i.e., T between ~280° and ~350°C | Pyrophyllite, dias, (ab), tpz, dumort, zon, m, chl | Cv, dig, cc, bn, en, py, hm, rt | Alum, APS, scorz-laz, ba |
| Advanced argillite (low T) | Kaolinite field on T vs. K/H diagram, i.e., T less than ~280°C | Kaolinite, dias, (ab) ± tpz, dumort, zon; chl | Cv, dig, cc, bn, en, py, me, hm, rt, anat, bhtk? | Alum, APS, ba |
| Intermediate argillite | Ksp remains unaltered. Kaolinite and montmorillonite with sericite and chlorite replace biotite or plag rather than ksp or ab | Sericite, kaolinite, montmorillonite, chl | Cp, py, (po), km, sl, (asp, wol), cas, stan, bht, frit, km, ln, rt | Ca, rhod, fl, ba |
| Propylitic | Weakly metasomatized rocks with volatiles CO2 and H2O added; ore minerals scarce; refertilization of ksp ± mt | Ab, sericite, montmorillonite, chl, ep, (po, em, zo, act); refertilization of (bio, zir) | (Bn, cp, py, po, mo, sl, ga, bn, ln, tit, bht, rut, tit, m, titt, rt) | Ca, ank, ± relic ap |
| Silicite | Metasomatic addition of silica in excess of the silica generated by hydrolysis reactions of feldspars | Qtz, tour? | (Mo, fl, titt, zir) | Fl, (ap, mona) |
| Sodic-calcic | Sodic plagioclase replaces ksp, and act, chl, ep, (po, em, zo, act); chlorite replaces mafic minerals and mt | Act, ep, (plag) | Tit, (rt) | (Ca), ap |
| Sodic | Ab replaces ksp and plag (aca, ser, ep), chl, ep, tour, py, ser replace mafic minerals | Ab, chl, ep, tour, veron | Py, rt, (tit) | Ca, ap |
| Calcic-potassic | Metasomatic addition of Ca and K; igneous protoliths altered to calc-silicate minerals, bht, mt | Ksp, (ab), bio, gr, diop, act, ep | Bn ± mt, bn, (rff) | Ca, ank, ap |
| Sodic-ferric | Metasomatic addition of Na and Fe | Amph, plag | Mt, tit, mb, ep, ilm, tit | Ap |</p>
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<th>Notes on occurrences and processes</th>
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<td>Potassic (K-silicate)</td>
<td>Biotite veinlets; sugary quartz veinlets (A); veins with muscovite and potassic envelopes; banded quartz veins with analbite (B')</td>
<td>Formed by cooling of magmatic-hydrothermal fluids; ubiquitous in porphyry Cu and porphyry Au deposits; varying from high grade to barren; tsp present in porphyry W and certain porphyry Mo deposits; tour-qtz is an equivalent in porphyry Sn deposits</td>
<td>Refugio (Muntean and Einaudi, 2000); Bajo de la Alumbra (Proffett, 2003); El Salvador (Gustafson and Hunt, 1975); Grassie and Bell (Carron and Jamieson, 1974); Panguna (Ford, 1978); Rodrigo (Kimura et al., 1978); Henderson (Carter et al., 1988b); Lillaguna (Grant et al., 1990); Mount Pleasant (Koizma et al., 1996)</td>
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<td>Transitional potassic-sericite</td>
<td>Gray banded quartz veinlets; greisen veins; banded quartz veins with molybdenite (B')</td>
<td>Assemblages with co-existing ksp and ser or mus; tour commonly present in porphyry Cu deposits</td>
<td>Refugio (Muntean and Einaudi, 2000); El Salvador (Gustafson and Hunt, 1975); Hall (Shaver, 1991)</td>
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<td>Sericite (phyllitic, quartz-sericite-pyrite)</td>
<td>Veins with sericite envelopes (D'); greisen veins</td>
<td>Present in all classes of porphyry systems; assemblages generally are pyritic but may contain muscovite instead; tsp present in some porphyry W and Mo deposits; tour common in porphyry Cu and Sn deposits</td>
<td>Bingham (Parry et al., 2003); Chupicuaro (Ossandon et al., 2001); El Salvador (Gustafson and Hunt, 1975); Hall (Shaver, 1991); Henderson (Seedorff and Einaudi, 2004a); Chorolque (Grant et al., 1980)</td>
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<td>Transitional potassic-advanced argillic</td>
<td>Sugary quartz veinlets (A')</td>
<td>Known only from a few porphyry Cu deposits; patterns of distribution and presence of co suggest warming of fluids</td>
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<td>Transitional potassic-sericite-advanced argillic</td>
<td>Green micro veins (EB'); quartz-bearing veinlets with biotite and sericite (C', 'EDM')</td>
<td>Known to date only from a few porphyry Cu and Mo deposits; tsp-bearing veinlets with biotite and sericite at Los Pelambres are a major host of Cu</td>
<td>Butte (Brimhall, 1977); El Salvador (Gustafson and Quirga, 1995); Los Pelambres (Atkinson et al., 1996); Turnley Ridge (Steebel and Athion, 1954)</td>
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<td>Transitional sericite-advanced argillic</td>
<td>Sulfide veins</td>
<td>Ser can coexist with andal (high T), pyrox (moderate T), or kaoi (low T); may be common in porphyry Cu systems with abundant advanced argillic and sericite alteration</td>
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<td>Advanced argillic (high T)</td>
<td>Sulfide veins and massive replacements</td>
<td>Less common than transitional sericite-advanced argillic with coexisting andal and ksp, ab, or bio; co- and diabasic-bearing, tsp-poor assemblages indicate warming of fluids</td>
<td>Qonyalet (Koumadj, 1986); Kudryavtsev, 1996</td>
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<tr>
<td>Advanced argillic</td>
<td>Sulfide veins and massive replacements</td>
<td>Tpz common in porphyry copper deposits; diabasic-bearing, tsp-poor assemblages probably represent warming of fluids</td>
<td>Butte (Meyer et al., 1988); El Salvador (Gustafson and Hunt, 1975); Watanabe and Hedenquist, 2001</td>
</tr>
<tr>
<td>Advanced argillic (low T)</td>
<td>Sulfide veins and massive replacements</td>
<td>Tpz common in porphyry copper deposits; diabasic-bearing, tsp-poor assemblages probably represent warming of fluids</td>
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<td>Intermediate argillic</td>
<td>Py and base-metal veins; in some cases not obviously related to veins</td>
<td>Weathering and/or lower T form of hydrothermal alteration compared to sericite alteration can be py- or sericite-clay-chlorite (SCC) of Silico (Grant et al., 1984a)</td>
<td>Refugio (Muntean and Einaudi, 2000); Bingham (Parry et al., 2003); various deposits (Silicote and Gape, 1984); Henderson (Seedorff and Einaudi, 2004a); Aldenarai (Muntean and Einaudi, 2001); Bajo de la Alumbra (Proffett, 2003a); Bingham (Bray, 1960); El Salvador (Gustafson and Hunt, 1975)</td>
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<tr>
<td>Propylitic</td>
<td>Cao, ep, hm, and chl veins with ill-defined propylitic envelopes</td>
<td>Examples in all classes of porphyry deposits; in silicate igneous rocks, contains relic ksp, with plag altered to ab, ca, ep, and dusting of ser, ill, or mont</td>
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</tr>
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<td>Silicic</td>
<td>Quartz veins with silicic and potassic envelopes</td>
<td>High-T, proximal alteration type in certain porphyry Mo deposits as qtz + fli possibly in porphyry Sn deposits as qtz + tour; rare in porphyry Cu systems</td>
<td>Ann-Mason (Dilles and Einaudi, 1992); Yerington mine (Carter, 1986); various deposits (Dilles et al., 1978)</td>
</tr>
<tr>
<td>Sodic-calcic</td>
<td>Act and op veins with sodic-calcic envelopes</td>
<td>Formed by warming to &gt;350°C of saline fluids, including documented examples produced by sediments</td>
<td>Yerington mine (Carter, 1986): Ridgeway (Wilson et al., 2003); various deposits (Dilles et al., 1995); Galore Creek (Enns et al., 1985); Ridgeway (Wilson et al., 2003); various deposits (Lang et al., 1993)</td>
</tr>
<tr>
<td>Sodic</td>
<td>Qtz, py, tour veins have ah-chl envelopes</td>
<td>Low-T form of sodium metasomatism compared to sodic-calcic alteration: complex origin due to overprinting relationships</td>
<td>El Salvador (Gustafson and Hunt, 1975); Watanabe and Hedenquist, 2001</td>
</tr>
<tr>
<td>Calcic-potassic</td>
<td>Calcic-silicate-rich veins; more common as massive replacements</td>
<td>Restricted to alkali porphyry Cu deposits, especially those associated with syenite intrusions; calcic-potassic alteration appears to precede potassic alteration</td>
<td>Taranaki (Cox, 1985); Island Copper (Araniehia and Clark, 1996); Kolovos (Chivas, 1978); Panguna (Ford, 1978)</td>
</tr>
</tbody>
</table>

See Table 2 for list of abbreviations.

1 Temperature estimates will vary, depending on factors such as pressure and silica activity (e.g., Hemley et al., 1980; Sverjensky et al., 1991; Watanabe and Hedenquist, 2001)
2 See Einaudi et al. (2003) for more information.
requiring chemical analyses of minerals or rocks. The choice of terms and related semantic issues are addressed in Appendix 1. The summary in Table 3 defines each alteration type, identifies key mineral replacement reactions and estimated formation temperatures partly based on phase petrology, and lists essential and accessory minerals, associated vein types, and examples with references. Although veins are an integral part of alteration and mineralization processes, certain vein types span multiple alteration types; hence, veins are discussed separately below.

The system $K_2O-Al_2O_3-SiO_2-H_2O-KCl-HCl$. The phase diagram of this system is a useful starting point for organizing alteration types (Fig. 5), as the bulk composition of the system is ideal for quartz-bearing aluminosilicate rocks. Figure 5A displays the relative stabilities of K-feldspar, muscovite (proxy for “sericite,” the traditional textural term for fine-grained K-mica), and the aluminosilicate minerals kaolinite, pyrophyllite, and andalusite. The principal phase boundaries in this diagram define the fields for potassic (potassium silicate), sericite (phyllic, quartz-sericite-phyllite), and advanced argillaceous alteration, listed in order of lower $KCl/HC1$ and pH of the fluid. An assemblage that plots in a phase boundary can be defined as transitional between the various types adjacent to the boundary (e.g., K-feldspar and sericite is a transitional potassic-sericite assemblage).

An important restriction on the definition of sericite alteration, which is implicit in the phase diagram of Figure 5A, is that K-feldspar (not just plagioclase), if present, be altered to sericite + quartz. Advanced argillaceous alteration involves formation of kaolinite, pyrophyllite, or andalusite. Sericite and advanced argillaceous alteration chemically represent progressively more intense hydrogen metasomatism and base-cation leaching, which in the extreme results in vuggy silica, though only

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**Fig. 5.** A. Phase diagram for the system $K_2O-Al_2O_3-SiO_2-H_2O-KCl-HCl$ at $P(H_2O) = 1$ kbar, plotting temperature versus $\log (m_{KCl} \cdot k_{HCl})$. Phase equilibria are modified from Henley and Jones (1964), Montoya and Henley (1975), and Henley et al. (1989). B. Phase diagram for the system $K_2O-Al_2O_3-SiO_2-H_2O-KCl-HCl$ at $P(H_2O) = 1$ kbar, plotting $\log (m_{KCl} \cdot k_{HCl})$ versus $\log (a_{K^+}/a_{H^+})$ at 400°C and stabilities of K-feldspar, muscovite, andalusite, and topaz. The diagram is constructed for the fayalite-epidote end member, but the topology of the diagram is little affected by incorporation of effects of solid solution or changes in temperature (e.g., Seedorff and Emard, 2004b). Fields corresponding to alteration types are projected vertically as dashed lines, as labeled at the top. As the diagram illustrates, topaz can be a member of potassic-sericite, or advanced argillaceous alteration types in porphyry systems, depending on the geologic occurrence. Based on data for topaz from data of Barton et al. (1982) and SUPCRT92 database (Johnson et al., 1992). C. Stacked series of phase diagrams that plot $\log (a_{K^+}/a_{H^+})$ versus $\log (a_{K^+}/a_{H^+})$ at various temperatures from 700°C to 150°C. The topology of the diagram changes somewhat with temperature, and invariant points shift to higher values of $\log (a_{K^+}/a_{H^+})$ as temperature declines. Mineral abbreviations are in Table 2. Shaded circles are keyed to alteration types (see text) in appropriate positions in the phase diagrams. Labels for transitional alteration types (e.g., transitional between potassic and sericite) are omitted for clarity. Data are from SUPCRT as reported by Bowers et al. (1994), with additions from SUPCRT92 database (Johnson et al., 1992). Phase boundaries shown at 600°C and 700°C are approximate with respect to $a_{K^+}/a_{H^+}$. Chlorite (chlorite, from Walsh, 1988), cordierite $(X_{Mg} = 0.6)$, and biotite $(X_{Mg} = 0.5)$ activities are given by an ideal, simple solution model for the Mg end member. Positions of eclogite hornblende are estimated from natural phase assemblages and NaK buffered by feldspar.

**Key:** (A) Intermediate Argillic (B) Propylitic
(A) Advanced Argillic (B) Sericitic (C) Potassic
in shallow environments characteristic of high-sulfidation epithermal systems (e.g., Stoffregen, 1987). We use the term greisen only as a textural modifier for coarse-grained aggregates, generally of muscovite, quartz, and other minerals, following the usage of Shaver (1991), Reed (1997), and Seedorff and Einandi (2004a).

Adding F and other anions: In the presence of fluorine-rich fluids with high activity of HF, K-feldspar gives way to topaz, muscovite gives way to topaz, and the aluminum silicate phases andalusite, pyrophyllite, and kaolinite also give way to topaz (Fig. 5B; Seedorff and Einandi, 2004a). In other words, depending on the geologic environment, topaz can be a member of potassic, sericite, or advanced argillaceous alteration types.

The same approach can be used to deal with sulfate, boron, and other anions (Table 3). Addition of sulfate provides for stability of alunite. Quartz + alunite is an important mineral assemblage in advanced argillaceous alteration, which may form by hypogene vapor condensation in the upper levels of some porphyry systems (e.g., Sillitoe, 2000). In boron-rich solutions, K-feldspar and muscovite give way to tourmaline and aluminum silicate phases give way to dumortierite. Hence, potassic alteration has an analogue in the assemblage quartz and tourmaline. Indeed, the spatial position and salinity characteristics of the quartz and tourmaline assemblage in porphyry Sn deposits are similar to those of potassic assemblages in many other porphyry deposits (Grant et al., 1980).

Adding Mg and other cations: The component MgO can be added to model mafic wall rocks or minerals (Fig. 5C). In the advanced argillaceous environment, corresponding to the left-hand side of the diagram at low values of $a_k/a_{Al}$, andalusite, pyrophyllite, or kaolinite are joined by cordierite, chlorite, or biotite at high values of $a_{Mg}/a_{Al}$. Cordierite is favored at high temperatures and chlorite at low temperatures, but the talc field may not be accessible to natural solutions at such low values of $a_k/a_{Al}$. Cordierite occurs as a hydrothermal mineral in the porphyry systems of Green Talc, Nevada (Hudson, 1983), El Salvador, Chile (Gustafson and Quiroga, 1995), and Batu Hijau (PA. Mitchell, pers. comm., 1998). At the intermediate values of $a_k/a_{Al}$, that characterize the sericite environment (Table 3), the principal sericite mineral muscovite can be joined by other minerals such as chlorite and rarely biotite (Fig. 5).

The environment of potassic alteration on the right-hand side of Figure 5 at high values of $a_k/a_{Al}$, and relatively low values of $a_{Mg}/a_{Al}$, can contain K-feldspar, K-feldspar and biotite, or biotite only. At higher values of $a_{Mg}/a_{Al}$, however, cordierite could form at high temperatures and chlorite could form at lower temperatures as part of the potassic alteration assemblages (e.g., Titley and Beane, 1981; Frestoe and Beane, 1982).

For some alteration types, reference to the $a_k/a_{Al}$ scale cannot be the sole criterion for classification. An example is propylitic alteration, where we follow Meyer and Hemley (1967) in reserving the term for weakly metasomatized rocks, in which the main change in chemical composition is addition of volatiles such as CO$_2$, H$_2$O, and/or $s$. Where this criterion of propylitic alteration is met in porphyry deposits (Table 3), propylitic alteration is characterized by (1) presence of relict K-feldspar, if present in the original rock, with albite, calcite, epidote, and a ductile of sericite or illite as alteration products of plagioclase; (2) absence or scarcity of ore minerals; (3) paucity of iron sulfide and oxide minerals, typically including minor pyrite, local hematite, and relict magnetite; and (4) chlorite and/or actinolite with local relict biotite. In Figure 5, propylitic alteration plots along the chlorite-Kfeldspar phase boundary, projected back onto the T-log $a_k/a_{Al}$ diagram, propylitic alteration plots inside the K-feldspar field at moderate to low temperatures (Fig. 5). Care must be exercised in applying the term propylitic where K-feldspar is not present in the wall rocks as an aid in monitoring the chemical environment. For example, altered rocks with abundant chlorite + sericite in mafic protoliths represent a similar log $a_k/a_{Al}$, environment to that of quartz + sericite alteration in felsic protoliths, not propylitic alteration (Fig. 5; e.g., Seedorff et al., 1991).

Intermediate argillic alteration (Meyer and Hemley, 1967) is a weaker and lower temperature form of hydrolytic alteration as compared to sericite alteration, and it forms as sericite alteration wanes (cf. Sillitoe, 1993, 2000; Aracibia and Clark, 1996). A key characteristic is the presence of relict alkali feldspar in K-rich rocks. Clay minerals such as kaolinite, illite-smectite, and smectite are present with illite (sericite) and chlorite (Table 3), but these minerals tend to replace biotite or plagioclase rather than K-feldspar or Na-rich plagioclase. Stability of these clay minerals, with the rarity of dickite and absence of muscovite and pyrophyllite restricts intermediate argillic alteration to temperatures below to ~200°C (Fig. 5).

Silicic alteration, as used here, is a high-temperature, proximal alteration type that occurs in the core of certain porphyry Mo deposits, in which quartz veins exhibit inner silicic and outer potassic envelopes (e.g., Carten et al., 1988b). In the inner envelopes, there is metasomatic addition of silica, with feldspars replaced by quartz and minor feldspar. Coalescence of the inner silicified envelopes in the vicinity of closely spaced quartz veins produces the deposit-scale high-silica alteration zones. High-temperature quartz-tourmaline alteration in porphyry Sn deposits might be regarded as an analog of silicic or transitional silicic-potassic alteration, with tourmaline pyrolyzing for K-feldspar at high boron concentrations in the fluid (see below). Silicic alteration is uncommon in porphyry Cu deposits but is found locally. At Robinson, Nevada, for example, wall rock is replaced over tens of meters mostly by granular quartz with a few percent each of hydrothermal biotite and chloropyrite and may have once had anhydrite (Seedorff, 1993; cf. Westra, 1982). Although the rock contains remarkably few obvious veins, its overall textural appearance is like a giant A vein in the sense of Gustafson and Hunt (1975; see below), and it hosts the highest copper and gold grades in the entire deposit.

A few porphyry Cu deposits, including Ok Tedi, Papua New Guinea, exhibit "quartz cores," which are areas of high densities of quartz veins, also located in the core of orebodies (Watnuff, 1978; Sillitoe and Gappe, 1984; Rush and Seevers, 1990; Zvezdov et al., 1993; Einandi, 1994). In some deposits, these quartz veins lack associated silicic alteration envelopes (i.e., are B veins in the sense of Gustafson and Hunt, 1975; see below) and thus are distinct from the silicic alteration environment described above, but from the descriptions of other deposits we cannot tell if they have silicic alteration.
Other types of silicification occur in porphyry systems. Intense sericitic alteration can include a component of silicification, as at Butte and Resolution (Meyer et al., 1968; Manske and Paul, 2002), where an inner zone of silicification is enclosed in sericitic alteration and locally encloses an innermost zone of advanced argillic alteration. In many other porphyry Cu deposits, inner silicified sericitic envelopes are enclosed in outer envelopes of sericitic alteration (Dilles et al., 2000b). A pyrite-poor form of quartz + sericite alteration of breccia fragments involves silicification in many breccia pipes (e.g., Zwing and Clark, 1995; Skewes et al., 2003). In these settings, magmatic aqueous fluids dominate.

In the shallow parts of many porphyry copper systems and in certain related high-sulfidation epithermal deposits, other forms of silicic alteration are common, if not ubiquitous (e.g., Hedenquist et al., 2000, table 4). One important form is vuggy silica zones enclosed in quartz-albite-pyrite ± pyrophyllite ± dickite assemblages of advanced argillic alteration that characterize shallow, high-sulfidation epithermal deposits (e.g., Stoffregen, 1987; Hedenquist et al., 1994) and relatively shallowly emplaced porphyry copper systems where the near-surface environment is preserved, including Far Southeast, Yerington, and El Salvador (Hedenquist et al., 1998; Lipske and Dilles, 2000; Watanabe and Hedenquist, 2001). These zones may be restricted to steep structures but commonly are blanketlike (e.g., lithocap of Sillitoe, 2000) with roots extending to depth; the zones may be blanked by propylitic alteration and overlie sericite alteration. The vuggy silica zones are produced where a low-density magmatic-hydrothermal gas containing SO2 rises and condenses into shallow ground water to produce extremely low pH, S-rich alteration (e.g., Hedenquist and Lowenstern, 1994).

The additional condition required to define sodic-calcic and sodic-calcic alteration (albitization of Meyer and Henley, 1967) is that there be mineralogic evidence of metasomatic introduction of sodium ± calcium. Sodic-calcic alteration can be regarded as the inverse of potassic alteration (Carten, 1966; Dilles et al., 1995), whereby K-feldspar is replaced by oligoclase, mafic minerals are replaced by actinolite and titanite, or they are destroyed. In contrast, sodic alteration is characterized by lower temperature assemblages involving albite, chlorite, and epidote; in this case, the distinction with propylitic alteration relies on recognition of metasomatic introduction of sodium based on replacement of K-feldspar by albite.

Significant metasomatism of additional components, other than those involving the cations K+, Na+, Ca2+, and SiO2, occurs in certain porphyry deposits. Coupled sodium and iron metasomatism occurred in certain deposits where the assemblage amphibole + plagioclase + magnetite formed (Clark, 1995). Coupled calcium and potassium metasomatism is reported at alkaline porphyry Cu deposits (Lang et al., 1995; Wilson et al., 2003).

Sulfide and oxide minerals: Porphyry deposits contain many sulfide and oxide minerals. Sulfide and oxide phases commonly reequilibrate with decline in temperature (e.g., Barton et al., 1963), but the original sulfide-oxide assemblages commonly can be inferred. Sulfidation state (Barton, 1970), which is a function of both fSO2 and temperature, provides a convenient and consistent framework for systematizing sulfide-oxide mineral assemblages and studying the evolution of hydrothermal systems. Other variables that are less directly constrained, such as total sulfur, also must be considered when evaluating hydrothermal processes. In aqueous solutions, sulfur may occur in the oxidized state as SO42– (n = +6) or in a reduced state as H2S (n = −2). We choose to use a phase diagram that plots the fugacity of sulfur gas (fSO2) versus temperature for organizing sulfide-oxide mineral assemblages from porphyry deposits, although other representations, such as Rg (Rg = log fH2S/fSO2) versus temperature, are equivalent (Einaudi et al., 2003). A characteristic of such diagrams is that the fSO2 covaries directly with fH2S for low temperature ore solutions that are dominated by aqueous H2S, as related by the reaction

\[ S_2 + 2H_2O = 2H_2S + O_2 \]  

In aqueous fluids at high temperature where SO42– may be the dominant sulfur species, fSO2 is inversely proportional to the cube root of fH2S at constant pH as follows:

\[ 2SO_4^{2-} + 4H^+ = S_2 + 3O_2 + 2H_2O. \]  

Because sulfide ore minerals are directly precipitated from reduced sulfur, whereas sulfates such as anhydrite and barite precipitate from oxidized sulfur, total sulfur, sulfidation state, oxidation state, salinity, and acidity are key variables that control ore mineral deposition (e.g., Barton, 1970).

Five categories of sulfidation state can be delimited as bands across the phase diagram, each bounded by a sulfidation reaction of the form A + S2 = B where A and B are sulfur-poor and sulfur-rich minerals, respectively (Fig. 6), as defined by Einaudi et al. (2003). These reactions have a positive slope on the plot of log fSO2 versus 1,000/T (temperature in K), so both sulfur fugacity (fSO2) and temperature vary for any given sulfidation state. A few key reactions define the limits of the various sulfidation environments, ranging from low-
very high sulfidation states, but a network of other sulfidation reactions plot in the diagram (e.g., Barton and Skinner, 1979) and help to locate sulfide and oxide assemblages from porphyry deposits in the sulfidation grid.

Minerals that by themselves define a relatively limited range of sulfidation states include pyrrhotite, covellite, the As minerals loellingite, arsenopyrite, tennantite, and enargite, the Bi minerals native bismuth and bismuthinite, and the Fe content of sphalerite in equilibrium with an Fe sulfide mineral (Barton and Skinner, 1979; Einaudi et al., 2003). More commonly, sulfide mineral assemblages need to be defined in order to constrain sulfidation state. In Figure 6, increasing sulfidation state is shown by assemblages pyrrhotite + loellingite, pyrrhotite + arsenopyrite, pyrite + chalcopyrite, pyrite + bornite or digenite, and covellite. It is essential that assemblages involving Cu-Fe sulfides that lack pyrite (e.g., bornite + magnetite) be distinguished from those in which pyrite co-precipitated with Cu phases (e.g., bornite + pyrite). The late sulfidation of earlier assemblages is common. Indeed, the predominance of chalcopyrite over other Cu-Fe sulfides in certain deposits is due to a widespread late-sulfidation overprint atop sulfide-oxide assemblages containing digenite or bornite (Proffett, 2003a).

Challenges and applications: The stabilities of hydrothermal mineral assemblages define a petrogenetic grid that can provide insight into the nature of ore fluids and the conditions of ore formation (Table 3; Reed, 1997). In the exploration setting, the sulfide ore minerals are commonly absent in the weathering environment but the associated silicate and other gangue alteration minerals commonly remain and serve as guides to ore. The challenge of identifying mineral assemblages in spite of superimposed events can be addressed during mapping and logging (e.g., Proffett, 2003a, b) and then during petrography, using various techniques (e.g., Anderson, 1983; Pennistron-Dorland, 2001). The possibility of postcrystallization changes in composition of silicates must be assessed (e.g., Munoz, 1984) but is less widespread than the nearly ubiquitous reequilibration of Cu-Fe sulfides with decline in temperature (e.g., Barton et al., 1963).

Identification of minor sulfide and oxide phases can have substantial practical implications, regardless of the degree of subsequent reequilibration. The mineralogic residence and textural occurrence of precious metals (e.g., Rubin and Kyle, 1997; Simon et al., 2000) affects whether they report to the concentrate or the tailings. Bismuth minerals commonly report to the Cu concentrate, and smelters penalize high Bi concentrates (e.g., Redwood, 2004), so segregation or blending of ore types may be required if Bi minerals are present. Likewise, As-bearing phases are a processing and environmental concern. For example, enargite generally is present where Cu is deposited in veins with advanced argillic alteration envelopes (e.g., Einaudi, 1995a), although exceptions exist (e.g., Manske and Paul, 2002).

Hyphogene gangue phases warrant attention for both scientific and practical reasons. For example, gypsum veins in the rock and leached cavities in veins could be regarded as clues to the possible former presence of anhydrite. Anhydrite has implications for the sulfur budget of the hydrothermal system, oxidation state of the fluid, and for chemical changes during alteration (e.g., Gustafson, 1979) and consequences in block caving, hydrology, and slope stability (e.g., Hunt, 1985; Kwapil et al., 1989). Similarly, soluble fluorine-bearing gangue minerals may have deleterious effects in tailings facilities, waste rock piles, and solvent extraction-electrowinning (SX-EW) circuits (e.g., Sutter, 2002). Insoluble fluorine-bearing gangue phases that report to a concentrate can have negative consequences at the smelter (e.g., Panguin et al., 1997).

Veins and Breccias

Vein dikes, veins, and breccias form throughout the life of porphyry magmatic-hydrothermal systems, contain a large proportion of the deposited ore minerals, and represent the locus of greatest fluid flux. As a consequence, vein and breccia mineralogy, textures, fluid inclusions, location, abundance, and orientation provide valuable information about the evolution in space and time of fluid composition and flow (e.g., Rehrig and Heidrick, 1972; Titley et al., 1986; Geraghty et al., 1988; Muntean and Einaudi, 2000, 2001; Proffett, 2003a, b). Vein dikes were first described by White et al. (1981) and Shannon et al. (1982). Several distinctive vein types were first described from the pre-Main Stage at Butte by Meyer (1965), Meyer et al. (1968), and Brinkhall (1977), but the primary frame of reference for porphyry deposits is one developed at El Salvador, where Gustafson and Hunt (1975) defined A, B, and D types of veins and Gustafson and Quiroga (1995) expanded the classification to include early biotite (EB) and C veins. Clark (1993) and Arancibia and Clark (1996) added M veins to the lexicon based on work at Island Copper, British Columbia, and Muntean and Einaudi (2000) defined a new type of vein, banded quartz veinlets, in the porphyry Au deposits of the Maricunga belt, Chile. Little of the existing terminology for veins has been used for porphyry Mo, W, or Sn deposits. Reviews of terminology on geisen by Shaver (1991) and on a genetic classification scheme for breccias by SkillHit (1985) also are notable. Veins and vein dikes are illustrated in Figure 7. Characteristics of breccias are summarized in Appendix Table A1, and breccias are illustrated in Appendix 2.

Vein dikes

Vein dikes provide a link between magmatic and hydrothermal processes. As described by Henderson by Carter et al. (1988b), vein dikes commonly contain euhedral crystals of quartz + biotite ± molybdenite that project inward from vein walls into a center occupied by aplite porphyry ± molybdenite (Fig. 71). Occurrences of vein dikes have a close spatial relationship to the occurrence of unidirectional solidification textures in porphyry stocks (described below). With increasing distance from their source stock, vein dikes may grade into open-space veins of coarse-grained quartz ± fluorite.

Classification of veins

Veins can be classified on descriptive criteria including morphology, texture, mineralogy of vein filling and alteration envelope(s), and orientation. We select certain of the most common and distinctive types described from well-studied deposits and assess their abundance in various classes of porphyry deposits. Veins are grouped loosely on the basis of associated wall-rock alteration (including transitional alteration types) and are described in general order of decreasing age in a given location within a deposit. Generally, but not always,
younger veins are formed at lower temperature. Temperature estimates are approximate for many veins and are generally based on mineral assemblages, fluid inclusions, and sulfur isotope fractionation.

**Veins associated with potassic alteration**

**Veins with silicic and potassic envelopes:** Most high-temperature veins in porphyry Mo deposits are quartz veins that differ significantly from the A and B veins (see below) described at El Salvador. At Henderson, Carter et al. (1988b) described both replacement and open-space veinlets. Both types contain quartz, fluorite, molybdenite, lesser K-feldspar, and biotite; pyrite is absent. Both vein types at Henderson have alteration envelopes that vary as a function of distance from stocks. Near the apex of a stock, where the veins tend to be molybdenite poor, they exhibit zoned alteration halos with inner silicic envelopes (quartz + fluorite after K-feldspar and plagioclase) and outer potassic envelopes (K-feldspar after plagioclase). Farther from stocks, where the veins are molybdenite bearing, only potassic halos are present. Veins similar to those at Henderson occur at other rhyolitic porphyry Mo deposits, such as Climax, Colorado; Questa, New Mexico, and Mount Hope, Nevada (Wallace et al., 1968; Cline and Bohnar, 1994; Westra and Riedell, 1996).

The earliest and highest temperature veinlets in porphyry Mo deposits of the Mo-Cu quartz monzonitic to granitic sub-class, such as Hall, Nevada, tend to be barren and locally have silicic envelopes, and secondary K-feldspar, rather than biotite, is the predominant potassic alteration phase. These characteristics are in common with those at Henderson. In contrast with Henderson, the ore-bearing quartz-molybdenite veins commonly contain chalcopyrite and pyrite (e.g., Shaver, 1991).

**Magnete-rich veinlets:** In some shallowly replaced porphyry Cu deposits such as Batu Hijau, magnetite ± Cu-Fe sulfide veinlets are present within biotite alteration (Clode et al., 1999), and they are cut by A-type quartz veinlets (see below). Magnetite-rich veinlets in porphyry Au deposits range from hairline streaks of magnetite ± biotite with minor quartz and chalcopyrite and K-feldspar envelopes to sugary quartz veinlets with magnetite and chalcopyrite and no alteration envelopes, and these have been called A veinlets (Muntean and Ennaudi, 2000, 2001). Porphyry Mo deposits such as Henderson also contain hairline magnetite veinlets that have narrow K-feldspar envelopes, and these formed at lower temperatures than veins with silicic and potassic envelopes (Sedoreff and Ennaudi, 2004a).

**Biotite veinlets:** The earliest, highest temperature veins at many porphyry Cu deposits are biotite veinlets that are related to widespread potassic alteration of wall rocks (Fig. 7A). These veins may be barren or mineralized; they may contain magnetite, sulfides, and other minerals, as at Butte (Brinhall, 1977; Field et al., 2005).

**Green mica veinlets:** Quartz-poor, commonly corundum-bearing, green mica veinlets have been described from a few porphyry Cu deposits. Mineral assemblages include combinations of green biotite, K-feldspar, andalusite, muscovite, cordierite, and corundum. The K-feldspar-andalusite-muscovite assemblage indicates formation at temperatures as high as ~600°C (Fig. 5). Veins of this type include the EB veins at El Salvador (Gustafson and Quiroga, 1995) and green mica veins at Butte (Brinhall, 1977) and Los Pelambres, Chile (Atkinson et al., 1996).

**Quartz-bearing veinlets with biotite and sericite:** Complex veinlets similar to green mica veinlets occur in a few porphyry deposits. Sericite (muscovite), biotite ± K-feldspar, and...
andalusite are present in the vein filling or envelope, but, in contrast with green mica veinlets, quartz is present in the vein filling and corundum is absent. Veinlets of this type include C veins at El Salvador, early dark micaceous veins at Butte (Fig. 7E), and type D veins at Los Pelambres. At Butte, spatial distributions suggest that early dark micaceous veins may be the shallow equivalent of green mica veins (Brinshill, 1977). There are rare reports of these veinlets across multiple classes and subclasses of porphyry deposits.

Sugar pyrite veinlets: Most porphyry Cu and porphyry Au deposits, with the exception of those associated with strongly alkalic host rocks, have sugary quartz veinlets that are broadly contemporaneous with potassic alteration (Fig. 7B). From the work at El Salvador, these are widely known as A veinlets. Gustafson and Hunt (1975) described A veinlets at El Salvador as granular assemblages of quartz (50–90 vol %), perthitic K-feldspar, anhydrite, bornite, and chalcopyrite, and rare biotite that are randomly oriented, discontinuous, and wispy, fringed by envelopes of K-feldspar, anhydrite, chalcopyrite, bornite, apatite, and rutile. Most A veinlets lack center lines or banding and tend to be irregular, discontinuous, and segmented, characteristics that suggest formation under ductile conditions at high temperatures during periods of high-strain rates (e.g., Fournier, 1999). At deeper levels of El Salvador, A veinlets contain andalusite, have lower sulfide contents, and display characteristics transitional to EB veinlets (Gustafson and Quiroga, 1995).

Veins associated with sodic or calcic alteration

Veins dominated by magnetite, amphibole, and plagioclase: Certain tonalitic-granodioritic porphyry Cu-(Au-Mo) deposits, such as Park Premier, Utah (John, 1989b), and Island Copper (Ananich and Clark, 1986) contain veins dominated by magnetite, amphibole, and plagioclase, which generally are the earliest and highest temperature veins in the deposits in which they occur. Certain veins of this type have plagioclase envelopes, whereas others lack alteration halos. These veins were termed M veins by Clark (1993) to emphasize the abundance of magnetite.

Actinolite veinlets: Veinlets of actinolite, or more rarely quartz, plagioclase, and tourmaline, control the distribution of sodic-calcic alteration (400–450°C) associated with oligoclase and actinolite, and are peripheral to potassic alteration in the deep part of the Yerington mine (Carter, 1986). Broadly similar actinolite or epidote veins (Fig. 7C) are observed controlling sodic-calcic alteration in a number of other porphyry Cu deposits (Battles and Barton, 1985; Dilles et al., 1985).

Epidote veinlets: Veinlets of epidote (pyrite-quartz) control the distribution of sodic (± calcic) alteration formed at moderate temperature (400°C or less) in various porphyry Cu deposits. In the Yerington district, these veins have alteration envelopes in which K-feldspar is replaced by albite, biotite is replaced by chlorite, vermiculite and epidote, and magnetite is destroyed (Carter, 1986; Dilles and Einaudi, 1982). At shallower paleodepths in the Yerington system, veins of quartz, pyrite, or tourmaline have albite-chlorite-sericite envelopes.

Quartz veins that lack wall-rock alteration

Banded quartz veins with molybdenite: Many porphyry Cu deposits have quartz veins that lack alteration envelopes, which in the El Salvador terminology are known as B veins (Gustafson and Hunt, 1975). The B type of veins has not been reported from other classes of porphyry deposits. The B veins at El Salvador are continuous planar structures with parallel walls and internal banding, including center lines (Gustafson and Hunt; 1975). They contain coarse-grained quartz, molybdenite, chalcopyrite, anhydrite, minor pyrite, and lesser tourmaline. In general, B veins lack alteration halos; however, with increasing depth at El Salvador, B veins are bordered by envelopes of K-feldspar with occasional albite, biotite, sericite, andalusite, or corundum, thereby blurring the distinction between B veins and other vein types (Gustafson and Quiroga, 1995). Veins sharing characteristics of both A and B types are classified as A-family (Fig. 7D) or AB veins by some workers (e.g., Clode et al., 1999).

Gray banded quartz veinlets: This distinctive type of vein is described from several porphyry Au deposits in the Maricunga belt by Muntean and Einaudi (2000, 2001). The veins lack alteration envelopes and contain dark gray bands whose color is due to abundant vapor-rich fluid inclusions and micrometer-sized grains of magnetite (App. 3A). The bands are commonly botryoidal and continuous through quartz veins, suggesting that quartz crystallized from a silica gel. Banded quartz veinlets also are present at three deposits in the western United States (Muntean and Einaudi, 2000) and in several in northern Peru (Gustafson et al., 2004).

Veins associated with strong hydrolytic alteration

Greisen veins: Though rarely described (Williams and Forrester, 1995), greisen veins occur in porphyry Cu and porphyry Mo deposits in several distinct settings. Greisen veins are distinguished by coarse-grained (diam > 1 mm) white mica. Greisen veins commonly change character over short distances along strike, back and forth, from muscovite + quartz aggregates with no vein filling to podlike vein fillings of quartz or sulfides (especially pyrite or chalcopyrite) with alteration halos of muscovite + quartz. In some deposits, greisen veins containing coarse-grained muscovite and quartz occur instead of, or in addition to, veins with sericitic envelopes (see below).

The Hall porphyry Mo deposit lacks pyritic veins with sericitic envelopes but has two types of veins that contain coarse-grained muscovite (Shaver, 1991). The first consists of veins with coarse-grained muscovite and lesser quartz, molybdenite, pyrite, and chalcopyrite with conspicuous envelopes of secondary K-feldspar. The second contains muscovite and lesser quartz, pyrite, and fluorite in the vein filling, with quartz, muscovite, pyrite, and minor fluorite in the envelope.

Pyritic greisen veins are especially common in porphyry Mo systems of the quartz monzonite to granite Mo-Cu subclass. Turnley Ridge, Montana, has both greisen veins and sericitic alteration related to D veins, but andalusite is locally associated with muscovite and quartz (Steefeld and Atkinson, 1984). Greisen muscovite occurs locally in porphyry Cu deposits in nonpyrite assemblages with bornite ± chalcopyrite (Fig. 7F), such as in the Highland Valley district, British Columbia (Casselman et al., 1995), where it may represent relatively deep levels of the system.

Veins with sericitic envelopes: Through-going veins with sericitic envelopes (Fig. 7G, H) formed at moderate temper-
nature in many classes of porphyry deposits and in the El Salvador terminology are known as D veins. The D veins at El Salvador described by Gustafson and Hunt (1975) are pyrite ± quartz veins that occupy continuous, systematically oriented fractures. Alteration halos contain sericite or sericite + chlorite, as well as pyrite, quartz, anhydrite, other sulfide minerals, and rutile. Fairly systematic changes with depth are observed at El Salvador. At the shallowest levels, bornite and enargite may coexist with pyrite. At progressively deeper levels pyrite commonly is present with chalcopyrite in D veins, then pyrite is the dominant or only sulfide mineral present, then tennantite may be present, and finally both anhydrite and calcite become abundant and tourmaline becomes more common (Gustafson and Hunt, 1975; Gustafson and Quiroga, 1995).

Some definitions of veins with sericitic envelopes, including the original definition of D veins at El Salvador (Gustafson and Hunt, 1975), include minerals characteristic of advanced argillic alteration and high-sulfidation states, suggesting a transition to the base metal lodes of Butte and other districts. A narrower definition could exclude obvious advanced argillic alteration. Similarly, D veins at El Salvador were defined as pyritic veins (Gustafson and Hunt, 1975), but our definition is not restricted with regard to the nature of the vein filling and includes those veins with sericitic envelopes enveloping veins filled with pyrite, quartz, magnetite, specular hematite, and other minerals (e.g., Atkinson et al., 1996; Seedorff and Einaudi, 2004a).

Late, low-temperature, or distal veins

Veins associated with propylitic alteration: Veins containing epidote, pyrite, quartz, chlorite, and calcite are associated with propylitic alteration on the distal edges of many porphyry systems in alteration assemblages with albite, epidote, and chlorite (e.g., MacKenzie, 1970; Gustafson and Hunt, 1975; Koski and Cook, 1982).

Base metal veins: Most porphyry districts contain moderate- to low-temperature veins, lodes, and replacement deposits of base and precious metals. Some are transitional to D veins such as the proximal base metal lodes that have zoned envelopes with an inner selavage of local advanced argillic and strong sericitic alteration, an outer intermediate argillic zone, and a distal propylitic alteration zone (Meyer et al., 1968; App. 3B). More distal lodes and small veins may display sericitic alteration envelopes, but intermediate argillic alteration envelopes are more common and some lack wall-rock alteration. Their sulfidation state ranges from very high to intermediate. These veins are of exploration interest because they commonly characterize the region above and beyond the bulk-tomage orebodies, and they may be exploration targets on their own. Moreover, the mineralogy and metal ratios of the base metal veins correlate fairly closely with the class of the underlying porphyry system (App. Table A2).

Sulfide-poor veins without alteration envelopes: Sulfide-poor veins that commonly lack alteration envelopes form in the final stages of porphyry systems. Carbonate ± silica minerals are common, and prehnite and zeolites may occur in more mafic wall rocks (e.g., Sillitoe and Gappe, 1984; Riedell et al., 1996).

Brecia

Descriptive characteristics: Brecias are fragmental rocks that are as important as veins in a few deposits and occur across all classes and most subclasses of porphyry deposits (Table A1). Brecia bodies exhibit diverse characteristics (Sillitoe, 1985), so detailed recording schemes are required to describe them (e.g., Landwehr et al., 2002). The key characteristics of brecias are (1) the composition of the matrix, (2) the characteristics of the cement, (3) the nature of the fragments, and (4) the geometry and dimensions of the brecia bodies. Table A1 employs a descriptive classification of the six brecia types that are most common in porphyry systems, correlates the descriptive types to the genetic categories of Sillitoe (1985), and assesses the importance and origin of the various types.

Types of brecias: Two types, igneous-cemented brecias and tabular bodies of brecia with angular wall-rock fragments in a matrix of crushed rock (fault brecias), are not distinctive of the porphyry environment and are not related to hypogene ore formation.

Open-space filling, hydrothermally cemented brecias (App. 2) are widespread in porphyry systems and are of special interest because they can be spatially associated with higher ore grades than nearby veinlet-related mineralization. Ore minerals commonly form part of the cement in these brecias (App. 2A, B), although in some cases, the bulk of the ore is contained in crosscutting veinlets that are superimposed on relatively low grade breccia (Zweng and Clark, 1995). The mineral assemblages of the cement and related alteration of fragments vary widely (Sillitoe, 1985), exhibiting much of the mineralogic variability displayed by alteration types and veins (Tables 3, A1), and they form at any time during the evolution of porphyry systems except for perhaps the very latest, lowest temperature stages. The corresponding fluid salinities and interpreted temperatures of formation exhibit wide variations (e.g., Scherkenbach et al., 1985; Cline and Bodnar, 1994; Vargas et al., 1999), similar to deposits that lack significant breccia. Hydrothermally cemented brecias are strongly zoned vertically in some deposits (e.g., Perry, 1961). Certain of these brecias are known to root in porphyry intrusions (e.g., Baldwin et al., 1978; Zweng and Clark, 1995) or silicate-sulfide pegmatite bodies (e.g., Wodzicki, 2001).

Fragmentation of rock is typically related to mechanical (PΔV) energy released by two means: separation of an aqueous fluid from a water-saturated magma, and expansion of already evolved magmatic-hydrothermal fluids that decompress from lithostatic to hydrostatic conditions and undergo liquid-vapor separation (Burnham and Ohimoto, 1980; Burnham, 1985; Fourrier, 1990). High-temperature varieties with hydrothermal biotite and K-feldspar cements are related to release of magmatic fluids (Sillitoe, 1985). Those that are cemented with quartz-tourmaline, quartz-sericite, or chlorite also likely were formed by similar magmatic-hydrothermal fluids from cooling stocks (Sillitoe, 1985). Alternatively, brecciation may be initiated by other mechanisms without introduction of new juvenile fluid. For example, emplacement of a water-poor, unsaturated intrusion into a preexisting hydrothermal system may suddenly add heat to the system,
causing preexisting brine under lithostatic pressure to be heated, inducing rapid expansion, brecciation, and subsequent boiling (Fournier, 1999).

The other three types of breccias (Table A1) contain progressively more juvenile material, which is interpreted as reflecting progressively greater degrees of interaction of magma and external waters. Poorly cemented breccias in rock-water matrix contain variably altered fragments set in a matrix of poorly cemented, sand- to silt-size rock flour. They may lack a juvenile magmatic component where not associated with significant hydrolytic alteration but nonetheless commonly show a close spatial and temporal relationship to late, barren dikes. There are two end-member varieties that differ mainly in their form and degree of mixing and regularity of clasts, and the more common variety is described as pebble breccias (Farinini, 1984; Sillitoe, 1985). Poorly cemented breccias in rock-water matrix are interpreted as forming from indirect interaction of magma and dominantly meteoric water, in which magmatic heat causes water to flash to vapor. The resulting explosion brecciates the surrounding rock and may result in an eruption.

Heterolithic, matrix-supported breccias with a subordinate juvenile component are matrix rich and contain a juvenile tuffaceous component of broken and unbroken phenocrysts (Table A1). Breccia fragments, up to meters in diameter, are dominated by clasts of wall rock but include poorly vesiculated magma and local pumice. These breccias are interpreted as resulting from direct interaction of magma and an external source of water, such as a confined aquifer. The resulting multiple explosions lead not only to brecciation of country rock but also to fragmentation and incorporation of magma. The conduit that contains the breccia, known as a diatreme, represents the roots of maars volcanoes that vent pyroclastic surge and fall deposits.

Clast-rich breccias with predominantly juvenile fragments have a matrix of rock flour, generally with a tuffaceous component (Table A1). Juvenile clasts include not only poorly vesiculated cognate fragments and vesiculated magma but also intrusive rock that is commonly phenocryst poor and flow banded (e.g., Sharp, 1978, 1979). This type of breccia seems to have a closer spatial and temporal relationship to ore deposition than the previous two varieties.

Challenges and applications

Veins are closely linked with hydrothermal walls-rock alteration via the relationship of veinlet filling and associated alteration envelope. A significant problem with high-temperature veins is that they are commonly reopened and act as channels for lower temperature fluids (e.g., Rusk and Reed, 2002; Redmond et al., 2004); hence, genetic ties between vein gangue minerals, ore minerals, and alteration halo can be obscured. For quartz veins, cathodoluminescence imaging is capable of revealing internal crystal growth and dissolution zones that may further understanding (cf. Penniston-Dorland, 2001).

In a given deposit, many vein types encompass multiple mineral assemblages, as evidenced by the routine presence of FeNi sulfide minerals in a vein type (e.g., Gustafson and Quiroga, 1995). The specification of a vein type narrows the possible physiochemical environment of formation, but such limits remain broadly compared to the environment of a single mineral assemblage. Mapping the distribution of vein types and relict sulfides encapsulated in quartz can be a useful method of exploration targeting and grade prediction (Gustafson and Hunt, 1975; Titey et al., 1986; Muntean and Einaudi, 2001; Proffitt, 2003a).

Hydrofracturing of rock and hydrothermal fluid flow produce both veins and breccias and link these features. The gravitational and interrelated nature of intrusive, extrusive, and hydrothermal processes can be deciphered from a descriptive classification and genetic understanding of breccias. The clasts in breccias can provide samples of deeper parts of a system than is exposed at the surface. Certain of the open-space filling, hydrothermally cemented breccias represent high-grade exploration targets, but other breccia types destroy or dilute the grade of earlier mineralization (Table A1).

Relative Ages of Hydrothermal Alteration and Mineralization

Early workers on porphyry deposits recorded temporal relationships, typically summarized in terms of a paragenetic diagram that plotted individual minerals on the vertical axis and time on the horizontal axis. Careful consideration of relative ages of events in porphyry systems, in the context of mineral assemblages and vein types, is a relatively recent advance. Wallace et al. (1968) used intrusive contacts to establish relative ages of alteration and mineralization events of the Climax porphyry Mo deposit. They also described porphyry intrusions being emplaced as the hydrothermal system evolved—the intermineral porphyry intrusions of Kirkham (1971). Gustafson and Hunt (1975) emphasized the importance of time in all aspects of hydrothermal alteration and mineralization.

As a result of the advances in understanding relative ages of veins and intrusions, a new form of the conventional paragenetic diagram has been created, called a space-time diagram. In this diagram, veins and hydrothermal mineral assemblages are arrayed along one axis of the diagram as a function of position along a profile through the deposit; the other axis represents geologic time, using relative ages calibrated using the sequence of intrusive events (e.g., Seedorff, 1988; Proffitt, 2003a).

Crosscutting relationships

Crosscutting relationships between veins: The most reliable basis for determining relative ages of hydrothermal events is the documentation of a vein cutting and offsetting another vein (Fig. 7). K. Vein intersections without offset are commonly not definitive because of potentially deceiving exposures, such as cases in which minerals are selectively deposited where younger veins cross older veins and where earlier veins are reopened and utilized by younger fluid (cf. Fig. 7L and discussion in Meyer and Hemley, 1967). Representative published photographs of vein-vein crosscutting relationships include those for El Salvador (Gustafson and Hunt, 1975), Henderson (Carter et al., 1988b; Seedorff and Einaudi, 2004a), Cuvachona deposit at La Pampa, Chile (Muntean and Einaudi, 2001), and Goonumbla, New South Wales (Lickfield et al., 2003). A crosscutting relationship between veins establishes the relative ages of the two events at a single point in space. Certain age relationships might be
observed repeatedly in space through a porphyry system, such as pyritic veins cutting and offsetting quartz veins; however, there may be locations in a system where the reverse occurs. Such reversals are potentially critical for understanding the evolution of a hydrothermal system (see below).

**Crosscutting relationships between veins and porphyry intrusions:** This type of crosscutting relationship (App. 3C) potentially allows the correlation of ages of individual hydrothermal features in space because most porphyry intrusive contacts formed in a short interval compared to the duration of the hydrothermal system. It also is one of the best ways to limit the absolute age of formation of a vein by tying it to the age of emplacement of a well-dated intrusion (e.g., Ballard et al., 2001). Representative examples of published photographs of porphyry-vein crosscutting relationships include El Salvador (Gustafson and Hunt, 1975), Mount Pleasant, New Brunswick (Kooiman et al., 1986), Fine Grove, Utah (Keith et al., 1986), Bingham (Redmond et al., 2001), and Goonubula (Lickford et al., 2003). Relative ages between porphyries and veins can be used to classify intrusions as premineral if they are cut by all types of veins, intraminal (or intermineral) if they cut off some mineralized veins and are cut by other mineralized veins, and postmineral if they cut off all veins.

**Challenges and applications:** The relative ages of veins and intrusions are still poorly documented at most deposits. Intraminal or postmineral porphyries that cut off mineralized veins are critical to grade control during mining and identification of their shapes and locations improves estimates of mineral resources. During exploration, failure to identify such intrusive contacts may result in a prospect being undervalued or overvalued, depending on the location of existing drill holes. Drill holes in diatremes or late barren intrusions may be only slightly offset from the highest grade part of an orebody; conversely, unsuspected barren intrusions near a high-grade intercept may dilute or completely cut out large volumes of potential ore. One practical implication of recognizing reversals in the relative ages of veins is that they indicate that another mineralizing stock or intrusive center is nearby.

**Igneous Rocks and Magmatic Systems**

The importance of igneous rocks in hydrothermal ore deposits was recognized long before Lindgren (1933) published his classification scheme (e.g., Spurr, 1923). Compositions of porphyries and associated minerals have been studied for differing reasons, ranging from (1) distinguishing potentially barren individual intrusions and intrusive suites from productive and high-grade intrusions and suites (e.g., Stringham, 1966; Hendry et al., 1985, 1988), (2) determining the sources of metals on a crustal scale (e.g., Lang and Titley, 1985), (3) inferring plate tectonic environments (e.g., Kesler et al., 1975; Keith and Swan, 1996), (4) linking metal contents of deposits to compositions of associated igneous rocks (e.g., Westra and Keith, 1981; Carter et al., 1993), and (5) understanding the evolution of volatiles and metals from magma chambers (e.g., Mutscherl et al., 1981; Hattori and Keith, 2001).

**Textures and form**

**Volcanic rocks:** Volcanic rocks in porphyry deposits may be older and unrelated to the mineralized magmatic system. In other cases, they are broadly related to the same magmatic system, generally predating the mineralizing porphyries and occurring as wall rocks to the porphyry intrusions or occurring as late postmineral deposits (e.g., Dilles and Wright, 1988; Harris et al., 2004).

**Hypabyssal porphyry intrusions:** Virtually all deposits have exposures of porphyry, a strongly porphyritic hypabyssal intrusion. The characteristic fine-grained aplitic (sugary) groundmass texture (Fig. 7M, N) is caused by pressure quenching of the porphyry upon rapid ascent and loss of volatiles from the magma (Fournier, 1967; Jahns and Burnham, 1969). Although some intrusions in porphyry Cu systems exhibit wide textural variation, the ore-forming intrusions are relatively texturally homogeneous and have abundant phenocrysts (ranging from 35 to 55 vol %), including plagioclase phenocrysts ranging from 1 to 3 mm in length; the typical grain size of the groundmass is 0.02 to 0.3 mm. In other types of porphyry deposits, such as Henderson (Carter et al., 1986b), stocks commonly are texturally zoned and the range in phenocryst contents is greater and may vary widely within a single intrusion.

In districts containing exposures below the level of orebodies, the groundmass grain size coarsens with depth (cf. App. 3D), such that the texture eventually becomes seriate, then porphyritic with a hypidiomorphic-granular matrix, and finally hypidiomorphic-granular or granitic, as at Yerington, El Abra, Chile, and Henderson (Ambrus, 1977; Carter, 1996; Carter et al., 1986b). Although the individual hypabyssal porphyry intrusions are distinct bodies at the level of the ore deposit, each coincident with a discrete fluid release event, they likely continue downward and merge into the same underlying magma chamber (e.g., Dilles, 1987).

The forms of these hypabyssal intrusions vary greatly (e.g., Sutherland Brown, 1976b), from a few isolated dikes representing as little as 1 vol percent of ore-bearing rocks, as at Butte (Meyer et al., 1968), to dikhe swarms, as at Muñong (Basa), Philippines, Island Copper, and Yerington (Sillitoe and Gappe, 1984; Dilles and Proffenst, 1985; Ferrell et al., 1995). The intrusions also can be cylindrical stocks, as at Berg, British Columbia, Climax, El Salvador, Panguna, Papua New Guinea, and Santo Tomas II, Philippines (Wallace et al., 1989; Gustafson and Hunt, 1975; Panteleyev et al., 1976; Serafica and Balada, 1977; Clark, 1990), and broad, domal intrusions, as at Mount Tolman, Washington (Lassman and Utterback, 1995). The number of recognized hypabyssal porphyry intrusions in a deposit ranges from two to three at Chuquicamata, Chile, to six to seven at Bingham, Yerington mine, and El Salvador, to more than 14 at Henderson. Aggregate cross-sectional areas of porphyry Cu intrusions temporally linked to ore ranges from 0.2 (Batu Hijau, Butte, Yerington mine) to 0.5 km² (Tocquepala, Perú, El Salvador) and do not correlate with the mass of contained metal.

**Plutonic rocks:** Plutonic rocks that host porphyry deposits may be older and unrelated to the mineralized system, such as the Proterozoic granites in southwestern North America (Titley, 1982e) and the batholiths only a few millions of years older than the porphyries at Butte, Santo Tomas II, and Tanamá, Puerto Rico (Sillitoe and Gappe, 1984; Cox, 1985; Lund et al., 2002). Plutons more closely linked in time to the porphyry system also can host porphyry deposits. Early
hypidiomorphic granular plutons, ranging from large bodies to small stocks, are wall rocks to younger porphyry dikes and are important host rocks at Bingham, El Abra, Goomalby, Highland Valley, Los Pelambres, and Yerington (Bray, 1969; Angrisani, 1977; McMillan, 1985; Dilles and Proffett, 1965; Atkinson et al., 1990; Lickford et al., 2003). More commonly, broadly co-genetic plutonic rocks are only present or inferred at depth, hidden below the porphyry intrusions.

Unidirectional solidification textures: The term unidirectional solidification texture (UST) refers to minerals that define a growth direction from a solid substrate (Shannon et al., 1982), and they are found in diverse types of magmatic-hydrothermal deposits. Some of the most spectacular examples are from porphyry Mo deposits (White et al., 1981; Shannon et al., 1982; Carten et al., 1988a, b), but they also occur in many porphyry Cu deposits (see compilation of Kirkham and Sinclair, 1988).

Unidirectional solidification textures occur in the apical portions of porphyry stocks. Crystals grow inward from the roof and walls of a magma chamber, forming creulite layers (Shannon et al., 1982). One of the most common types of creulite layers, forming in the roof zones of intrusions, has rhythmic, monomineralic layers of euhedral quartz interlayered with apatite, apatite porphyry, or apatite rhyolite (Fig. 70). Other varieties include pegmatoidal and less common dendritic layers of quartz and/or K-feldspar (App. 3E).

Prominent horizons of unidirectional solidification textures occur in bands at the boundaries between internal textural subdivisions of individual stocks at Henderson (Carten et al., 1988a, b). Here, orebodies persist downward only as far as the lowest occurrences of unidirectional solidification texture horizons, and individual vein dikes and open-space veins (described above) have been mapped as originating in individual creulite layers. In a few deposits, the creulite bands may also contain ore minerals and may constitute ore-grade rock, but in other deposits creulite layers may be best developed in the least mineralized intrusions (Carten et al., 1988a; Kirkham and Sinclair, 1988). Lowenstern and Sinclair (1996) demonstrated that the quartz in creulite layers is compositionally distinct from quartz that occurs as phenocrysts and groundmass grains in the interlayers. They concluded that interlayer quartz is magmatic but that creulite quartz is hydrothermal. Hence, these layers texturally record the transition between magmatic and hydrothermal conditions and demonstrate that aqueous fluids locally accumulated in the apex of a porphyry stock during crystallization (e.g., Carten et al., 1988a, b; Kirkham and Sinclair, 1988; Lowenstern and Sinclair, 1996).

Compositions

The compositions of rocks petrologically related to porphyry systems span virtually the complete range of modern volcanic rocks, occupying all fields in a plot of total alkalis versus silica (Fig. 8A). The porphyry intrusions most closely linked in time and space with ore, however, exhibit a more restricted range of SiO₂ contents of ~55 to 75 wt percent. The silicic rocks associated with certain rhyolitic porphyry Mo deposits are highly fractionated, as evidenced by their extremely enriched or depleted trace element contents (Carten et al., 1988a, b; Keith et al., 1993), but the compositions of most porphyry intrusions are not exceptional (e.g., Lang and Titley, 1998).

Rocks broadly associated with porphyry systems include sparse mafic and ultramafic compositions (Fig. 8A). Diorites and pyroxenites occur in certain porphyry Cu deposits associated with mafic and alkaline rocks, such as Kolombia, Solomon Islands, and Lorraine, British Columbia (Chivas, 1978; Lang et al., 1995).

There are minor occurrences of lamprophyres, which are a diverse group of volatile-rich ultramafic rocks characterized by phenocrysts of biotite and lesser clinopyroxene with plagioclase only in the groundmass (e.g., Mitchell, 1994). Kersantite occurs as small dikes in many porphyry Mo deposits with evidence of having conformed with leucocratic rocks (Bookstrom et al., 1998; Keith et al., 1998). Lamprophyres also are present in some porphyry Cu systems. Minette conformed with silicic magma at Bingham and may have contributed sulfur and chalcopyrite elements to the Bingham system (Keith et al., 1998; Hattori and Keith, 2001; Stavest et al., in prep.). Contributions from ultramafic magmas could account for enriched platinum-group element (PGE) concentrations in deposits such as Bingham and Skouries, Greece (Tarkian and Stibrivs, 1999).

A plot of CaO/(Na₂O + K₂O + CaO) versus Al₂O₃/(Na₂O + K₂O + CaO) for least altered rocks permits discrimination on the basis of alumina saturation (Fig. 8B). There is wide scatter of data, but general trends are visible. The majority of porphyries are metaluminous or weakly peraluminous. A few rocks from syenitic porphyry Cu-(Au) systems plot in the peralkaline field. The porphyry Sn and W and more than half of the Mo deposits tend to have weakly to strongly peraluminous compositions.

Analyses of rocks from porphyry Cu-(Au-Mo) deposits plot in an arc of high normative plagioclase and quartz toward the upper right-hand part of a normative Streckeisen diagram (Fig. 8C). Data for porphyry Cu-(Mo) deposits plot along another arc near the middle of the diagram. Data from alkaline deposits plot with low normative quartz and high orthoclase in the lower part of the diagram (a few quartz-saturated points would plot below the diagram—not shown). Porphyry deposits containing Mo, W, and Sn are associated with silicic rocks that plot near the granite minimum (Q = Or = Ab + An, Fig. 8C).

Mineralogy

Porphyry intrusions associated with most porphyry deposits contain phenocrysts of plagioclase; hornblende is common in those of intermediate composition; and biotite, K-feldspar, and quartz are common in those with more silicic compositions. For porphyries containing all the above phenocrysts, the temperature and water content of the magma at emplacement can be estimated from experiments at ~675°C to 700°C and ~4 wt percent, respectively (Nanney, 1983; Dilles, 1987). More mafic suites associated with porphyry Cu-(Au-Mo) and porphyry Au deposits may contain clinopyroxene and rarely orthopyroxene, as at Kolombia (Chivas, 1978). Rocks in alkaline suites also commonly have clinopyroxene phenocrysts, whereas silica-saturated syenitic suites may contain pseudoleucite and melanite garnet (e.g., Lang et al., 1995; Wilson et al., 2003). The peraluminous rhyodacites
related to the Lallagua porphyry Sn deposit, Bolivia, contain cordierite (Grant et al., 1980).

The most common primary accessory phases are magnetite, ilmenite, titanite, apatite, and zircon. Rhyolitic porphyry Mo deposits such as Henderson and Climax may also contain microphenocrysts of fluorite, ilmenorutile, monazite, and xenotime, and other trace minerals (Gunow, 1983; Carten et al., 1985a). Anhydrite has been reported as a microphenocryst in a few districts, such as Goonumbla (Lickford et al., 2003).

Compositions of coexisting phenocrysts constrain magmatic variables such as oxidation and sulfidation state and halogen fugacities (Einnaudi, 1982b; Munoz, 1984; Scaillet et al., 1998; Einnaudi et al., 2001). The common occurrence of magnetite and titanite, the local presence of anhydrite, and the compositions of igneous minerals such as biotite and apatite testify to the relatively oxidized nature of the magmas associated with most porphyry Cu deposits (Einnaudi et al., 1981; Dilles, 1987; Wones, 1989; Streck and Dilles, 1998). Similarly, certain porphyry Mo magmas are relatively oxidized (Keith and Shanks, 1985). In contrast, porphyry W and porphyry Sn systems may be relatively reduced (e.g., Burnham and Ohimoto, 1980).

Open-system behavior

Most magmatic chambers behave as open systems over time, periodically being replenished or underplated at the base, interacting chemically with their enclosing rocks and discharging magmas and volatiles out the top (e.g., Hildreth, 1981; Pallister et al., 1992; Eichelberger et al., 2000). Underplating of silicic magma chambers by mafic magmas is common—if not ubiquitous—and mafic plutons occur in the roots of batholiths and at the base of magmatic arcs (DeBarri and Coleman, 1989; Wiebe, 1996). The evidence noted above of commingled silicic and mafic magmas in porphyry systems (cf. Halter et al., 2004) suggests that injection of mafic magma into the base of the silicic chamber occurs in porphyry systems and may have occurred close to the time of porphyry ore formation for certain deposits (e.g., Hattori and Keith, 2001). Geologic relationships indicate that several mineralizing intrusions did not erupt magma at the surface and theoretical considerations strongly suggest that others did not erupt (e.g., Carten et al., 1986b; Dilles et al., 2000a; Proffett, 2003a). Viewed over a longer time period, however, it is equally certain that the magma chambers associated with many porphyry systems did erupt (e.g., Keith et al., 1986; Keith et al., 1993). Indeed, the silicic chamber at Red Mountain, Colorado, almost certainly vented between formation of the small Urad orebody and the giant Henderson deposit (Wallace et al., 1978; Stein and Crock, 1990; Seedorff and Einnaudi, 2004a).

Challenges and applications

The biggest challenge in understanding the petrology of porphyry systems is obtaining rocks and minerals that are
sufficiently fresh for meaningful interpretation of the analyses. Loss of volatiles, including ore fluids, from porphyry magmas during crystallization removes Cl, S, alkalis, and many trace metals. In some cases, fluid, melt, and sulfide inclusions are trapped in igneous minerals and provide information about the original magmatic volatile, metal, and sulfur budget. For example, the Cu/Au ratios of magmatic sulfide and bulk ore are similar at Bajo de la Alumbrera, Argentina (Halter et al., 2002). Furthermore, the role of mafic mantle magmas versus silicic crustal magmas in providing key ore components is commonly difficult to assess because magma of silicic or a mixed intermediate composition caps the magma chamber and provides most of the available samples. There is broad consensus that the Cu, Au, and PGE are derived from mafic melts, whereas Mo, W, and Sn are crustal, but the sources of water, chlorine, and sulfur are debated (e.g., Keith et al., 1993; Streck and Dilles, 1998).

Chemical compositions of igneous rocks can be used cautiously as a guide to distinguishing different classes and subclasses of porphyry deposits. This can be of exploration interest, because some subclasses such as quartz monzonitic-granitic porphyry Mo-Cu have substantially lower odds of generating an economic deposit than other subclasses such as quartz monzodioritic-granitic porphyry Cu-(Mo) and monzonitic porphyry Cu-(Mo-Au).

Spatial Distribution of Igneous and Hydrothermal Features

The deposit- and district-scale studies of porphyry systems that originated at about the time of the founding of Economic Geology focused on the spatial distribution of igneous and hydrothermal features. Early studies were completed mainly by field parties of the U.S. Geological Survey, mostly in the western United States but also abroad. Field-based studies complemented with laboratory analyses continued into the Society's second- and third-quarter centuries but have been less common for the last-quarter century, although such field work is still an essential part of future study.

Deposit- and district-scale descriptions led to descriptive models of the spatial distribution of igneous and hydrothermal features (e.g., Jerome, 1966; Lowell and Guillbert, 1970). Over time, the models incorporated more of the diversity of porphyry systems, in types of contained metals, compositions of porphyries, and wall-rock composition (e.g., Guilbert and Lowell, 1974; Sillitoe et al., 1975; Vila and Sillitoe, 1991; Zvezdov et al., 1993). The origin of the shapes of orebodies was studied in the former Soviet Union (e.g., Krivtsov et al., 1981) but has received less attention elsewhere (e.g., Seraphin and Hollister, 1976).

Descriptive elements

Distribution of igneous rocks: The broad spatial context of the porphyry system begins at the paleosurface, extends for kilometers on either side of a deposit, and continues downward for >10 km into and beneath the magma chamber (cupolas in Fig. 9) from which porphyry intrusions and related aqueous fluids may have ascended. The relative positions of successive intrusions have a large effect on the geometry and characteristics of a system, irrespective of the shapes of the porphyry intrusions (see above).

At some deposits, successive porphyry intrusions are emplaced at successively greater depths, more or less one under the other, such as at Climax (Wallace et al., 1988; White et al., 1981) and Pine Grove (Keith et al., 1986). In others, successive intrusions are emplaced at the same time but are derived from successively greater depths, as at the Yerington mine and the Ann-Mason deposit (Carter, 1956; Dilles, 1987). Intrusions may also be emplaced side by side, in some cases along a trend, as the intrusive centers at the Henderson (Carter et al., 1985b) and the porphyries at Mount Hope (Westra and Riedell, 1996) and Bajo de la Alumbrera (Proffett, 2003a, b). In another observed geometry, successively emplaced cylindrical intrusions may be nested inside the interiors of older intrusions, as at Santo Tomas II, Henderson, Panguna, and Goonumbla (Serafica and Baluda, 1977; Carter et al., 1985b; Clark, 1990; Lieckfold et al., 2003), and, with less regularity, El Salvador (Gustafson and Hunt, 1975). Where the late intrusions are barren as at Bajo de la Alumbrera, they may produce a barren core or accentuate an inward decline in grade (Sillitoe, 2000; Proffett, 2003a).

Distribution of grade and metal ratios: Metal distribution patterns can be described at different scales. There is more than one mineralized intrusive center at the district scale where orebodies occur in clusters related to a batholith (e.g., Titley, 1982c). Individual mines or orebodies may be a composite of several ore-forming intrusive events, as at Henderson (Carter et al., 1985b).

The determination of ore grades and metal ratios is useful for posing genetic hypotheses (Langton and Williams, 1982; Wilson et al., 2003; Seedorff and Imaudi, 2004b), but careful investigation is required to unravel the source of various additions (and subtractions) of metals. The factors include the superposition of multiple vein sets, local-scale hypogene sulfidation and redistribution, system-scale hypogene enrichment by an overprint of advanced argillic mineralization, widespread hypogene depletion by sodic-calcare alteration, and supergene enrichment and depletion (e.g., Carter et al., 1985b; Dilles and Imaudi, 1992; Titley and Marozas, 1995; Proffett, 2003a; Sillitoe, 2005). There are few deposits for which there is published documentation of the distribution of metal grades. The data are especially sparse for by-product and trace elements, so metal ratios and metal zoning are more poorly known.

Distribution of veins and minerals: The abundance of various veins, breccias, and hydrothermal minerals can be mapped or logged, and then contoured to illustrate the geologic controls on distribution (e.g., Titley et al., 1986; Carter et al., 1985b; Proffett, 2003a, b); many orebodies have been found by making and using such maps.

Deposition of the distribution of hydrothermal mineral assemblages, rather than outlines of alteration types and sulfide zones, is a greater challenge to convey clearly, as there may be a dozen or more assemblages and a considerable overlap in their distributions. Where semiquantitative estimates of the abundances of each mineral assemblage have been mapped, they can be contoured like assays. Geologic controls on distribution can be mapped where, for example, older assemblages are cut off by younger intrusions (e.g., Seedorff and Imaudi, 2004a).
FIG. 9. Generalized system-scale geologic maps and vertical profiles through selected porphyry systems, all at a common scale as defined by bar at left, selected for degree of study and exposure, coverage of various metal types, and variations on a theme. Vertical profiles include simplified rock type, alteration type, economic ore metals, and sulfide-oxide minerals, and corresponding plan maps illustrate geology and wall-rock alteration of key porphyry districts: 1 = Refugio Au, Maricunga district, Chile (Muntau and Einaudi, 2000, 2001); 2 = Batu Hijau Cu-(Au-Mo), Indonesia (Clode et al., 1999; Garwin, 2002); 3 = El Salvador Cu-(Mo), Chile (Gustafson and Hunt, 1975; Gustafson and Quiruga, 1986; Gustafson et al., 2001; Watanabe and Hedenquist, 2001); 4 = El Teniente Cu-(Mo), Chile (Carnes, 1975; Skewes et al., 2002); 5 = Resolution (Magma Porphyry), Superior district, Arizona (Manise and Paul, 2002); 6 = Butte Cu-(Mo), Montana (Meyer et al., 1986; Proffett, 1973; Houston, 2001; Field et al., 2005); 7 = Yerington Cu-(Mo), Nevada (Proffett, 1979; Proffett and Dilles, 1984; Carten, 1986); 8 = Anna-Mason Cu-(Mo), Yerington, Nevada (Dilles and Einaudi, 1992; Dilles and Proffett, 1996; Dilles et al., 2000b; Lipska and Dilles, 2000); 9 = Bajo de la Alumbrera (Unich et al., 2001; Proffett, 2003a, b); 10 = Bingham Cu-(Mo-Au), Utah (Babcock et al., 1995; Ballantyne et al., 1997; Phillips et al., 1997; Redmond et al., 2001); 11 = Hall Mo-Cu, Nevada (Shaver, 1996; 1991); 12 = Henderson Mo, Colorado (Seedorff and Einaudi, 2004a, b); 13 = Chorolque Sn Bolivia (Grant et al., 1980), plus observations of the authors. Columns and maps are highly generalized but to scale and are intended to show the key features of each deposit. Vertical position is approximately estimated, as indicated. In the vertical profiles, left-most columns illustrating hydrothermal alteration are arranged in general order of decreasing age from oldest on left to youngest on right. The rightmost columns, illustrating sulfide-oxide minerals, are in general order of decreasing age from oldest on left to youngest on right.
System-scale observations

Depth and level of exposure: The paleosurface is a useful frame of reference, although it is generally not flat and its position may change during the life of a hydrothermal system (e.g., Sillitoe, 1994). The level of the paleosurface can be established within close limits at a few deposits (Fig. 9). For certain upright and gently tilted deposits, the paleosurface can be projected over the top of the orebody, as at Bajo de la Alumbrera (Proffett, 2003a), Bingham (Einaudi, 1982b; Keith et al., 1995), Marte, Chile (Vila et al., 1991), and Pine Grove (Keith et al., 1986).

Depths of emplacement also are relevant to cupolas, which are the domal tops of evolving magma chambers that crystallize as equigranular to porphyritic granites (Emmons, 1927). Several cupolas may develop during the evolution of a single batholith as described above. Porphyry stocks and dikes commonly emanate from cupolas, cutting up through the crystallized roof of the cupola toward the surface (Dilles et al., 2000a).

Well-studied regions (e.g., southwestern North America) show that porphyry deposits are restricted to areas where the upper crust is preserved (e.g., Butler, 1915; Emmons, 1927; Barton, 1996). Systems that were faulted and/or tilted after they formed could display multiple levels or a wide range in levels of exposure (see below).

Examples from well-studied deposits: Figure 9 shows a compilation of igneous and hydrothermal features from well-studied systems representing a variety of classes and subclasses of porphyry deposits. Data are presented as plan maps and vertical columns referenced to the paleosurface through the center of the system, prior to any postmineral tilting or dismemberment. As the figure indicates, most porphyry deposits form within 1 to 6 km of the surface but deposits such as Butte developed as deep as 9 km (Roberts, 1975; Rusk, 2003). The tops of cupolas range from 3 to >10 km below the surface (Fig. 9). The figure also shows simplified distributions of rock types, wall-rock alteration, sulfide and oxide assemblages, locations of related deposits, and metal grades.

Postmineral deformation and importance of tilted systems

Exposure in relatively intact, upright systems: The total direct vertical exposure (sum of topography, man-made openings, and drilling) rarely exceeds 2 km. Bingham, Chuquicamata, Grasberg, El Salvador, and El Teniente are deposits with some of the greatest exposure in the vertical dimension. Compact systems such as porphyry Sn, W, and most porphyry Mo deposits (Fig. 9) require a smaller absolute exposure to obtain a comparable degree of understanding of a large porphyry Cu system. El Abra, Bingham, and Butte are examples of huge systems with extensive lateral exposure (Ambrus, 1977; Babcock et al., 1995; Houston, 2001).

Relatively upright systems exposed at several levels by normal faulting: Normal faults cut out structural section, with deeper structural levels exposed in the footwall block. Markedly different levels of the igneous and hydrothermal system potentially can be juxtaposed from fault block to fault block. Nonetheless, the system may be relatively unaltered by postmineral deformation. Hence, the plunges of intrusions are little different from their plunges at the time of emplacement, which is generally within 10° to 20° of vertical. Examples include the Christmas, Arizona, Peschanka, Russia, and Butte deposits (Einaudi, 1982a; Zvezdov et al., 1993; Houston, 2001).

Systems strongly dismembered and tilted by normal faults: Some porphyry deposits are located in areas that have been subjected to considerable crustal extension (100% is not uncommon) since the porphyry deposits were formed. In areas with large amounts of crustal extension, there are multiple sets of normal faults (e.g., Proffett, 1977; Gans et al., 1995). Faults commonly were initiated at high angles and were rotated to lower angles as they moved and as younger, steeper faults initiated, causing the fault blocks that they bound to be tilted.

The most complete three-dimensional view of alteration and mineralization in a porphyry system is available at the Yerington district, Nevada (Fig. 9), which has a cluster of Jurassic-age porphyry centers with associated Cu-Mo mineralization. A sequence of Tertiary volcanic rocks straddles the period of extension, constraining the age of extension and aiding restoration of faults to generate pretilt views of the porphyry systems (Proffett, 1977; Dilles and Gans, 1995). Postmineral deformation tilted the system about 90° westward. Porphyry centers, their cogenetic batholith, and Cu skarns hosted in adjacent Mesozoic sedimentary and volcanic rocks are exposed at the present surface as a Jurassic cross section, with top to the west (Einaudi, 1977; 2000; Proffett and Dilles, 1984; Dilles, 1987). The batholith exhibits multiple cupolas across the district (Dilles and Proffett, 1995; Dilles et al., 2000a). In the single major fault block that contains the Ann-Mason deposit, the present surface provides virtually continuous exposure of ~6 km of paleodepth (Dilles and Einaudi, 1992). In each of several fault slices, the lateral exposure is several kilometers, and drill holes of a kilometer or more provide information in the perpendicular pretilt lateral dimension (Proffett and Dilles, 1984). The Yerington system also includes exposures of sodic-calcic alteration at the sides of the system and advanced argillic alteration at the top of the system (Hudson, 1983; Lipske and Dilles, 2000). Normal faults also have provided oblique or cross-sectional views through many other porphyry systems in the North American Cordillera (e.g., Seedof, 1991a; Wilkins and Heidrick, 1995). Other significantly extended and tilted deposits include Ajo, Arizona (Hagstrom et al., 1987); Cuatro Hermos, Sonora (Gans, 1997; Zurcher, 2002); Endako, British Columbia (Selby et al., 2000; Lowe et al., 2001); Globe-Miami and Ray, Arizona (Wilkins and Heidrick, 1995); Hall (Shaver and McWilliams, 1987), Robinson (Seedorff et al., 1996; Gans et al., 2001), Royston, Nevada (Seedorff, 1991b), and San Manuel-Kalamauro, Arizona (Lowell, 1968). Most of these deposits are overlain by postmineral rocks that are tilted at least 50° and contain <3 km of paleovertical extent.

Systems dismembered by reverse faults: Reverse and thrust faults disrupt structural section via relative uplift of the hanging-wall blocks. Rotation of beds is localized, mainly in the hanging walls of folds and in fault-bend folds (e.g., Suppe, 1983); hence, beds in both the hanging wall and footwall may not be rotated. Because tiling is less compared to extensional settings and duplication is common, reverse faults rarely produce large vertical exposure of a porphyry system.
Systems that are dismembered by reverse faults include Cadia (Holliday et al., 2002), the Kerr porphyry deposit and the greater Sulphurets district, British Columbia (Kirkham and Margolis, 1995; Bridge et al., 1996), and Potrerillos, Chile (Olson, 1989; Marsh et al., 1997).

**Systems penetratively deformed by contractional deformation:** Certain systems, such as Gibraltar, British Columbia (Bysouth et al., 1995), have been deformed in the ductile regime and have been folded and metamorphosed but retain enough of their characteristics to be clearly recognizable. Other deposits, such as Aitik, Sweden (Wanhainen et al., 2003), exhibit sufficiently intense ductile deformation that the origins of the deposits are far less clear.

**Systems dismembered by strike-slip faults:** It has been argued that the emplacement of certain porphyry Cu deposits, especially those in the Cordillera de Domeyko of northern Chile, have been localized by arc-parallel strike-slip faults (e.g., Lindsay et al., 1995; Tomlinson and Blanco, 1997a; Richards et al., 2001), although some workers have alternate interpretations of the tectonic environment (e.g., Skárneta et al., 2003). Several deposits are cut by postore faults; the El Abra and Chuquicamata deposits are offset by faults interpreted by several as strike-slip (Sillitoe et al., 1996; Dilles et al., 1997; Tomlinson and Blanco, 1997b), and others as dip-slip (Skárneta et al., 2003). Strike-slip faults primarily produce lateral translation, so dismemberment only moderately changes the depth of exposure of the system (e.g., McInnes et al., 1999; Tomlinson et al., 2001).

**Configuration of alteration patterns**

The deep, proximal part of porphyry systems generally is dominated by potassic alteration. The upper part of systems tends to be dominated by hydrolytic alteration of the sericitic and advanced argillic types. The deep flanks, distal sides, and peripheral upper areas are zones of propylitic and, locally, sodic-calcic alteration. Among a continuum of possible overall patterns, we distinguish three of the most obvious endmember configurations: (1) hydrolytic alteration largely above potassic alteration, (2) intense hydrolytic alteration enclosed within and extending above potassic alteration, and (3) both sodic-calcic and potassic alteration located in the center of the system (Fig. 10).

**Hydrolytic alteration largely above potassic alteration:** Potassic alteration is overlain by sericitic alteration, which narrows upward to form a hood and may be overlain by relatively fresh rock (Fig. 10A). Examples include Bajo de la Alumbrera, San Manuel, Henderson, Climax, and Chorolque, Bolivia (Lowell, 1968; Steininger, 1973; Grant et al., 1980;...
Proffett, 2003a; Seedorff and Einaudi, 2004a), and this configuration corresponds to the Lowell and Guilbert or Climax-type models (Lowell and Guilbert, 1970; White et al., 1981). In another variant, the systems have extensive advanced argillic alteration, mostly at shallow levels in the system and inferred to extend to the paleosurface (Fig. 10B). Examples include Batu Hijau, Cerro Rico, Bolivia, El Salvador, Far Southwest-Lepanto, Philippines, Grasberg, Red Mountain, Arizona, and Sunnyside, Arizona (Cron, 1975; Gustafson and Hunt, 1975; Macdonald and Arnold, 1994; Grasb., 1996; Hedenquist et al., 1998; Sillitoe et al., 1995; Clode et al., 1999; Watanabe and Hedenquist, 2001), and this configuration might be regarded as the lithocap variant on the above models (Bodnar and Beane, 1980; Sillitoe, 1995b). A plan map through the center of deposit for either variant may have potassic alteration ringed by hydrothermal alteration, in turn succeeded outward by propylitic alteration. In many cases, the distinction between the variants is surely only one of exposure: perhaps all shallowed emplaced porphyry copper and gold systems were overlain by zones of advanced argillic alteration. It is not clear, however, whether deeply emplaced porphyry copper systems, which may not undergo brine-vapor separation ever had a lithocap. Likewise, other classes of porphyry deposits have few examples of advanced argillic alteration, and the evidence at some deposits such as Henderson indicates that hydrothermal alteration grades upward into fresh rock without development of advanced argillic alteration (Seedorff and Einaudi, 2004a).

**Intense hydrolytic alteration enclosed within and extending above potassic alteration:** The distribution of most intense hydrolytic alteration expands upward, with an overall geometry of an upright funnel (Fig. 10C). Intense hydrolytic alteration penetrates deeply into older potassic alteration. A plan map through the center of the deposit has a central region of hydrolytic alteration, succeeded outward by potassic alteration and then propylitic alteration, i.e., the relative positions of hydrolytic and feldspar-stable alteration in plan view are reversed compared to the previous configuration. Many of the world’s great porphyry Cu orebodies have late base metal lodes displaying this configuration, including Butte, Chuquicamata, Escondida, Chile, and Onyar, Kazakhstan (Meyer et al., 1968; Kudryavtsev, 1996; Ossandón et al., 2001; Padilla et al., 2001). This configuration corresponds to what comes the northern Chilean model (J. Perelló, pers. commun., 1997).

**Both sodic-calcan and potassic alteration located in the center of the system:** Sodic-calcan alteration is widespread and occupies a bell-shaped volume beneath potassic alteration (Fig. 10D). Fingerlike projections of sodic alteration extend up through the center of the system. A plan map through the center of a deposit with this configuration may have a central region with prominent sodic alteration, ringed by potassic alteration and succeeded outward by sodic-calcan alteration. Both sericitic and propylitic alteration may be present only at higher levels, as at Ann-Mason, or sericitic alteration may be present at this level, as at the Yerington mine (Carter, 1986; Dilles and Einaudi, 1992). In the Yerington district the sericite zones grade upward into and are overlain near the paleosurface by quartz-aluminate advanced argillic alteration (Dilles et al., 2000a). Sodic-calcan alteration is documented elsewhere in the United States (Dilles et al., 1995), Canada (Lang et al., 1995), and Mongolia (Kirwin et al., in press).

**Geometry of orebodies**

The shape of a hypogene orebody depends on an assortment of factors, including the number of mineralizing intrusions and their relative positions, the sequence and relative positions of mineralizing versus barren intrusions, the dominant structural style of the deposit, and the types and orientations of mineralized structures (e.g., Carter et al., 1988b; Proffett, 2003a; Seedorff and Einaudi, 2004b). Late diatremes and intrusions can cut out part of the orebody. In addition, ore deposition is strongly dependent on various physicochemical factors. Ore minerals may be precipitated in veins and disseminations in one or many mineral assemblages and formed contemporaneously with potassic alteration, a transitional stage associated with little wall-rock alteration, sericitic alteration, or advanced argillic alteration, or any combination of these (Seedorff and Einaudi, 2004b). Previously deposited Cu also can be removed or redistributed by fluids associated with sericitic, advanced argillic, and sodic-calcan alteration (Brinhall, 1980; Brinhall and Chioros, 1983; Dilles and Einaudi, 1992).

The geometries of orebodies cut across the classes of porphyry deposits. In porphyry Cu deposits, the thick-walled cylinder extending vertically for as much as 2 km or more is a common geometry (Sillitoe and Gapp, 1984) but many other geometries occur, including two domes or inverted cups for the pre-Main Stage at Butte (Field et al., 2005) and an inverted cup at Mamut, Malaysia (Kosaka and Wakita, 1978). The inverted cup is widely associated with porphyry Mo deposits because the three examples at Climax are well described (Wallace et al., 1968), but at Henderson the six major mineralizing intrusions display four different shapes: inverted cup, open cone, dome, and lenticular geometries (Carter et al., 1988b). The orebody at Mount Emmons, Colorado, was originally interpreted as an inverted cup, but additional drilling later demonstrated that the area above the stock is low grade, so the geometry is more like a cylinder or open cone (Ranta et al., 1984).

Although multiple overlapping ore zones are widely associated with porphyry Mo deposits, as at Climax (Wallace et al., 1968), Henderson (Carter et al., 1988b), and Hall (Shaver, 1991), they also are known from other classes of porphyry deposits, including porphyry Cu deposits (e.g., two domes for pre-Main Stage at Butte).

**Metal zoning patterns**

At the system scale, the proximal part usually contains the metals for which the class is named (e.g., Cu in porphyry Cu deposits). Most districts are zoned upward and outward to regions with higher contents of Pb, Ag, and/or Au, although in many Au-rich porphyry Cu deposits the Au content can be highest within the central Cu zone (Sillitoe, 2000). Among the more Au-rich and mafic systems, there is a zonation laterally outward to higher Au/Cu ratios (e.g., Einaudi, 1994). In contrast, Ag/Au ratios in the Bingham and Copper Canyon districts show a systematic increase of several orders of magnitude from central porphyry Cu-Au ore to distal, limestone- and hornfels-hosted Pb-Zn-Ag ores, and there is
a strong negative correlation between Ag/Au and Au content. Porphry-hosted ores at Tanumá (Cox, 1985) and Dos Pobres (Langton and Williams, 1982) also show an increase in Ag/Au outward, and highest Au grades are centrally located and associated with lowest Ag/Au. Thus, Ag/Au ratios are a useful zonal indicator and can be used as a vector for Au potential. The data are sparse for the more Ag rich and felsic systems, but they probably grade outward to higher Ag/Cu and Ag/Mo ratios (e.g., Eidel et al., 1968; Babcock et al., 1995).

In porphyry Mo deposits such as Henderson, Cu occurs in trace amounts associated with sphalerite, outboard of Mo (Seedorff and Einaudi, 2004b). Tungsten occurs as wolframite and also generally occurs outboard of Mo but inboard of Zn and Cu.

In porphyry Cu deposits, the relative positions of Mo and Cu are variable. At Bingham and Butte, the Mo zone lies beneath but overlaps the Cu zone (John, 1978; Babcock et al., 1995; Field et al., 2005). In untitled plan view, Cu is also distal to Mo at Ann-Mason, but the Cu may have moved outward from an original position inboard of Mo during late sodic alteration (Dilles and Einaudi, 1992). At the nearby Yerington mine, Mo is distal to Cu (M.T. Einaudi, unpub. data, 1970). The Mo shell is not particularly well developed at either Dos Pobres (Langton and Williams, 1982) or Saimdak, Pakistan (Sillitoe and Khan, 1977), but the patterns suggest that Mo is distal to Cu at both deposits. The position of Cu relative to Mo is not clear at El Salvador (Gustafson and Quiroga, 1995) and Sar Cheshmeh, Iran (Waterman and Hamilton, 1975).

The data that are available for W at El Salvador indicate that Mo is distal to W, which is opposite of the zoning for Climax (Wallace et al., 1968) and Henderson (Seedorff and Einaudi, 2004b).

Challenges and applications

Industry geologists have a strong interest in spatial distributions at all scales and use them to site drill holes for exploration, geotechnical, metallurgical, and environmental purposes. In spite of complications within single deposits and wide variability between deposits, the geometries and spatial distributions of porphyry dikes and source granitoid cupolas, hydrothermal alteration and mineralization zones, ore minerals, and veins are useful for exploration targeting in conjunction with geologic hypotheses for their spatial distribution. Maps and sections that display rock contacts, structures, and alteration and mineralization zones have the advantage that they convey important geologic information simply on one layer. Many orebodies have also been found by mapping the abundance of vein and breccia types and of hydrothermal minerals (e.g., Perry, 1933; Hunt, 1985; Marsh, 2001). Scientific understanding of hydrothermal zones and metal distributions is relatively poor today but requires knowledge of fluid-flow directions and paths as well as understanding of fluid-rock reactions that dictate ore deposition and wall-rock alteration (Reed, 1997; Geiger et al., 2002).

Although chemically similar metals such as Cu and Au may not be strongly fractionated in some deposits (Halter et al., 2002), hydrothermal precipitation of metals during evolution of the hydrothermal system in general fractionates metals and forms metal zones (cf. Seedorff and Einaudi, 2004b). Metal ratios are a reconnaissance exploration tool in the form of rock chip analyses (e.g., Lowell, 1991; Garvin, 2002) and are also particularly useful in finding extensions and new orebodies in existing mining districts (Petersen et al., 1977; Atkinson and Einaudi, 1978; Manske and Paul, 2002).

Space-Time-Temperature Relationships and Geochemical Evolution at the System Scale

Wallace et al. (1968) and Gustafson and Hunt (1975), respectively, summarized the evolution of the Climax porphyry Mo system and of the El Salvador porphyry Cu system with cumulative cross-sectional time panels. This approach remains a powerful means of display (e.g., Proffett, 2003a).

Space-time diagrams plot igneous and hydrothermal features as a function of time versus position in the orebody (Barton, 1982), which allows for the temporal evolution and spatial zoning to be portrayed in a single illustration (e.g., Dilles et al., 2000a). Seedorff and Einaudi (2004a) introduced evolutionary tree diagrams using axes of time and temperature to illustrate the linkages among assemblages in space and time as a function of temperature.

Constraints on time

Absolute time and geologic calibrations: Geochronologic methods applicable to dating the hypogene aspects of porphyry deposits have improved greatly in the last quarter century with improvements in mass spectrometry, use of lasers and ion microprobes, and addition of new isotopic systems (Chesley, 1999). The methods most commonly used have been K-Ar (e.g., Dannon et al., 1981; Arribas et al., 1995); 40Ar/39Ar (e.g., Geissman et al., 1992; Marsh et al., 1997), Re-Os (e.g., McCandless and Ruiz, 1983; Stein et al., 1998), and U-Pb (e.g., Dilles and Wright, 1985; Ballard et al., 2001). Apparent analytical uncertainties of -0.25 percent are obtained for the 40Ar/39Ar dates, -0.50 percent for U-Pb and Re-Os dates, and generally greater uncertainties for K-Ar dates (Chesley, 1990); however, these systems may be affected by inheritance or thermal/chemical resetting, so the geologic uncertainties are typically greater.

Porphry systems provide a challenge to date radiometrically because multiple thermal pulses related to intrusions may cause the analyzed grains to be partially to fully reset. Not surprisingly, geochronology studies sometimes produce unsatisfying results (e.g., Gustafson et al., 2001). In an attempt to meet the challenge, most complicated deposits are now being attacked with multiple isotopic methods (e.g., Lund et al., 2002; Maksaev et al., 2004). The most difficult problem is establishing the duration of the hydrothermal system(s). For several of the giant porphyry Cu deposits, including Chuquicamata, Escondida, and Butte, geochronology suggests either that hydrothermal systems were active over a period of several million years or that there were multiple pulses of activity (Reynolds et al., 1998; Lund et al., 2002; Maksaev et al., 2004; Padilla et al., 2004). In other districts, there is clear evidence for petrologically unrelated hydrothermal systems being superimposed on one another (e.g., Newberry et al., 1991; Marsh et al., 1997), commonly bedeviling attempts to discern absolute ages and genetic relationships between magmatic events and ores and complicating efforts to ascertain the duration of any one system.
Other geochronologic data support durations of hydrothermal systems, including some very large deposits such as Far Southeast and Batu Hijau, which are considerably shorter. Two well-dated porphyry systems are only a little over a million years old: Far Southeast was active for about 150,000 yr (Arribas et al., 1995) and Ok Tedi for less than 1.2 m.y. (Page and McDougall, 1972). Other precisely dated porphyry deposits include Batu Hijau with a duration of 80,000 yr (Garwin, 2002) and several porphyry systems in the Potrerillos area, where none was active for more than a few 10,000 yr (Marsh et al., 1997).

**Thermal models:** Time-temperature (thermal) models provide time estimates of the sustainability of hydrothermal systems, with assumptions of either conductive or convective cooling of plutons (e.g., Norton, 1982; Hanson, 1996) and convection or stagnant crystallization of the magma chamber (Shinohara and Hedenquist, 1997). Results on single intrusions indicate that even large stocks, e.g., 100 km$^2$ and 1 km half-thickness, should cool in less than 100,000 yr, and porphyry stocks should cool in a few tens of thousands of years, which is presently beyond the resolution of existing radiometric dating techniques (e.g., Marsh et al., 1997). Multiple intrusions are required to prolong hydrothermal activity.

**Time lines:** When an intrusion or breccia body is emplaced, it cuts off preexisting features at its contacts and the time of emplacement serves as a time line in the evolution of the system. Numerous time lines can be created during the evolution of a hydrothermal system as successive dikes, stocks, or breccia bodies are emplaced. The relative temporal resolution of such geologic time lines is provided by the time to form a porphyry intrusive contact, which is likely to be days to years compared to >10,000-yr resolution of geochronologic methods.

**Space-time characteristics of porphyry systems:**

The commonly observed crosscutting relationships indicate potassic alteration forms earlier, followed by sericitic alteration; where it is present, sericitic alteration commonly moves downward as temperature declines at a given location. In contrast, theoretical considerations suggest that potassic and sericitic alteration must form contemporaneously at some point in the evolution of the system, because a cooling magmatic fluid first produces K-feldspar + biotite and then at a lower temperature produces sericite according to the mineral-fluid relationships such as shown in Figure 5A. Porphyry systems commonly display cyclic complexity of hydrothermal events similar to other types of igneous-related hydrothermal systems (e.g., Barton, 2000). The emplacement of multiple plutons should lead to cyclicity, i.e., repetitions of hydrothermal features. This underscores the point made in the earlier section on relative ages: hydrothermal features (e.g., vein types) are not time lines but rather are parts of an evolving sequence that may be repeated in part or in their entirety.

**Types of crosscutting relationships:** If vein types in a porphyry deposit are perfectly arranged in order of decreasing temperature and if temperature at any given point in space declines with time, then the early, high-temperature veins would be consistently cut and offset by lower temperature veins. This type of observation could be regarded as a “normal” crosscutting relationship between veins (cf. Seedorff and Einaudi, 2004a).

Given that most deposits have a series of porphyry intrusions, cyclicity is expected, as noted above, and “anomalously” crosscutting relationships should also be expected, i.e., situations where higher temperature veins cut and offset lower temperature veins. The “normal” and “anomalously” occurrences together with spatial location can be tabulated in a matrix of offsetting vein relationships in order to reconstruct the time-space-temperature dynamics of the hydrothermal system (Seedorff and Einaudi, 2004a).

**Frames of reference:** Because hydrothermal fluids flow in space as a function of time, there are two common frames of reference used in thinking about evolutionary pathways: one is the rock perspective, where the focus is the sequence of reactions in a small volume of rock; the second is the fluid perspective, where the focus is the changing geochemical state of a packet of fluid as it moves through the rock.

Cumulative time panels and space-time diagrams use the rock perspective. In contrast, the evolutionary tree diagrams also display the fluid perspective (Seedorff and Einaudi, 2004a), the common frame of reference for tracing evolutionary processes in phase diagrams. The trace of the arrows from left to right in a space-time diagram or from top to bottom of the evolutionary tree diagram shows the possible succession of mineral assemblages or vein types in a packet of moving fluid with time, for example in the space-time and evolutionary trees for El Salvador and Henderson (App. 4).

**Evolutionary styles:** Numerical models predict that an intrusion ought to be accompanied by a rapid period of heating of adjacent rock, followed by prolonged cooling (e.g., Marsh et al., 1997). In porphyry systems, from a rock perspective the few thermal progradations recognized are either small early events (e.g., Gustafson and Quiroga, 1995) or sudden reversals during a mid-life change (e.g., Seedorff and Einaudi, 2004). From a fluid perspective, a thermally prograding (heating) fluid path occurs where a nonmagmatic fluid flows toward an intrusion (e.g., Dilles and Einaudi, 1992), in contrast to magmatic-hydrothermal fluids that generally follow a thermally retrograding (cooling) path.

An evolutionary style, in the sense of Seedorff and Einaudi (2004a), refers to the degree of cyclicity of a hydrothermal system. Hydrothermal systems that are predominantly thermally retrograding exhibit evolutionary styles that are bounded by two conceptual end members, perfectly unidirectional and perfectly cyclical, with a continuum of variably cyclical behavior in between. In the perfectly unidirectional style, assemblages are deposited in an uninterrupted sequence from high to low temperatures, without reversals, irrespective of the number of intrusions, whereas in the perfectly cyclical style the entire range of high- to low-temperature mineral assemblages is repeated for each successive intrusion. Additional deposits may be referred to as composite systems, which are spatially overlapping by temporally distinct.

**Geochemical evolution**

**Evolutionary paths:** Here, we describe lithophile evolution as a function of temperature from the fluid perspective. We focus on sulfidation state because of its strong link to mafics but allude to other related parameters such as oxidation state and pH or $\alpha_{H_2O}/\alpha_{H_2}$. 
Porphyry gold and porphyry copper deposits: Mineral assemblages in porphyry Cu deposits suggest that these deposits evolve at relatively higher sulfidation and oxidation states than other types of porphyry deposits. Porphyry Au deposits have some similarities to shallow porphyry Cu systems (Monteau and Einnaudi, 2000, 2001). The highest temperature potassic assemblages are typically sulfide poor, and the highest temperature veins in most low-pressure (shallow) deposits contain magnetite veinlets with minor quartz, biotite, and ilmenite, compatible with intermediate- to low-sulfidation and oxidation states (Fig. 11). More oxidized conditions prevail in deeper potassic assemblages at ~500°C to 700°C as indicated by abundant magnetite, local hematite, and replacement of ilmenite and titanite by rutile ± hematite in El Salvador and titanite by ilmenite-hematite intergrowths at Butte (Gustafson and Hunt, 1975; Roberts, 1975). Relatively low ratios of KCl/HCl in solution at high temperature are required by assemblages containing corundum, andalusite, and sericite, together with biotite and K-feldspar at Butte (Brimhall, 1977), El Salvador (Gustafson and Hunt, 1975), and Batu Hijau (which lacks K-feldspar, Clode et al., 1999). In most deposits, slightly younger and lower temperature assemblages consist of bornite-digenite, bornite, bornite-chalcopyrite, or chalcopyrite with or without magnetite, as at Batu Hijau (Clode et al., 1999, and our work) and Bajo de la Alumbrera (Proffett, 2003a). In contrast, chalcopyrite-pyrite is the stable sulfide assemblage at temperatures of ~600°C in some deeper deposits such as Butte (Roberts, 1975; Brimhall, 1977; Field et al., 2005). All of these assemblages suggest that sulfidation state remains at intermediate values during the main period of ore deposition. Sulfides rarely exceed 2 vol percent of rock, and Cu grades are up to 1.5 wt percent, although typically half this amount. Anhydrite is characteristic of many potassic mineral assemblages, and it may be more abundant than sulfide, as in the Continental area at Butte (Brimhall, 1977), El Salvador (Gustafson and Hunt, 1975), and El Teniente (Camus, 1975); at Butte, sulfur isotope data indicate that sulfate predominates over reduced sulfide in solution at ~600°C (e.g., Field et al., 2005).

At temperatures below ~450°C, abundant (2–7 vol %) pyrite, typically associated with sericitic alteration, becomes the dominant sulfide in most porphyry Cu systems. At ~300°C porphyry Cu and porphyry Au deposits display a large range of sulfidation states, from low (pyrrhotite ±

![Fig. 11. Log fO2-1,000/T diagram defining the relative sulfidation states of hydrothermal fluids for various classes of porphyry deposits, based on figure 8 of Einnaudi et al. (2003). See text for discussion. Inset is log fO2-1,000/T diagram from Einnaudi et al. (2003) defining the relative sulfidation states of hydrothermal fluids based on sulfidation reactions involving two or more copper- and iron-bearing mineral phases. The Main line of ore-forming environment is from Barton (1970), and the generalized evolutionary path of fluids in porphyry copper and porphyry-related vein deposits is discussed in Einnaudi et al. (2003).]
pyrite), through some typical intermediate (pyrite with chalcopyrite, tennantite-tetrahedrite, and/or sphalerite with moderate iron content), to high and very high (pyrite with bornite, enargite, digenite, covellite or nakamidite, and low iron sphalerite; e.g., Einaudi, 1977b; Inan and Einaudi, 2002). In porphyry-related base metal lodes, these high-sulfidation assemblages are particularly characteristic of the proximal parts of porphyry-related base metal deposits, whereas intermediate-sulfidation states of sulfides prevail in distal portions rich in Mn and Ag and containing pyrite with moderate iron sphalerite, galena, and carbonates (e.g., Burtos, 1989). Sulfates in low-temperature advanced argillic assemblages include alunite and minor barite, and sericite alteration in El Salvador, Chuquicamata, and Far Southeast contains anhydrite (Gustafson and Hunt, 1975; Ossandón et al., 2001). The Cu-Fe sulfides commonly precipitate between about 550° and 300°C (Redmond et al., 2004; Field et al., 2005) as a result of temperature decrease and rock reactions that buffer pH and increase proportion of reduced sulfide, in accordance with Cu solubility experiments (Hemley et al., 1992).

**Porphyry molybdenum deposits:** Porphyry Mo systems contain few sulfides and oxides that fix sulfide state. In the high-temperature (500°–700°C) parts of the Henderson deposit in silicic and potassic alteration, molybdenite is the chief or only sulfide and anhydrite is absent. At about 400° to 500°C at Henderson, veinlets contain pyrite in apical positions with respect to porphyry intrusions and magnetite in distal positions of the Seriate center associated with sericite, green biotite, or topaz (Seedoff and Einaudi, 2004a). These assemblages suggest intermediate- to low-sulfidation states. Younger, shallower, and distal low-temperature (200°–325°C) alteration zones at Henderson are characterized by clay, carbonate, and Mn-rich garnet; sulfides include pyrite or pyrrhotite (<1 vol %) with minor low iron sphalerite, galena, local chalcopyrite, and local hematite, which indicate intermediate-sulfidation states (Fig. 11). Therefore, Henderson and other similar porphyry Mo deposits (Seedoff and Einaudi, 2004b) evolve at lower temperature than most porphyry Cu and porphyry Au deposits. This might arise in fluids that have fairly low total sulfur contents and are relatively oxidized.

**Porphyry tungsten deposits:** Porphyry W deposits at Mount Pleasant contain assemblages at both relatively high and moderate temperatures that indicate low- and locally very low-sulfidation conditions: biotite-K-feldspar alteration (~350°–500°C) with wolframite, arsenopyrite, local loellingite, pyrrhotite or pyrite, and molybdenite; and sericite alteration with native bismuth and bismuthinite (e.g., Koimima et al., 1986). At lower temperature (150°–300°C), assemblages evolve from sphalerite, galena, stannite, bismuthinite, pyrite, and chalcopyrite, to tennantite-tetrahedrite with pyrite and bismuthinite, representing a slight increase in sulfidation into the lowest parts of the field of intermediate-sulfidation state (Fig. 11). Nonetheless, the entire sulfidation evolution at Mount Pleasant represents a lower sulfidation path than for the porphyry Au, Cu, Mo, and Sn, possibly reflecting dominance of reduced sulfur species over oxidized sulfur species throughout the evolutionary path.

**Porphyry tin deposits:** High-temperature conditions in porphyry Sn deposits are poorly characterized, but at >400°C they are typified by cassiterite with minor pyrrhotite in association with quartz and tourmaline (Grant et al., 1980) and, therefore, have a low-sulfidation state. At Llallagua and Chorolque (Turner, 1935; Grant et al., 1980), assemblages that formed at 350° to 200°C contain pyrite or pyrrhotite, arsenopyrite, cassiterite or stannite, bismuthinite, franckeite, minor high iron sphalerite and chalcopyrite associated with tourmaline and sericite. The youngest alteration contains abundant pyrite with sericite, and in some cases also alunite, pyrophyllite, and dickite (e.g., Cerro Rico; Sillitoe et al., 1998); these pyritic assemblages indicate a transition from low- to intermediate-sulfidation state with decreasing temperature (Fig. 11).

### Sources of diversity

Here, we summarize how different classes of porphyry deposits evolve with respect to sulfidation state. Hydrothermal alteration and mineralization in the porphyry Cu and porphyry Au deposits that we have reviewed (see also Einaudi et al., 2003) starts at the transition from low- to intermediate-sulfidation states and may evolve upon cooling to moderate temperature to a wide range of sulfidation states (Fig. 11). At a given initial SO₂/H₂S gas ratio imposed by juvenile magmatic-gas input (~SO₂ gas flux" arrow, Fig. 11), the sulfidation trajectory will depend on the buffering capacity of the rock and the net ratio of fluid to rock (i.e., the balance between sulfur and iron species in fluid and rock). In deposits with a high water-to-rock ratio, sulfide deposition at ~300°C may take place at high- to very high sulfidation states in association with advanced argillic alteration in quartz-monzonite-hosted base metal lodes, as at Butte. Alternately, where wall rock has low buffering capacity, very high sulfidation states can be achieved, as in quartzite at Bingham (Inan and Einaudi, 2002). High-sulfidation states require cooling at relatively fixed or increasing f₂S, suggesting these systems evolved along the H₂S = SO₂ sulfur-gas buffer.

Porphyry Cu and porphyry Au systems may preferentially reach very high sulfidation states because of (1) high contents of the parent magma (see below) and (2) high oxidation state controlled by a high fluid-to-rock ratio and SO₂-rich gases following the S-gas buffer (Einaudi et al., 2003). Upon cooling disproportionation of SO₂ produces H₂S via the reaction:

\[
4\text{SO}_2(g) + 4\text{H}_2\text{O} = 3\text{H}_2\text{SO}_4(s) + \text{H}_2\text{S}(g) = \text{3H}_2\text{O}(l) + 3\text{HSO}_4^{-}(aq) + \text{H}_2\text{S}(aq).
\]

(3)

Under typical hydrostatic conditions characteristic of the transition from porphyry Cu to base metal lodes, the disproportionation reaction goes strongly to the right beginning at ~400°C (e.g., Ohmoto and Rye, 1979), leading to increased f₂S and a greater likelihood of sulfide precipitation. At temperatures >500°C in shallowly emplaced porphyry systems, the paucity or lack of sulfides and relatively high magnetite content (e.g., John, 1989b; Muntean and Einaudi, 2000; Ulrich and Heinrich, 2001; Frohlich et al., 2003a) is consistent with less SO₂ disproportionation at low pressure than at high pressure, all else being equal, and a low SO₂ content of the iron-rich brine as a result of fluid immiscibility and separation of SO₂-rich vapor (see Field et al., 2005). Conversely, in some deeply emplaced deposits such as Butte (Fig. 9), where there...
is evidence for high-temperature (−600°C) precipitation of pyrite-anhydrite-molybdenite ± chalcopyrite (Roberts, 1975; Brinnhail, 1977), higher fluid pressures cause an increase in $f_{\text{H}_2\text{S}}$ through reaction (3) (Field et al., 2005).

In contrast, rock buffering (Giggenbach, 1987) of $f_{\text{H}_2\text{S}}$ is characteristic of the low-temperature evolution of many porphyry systems. In rock-buffered fluids, iron-rich country rocks scavenge $\text{H}_2\text{S}$ by iron sulfide precipitation and reduce the effective $f_{\text{S}_2}$ via reaction (1). This effect is most readily seen in andesitic or basaltic host rocks near the periphery of sulfide zones. Examples of pyrite-rich zones grading outward into pyrite-pyrrhotite or pyrrhotite zones of low-sulfidation state occur at the Mamut (Kósaka and Wakita, 1978) and North Fork, Washington (Herdick et al., 1995) porphyry copper deposits.

Most porphyry Mo, Sn, and W deposits, however, do not show evidence for intermediate- to high-sulfidation conditions at high or moderate temperatures. These deposits have lower total sulfide, typically lack anhydrite, and are generally compatible with origin from hydrothermal fluids of low total sulfur content and low concentrations of $SO_4^2−$. In general, the assemblages are rock-buffered at temperatures <350°C at a sulfidation state near the low to intermediate boundary.

Thus, at high temperature the sulfidation state decreases in the order: porphyry Cu and Au, porphyry Mo, porphyry Sn, porphyry W. The likely cause of the different sulfidation states of high-temperature porphyry fluids is the parent magma. There are two key factors: sulfur capacity of the melt and sulfur species in the melt. The andesite and dacite magmas associated with porphyry Cu and porphyry Au deposits are typically water rich, oxidized (>QFM +3, i.e., greater than three log units above the quartz-fayalite-magnetite buffer), and S-rich (Dilles, 1987; Streck and Dilles, 1998). Oxidized, sulfate-rich, hydrous andesite-dacite melt inclusions trapped at >550°C commonly contain >1,000 ppm S (Wallace, 2003), in accord with measurements of volcanic gases containing 10,000 to 50,000 ppm total S with molar $SO_2$/$H_2S$ of ~1 to 10 at 700° to 900°C (e.g., Einaudi et al., 2003). Figure 11 shows the composition of gases in equilibrium with such an oxidized, S-rich magma.

Based on the mineralogical summary given above, porphyry Mo, Sn, and W magmas may have oxidation states lower than porphyry Cu magmas. Few quantitative data exist, but rhyolite associated with the Pine Grove porphyry Mo deposit lies at an oxidation state of QFM +1 to QFM +2 (Keith and Shanks, 1988). Under such conditions, only minor amounts of $H_2S$ and $SO_4^2−$ are dissolved in magma (Carroll and Rutherford, 1988), and gases would be expected to have $SO_2$/$H_2S$ of <0.1 to 1 (Ohmoto and Rye, 1979). Figure 11 illustrates how such a melt would produce only small amounts of $H_2S$-rich gas (<600–3000 ppm S species in magmatic gases), consistent with generation of porphyry Sn and porphyry W systems with relatively low sulfidation states.

Magma compositions also dictate the initial $m_{KCl}$, $m_{K/Na}$, of magmatic-hydrothermal fluid. Metaluminous magmas or magmas with high $K/Na$ (e.g., granites) will commonly produce potassic and sericitic alteration on cooling (path A, Fig. 12). In contrast, magmas that are more peraluminous or have low $K/Na$ (e.g., tonalite) produce hydrothermal fluids with low $m_{KCl}$, $m_{K/Na}$, that tend to cause advanced argillic alteration on cooling (path B, Fig. 12; cf. Batu Hijau). Other factors that generate advanced argillic alteration include condensation of vapor that separated from brine, at high water/rock ratios (path C, Fig. 12). In contrast, potassic and sericitic alteration tend to be produced by brines that have higher $KCl$/$HCl$ and smaller masses than related vapors and commonly remain near rock-buffered conditions, i.e., relatively low water/rock ratios.

In summary, the spectrum of magmas related to porphyry deposits appears to produce a continuum of oxidation states (Burnham and Ohmoto, 1980), sulfidation states (John, 2001; Einaudi et al., 2003), total sulfur contents, and $a_{K/Na}$ (or $m_{KCl}$, $m_{K/Na}$, $m_{HCl}$, $m_{Na}$) values that correlates with class of porphyry deposit. Individual deposits, however, exhibit variability in the lower temperature portion of their paths that depends on subsequent evolution of the fluid.

**Challenges and applications**

Systematically collected observations on crosscutting relationships between intrusions and veins and between various types of veins from throughout a deposit are essential to develop an accurate descriptive model of the space-time evolution of mineral assemblages. These data are only available for few porphyry systems and thus may not be representative. Because most porphyry systems exhibit a component of cyclic behavior, knowing the place of a sample in the space-time evolution of the entire system is necessary to make a meaningful interpretation of geochemical and geochronologic data.
Systematic studies of space-time relationships and documentation of anomalous crosscutting relationships are needed from the deposits that appear to have formed over long time intervals, to determine whether these intervals indicate long-lived, persistently active hydrothermal systems or spatially superimposed but temporally widely spaced pulses of short duration.

Classification of Porphyry Deposits

Early efforts to classify porphyry Cu deposits, based on principal metals, led to three classes: Cu-Mo, Cu-Au-Mo, and Cu-Au (Kesler, 1973; Cox and Singer, 1988, 1992; Kirkham and Sinclair, 1996). By the 1970s, it was recognized that there was considerable diversity in the morphology of orebodies that was a function of tectonic setting and depth of emplacement. Sutherland Brown (1976b) and McMillan and Pantelleyev (1980, 1985) created three end members termed plutonic, volcanic, and classic types. Cheney and Trammell (1996) proposed a three-dimensional classification using depth (porphyry vs. plutonic), predominant metal, and compositions of igneous rocks. Classification schemes also have included important petologic and tectonic components, such as continental versus island arc (Kesler et al., 1975), petologic and tectonic characteristics, metal ratios, and other criteria (Sillitoe, 1980; Mutscher et al., 1981; Krivtsov and Ageyeva, 1982; Carten et al., 1993; Zvezdov et al., 1993), and alkalinity and ferric/ferrous ratios of cogenetic igneous rocks (e.g., Keith and Swann, 1996).

Classification considerations

Use of quantitative geochemical data for classification of porphyry deposits has limitations, despite having the appearance of objectivity. For example: (1) overall mean grades and metal ratios are a function of depth of erosion and host rock, to name a few factors; (2) some elements are preferentially enriched and depleted by supergene processes so that metal contents of weathered or partially weathered deposits are skewed according to their position in the weathering profile; (3) igneous petrologic and geochemical classifications—especially those that involve alkalis—can be highly sensitive to even slight alteration; and (4) production data are strongly affected by differences in recovery between products and byproduct metals and by short-term economic variations. These limitations were recognized by Kesler (1973) but have remained a basis of classification by some (Cox and Singer, 1988, 1992).

Strategy

The proposed classification (Table A2) has similarities with those of Zvezdov et al. (1993), Carten et al. (1993), and Westra and Keith (1981). We choose to avoid criteria that depend on precise determination of ratios (e.g., Mo/Au ratio) or threshold values, as in the case of Au-rich porphyries where there is no natural boundary at 0.4 ppm Au. We employ metal ratios but generally only with the precision implied by a log scale depiction (Fig. 13), which are available to the exploration geologist at the prospect stage.

The top level of classification—class of deposits—is based on the predominant metal. For the next level down, we use ranges of igneous rock compositions together with general metal associations, guided also by distinctive geologic characteristics (Table A2). The compositional ranges overlap between subclasses (Fig. 8), so igneous compositions need not be precise. Further divisions are based on structural styles. The grade-tonnage (Fig. 3) and petrologic (Fig. 8) plots

![Fig. 13. Metal ratios in porphyry systems. A. Log-log plot of Au/Ag vs. Cu/Mo. Classified deposits are plotted, keyed to subclasses of deposits. Porphyry classes are outlined. See text for further explanation. B. Distribution of classified deposits on a Cu-Mo-Au triangular plot using conventions of Kesler (1973). C. Key to divisions of triangular diagram in (B) using classification system of Cox and Singer (1988, 1992), where PCD stands for porphyry copper deposit.]
presented earlier, coupled with plots of the Au/Ag versus Cu/Mo ratio and the Cu-Mo-Au triangle (Fig. 13), provide useful means of comparison.

Classes and subclasses and their distinctive features

Porphyry gold deposits: At present, the porphyry Au class consists of only one subclass: dotoritic porphyry Au (Table A2). The first descriptions were from the Maricunga belt of Chile (Vila and Sillitoe, 1991) but reported examples are now known from Persi, Nevada, and California. There are few data on metal abundances for porphyry Au deposits. The two deposits plotted in Figure 13 have fairly high Au/Ag ratios but Cu/Mo ratios that, surprisingly, are at the low end of the range exhibited by porphyry Cu deposits. The geochemical plots are consistent with the geologic data that suggest these deposits form a continuum with the porphyry Cu class, in particular the tonalitic-granodioritic porphyry Cu-(Au-Mo) subclass, with which they are linked through their gradational igneous compositions. The Cerro Casale deposit in the Maricunga belt of Chile and Cerro Corona in Peru are closest to being transitional into the porphyry Cu class.

Porphyry copper deposits: The porphyry Cu class consists of four subclasses. We distinguish two subclasses within the continuum of subalkaline rocks: the aforementioned tonalitic-granodioritic porphyry Cu-(Au-Mo) deposits and the more silicic quartz monzodioritic-granitic porphyry Cu-(Mo) deposits. To a first approximation, these two subclasses correspond to the two most abundant varieties of porphyry deposits recognized since the early 1970s, which were then known more for their typical tectonic settings in island and continental arcs, respectively. There also are the alkaline sub-classes of porphyry Cu deposits: the monzonitic porphyry Cu-(Mo-Au) deposits and silica-undersaturated syenitic porphyry Cu-(Au) deposits.

Members of the more mafic subalkaline subclass, tonalitic-granodioritic porphyry Cu-(Au-Mo) deposits, have fairly high Au/Ag ratios—typically above 0.3. Their reported Cu/Mo ratios vary by more than two orders of magnitude (Fig. 13), from which the argument can be made that this group deserves further subdivision. If so, the Mo-poor and/or more mafic examples, including Kola (Chivas, 1978), Salamat, Russia (Zvezdov et al., 1993), and perhaps Panguna (e.g., Ford, 1978), might be separated from the less mafic and/or more Mo-rich examples such as Island Copper (Perelló et al., 1995) and Sipalay, Philippines (Kinkel et al., 1995). Although tonalitic-granodioritic porphyry Cu-(Au-Mo) deposits predominate in island arcs, there are many examples in the continental arcs of North and South America (Table A2).

Members of the more silicic subalkaline subclass, the quartz monzodioritic-granitic porphyry Cu-(Mo) deposits, have similar Cu/Mo ratios but generally much lower Au/Ag ratios than their more mafic counterparts. This group of deposits includes many—but not all—of the porphyry Cu deposits of southwestern North America and South America, including Butte (e.g., Meyer et al., 1968), Morenci, Arizona (Moolick and Duke, 1966), and Chuquicamata (e.g., Osandón et al., 2001). The Au/Ag ratios of ore are more than an order of magnitude less than the values for average silicic, intermediate, and mafic rocks (Fig. 13A). Thus, this subclass of deposits produces large amounts of Ag but little Au.

The alkaline syenitic porphyry Cu-(Au) deposits, which include Galore Creek, British Columbia (Enns et al., 1995), are rare (Table A2). These deposits have some of the highest Cu/Mo ratios of all porphyry deposits and have intermediate Au/Ag ratios (Fig. 13). Members of the second alkaline subclass of monzonitic porphyry Cu-(Mo-Au) deposits are present in many of the major porphyry provinces of the world and include several giant deposits including Bingham (e.g., Babcock et al., 1995), Ok Tedi (Bush and Seegers, 1990), and Peschanka (Zvezdov et al., 1993), but they are less common than the two subclasses of subalkaline deposits (Table A2). The compositions of igneous rocks at Bajo de la Alumbrera and Grasberg probably are close to, but may be slightly less alkali than, the monzonites of the monzonitic porphyry Cu-(Mo-Au) subclass. Monzonitic porphyry Cu-(Mo-Au) deposits have similar Au/Ag and Cu/Mo ratios to the subalkaline, tonalitic-granodioritic porphyry Cu-(Au-Mo) deposits, except that the monzonitic deposits may exhibit a more restricted range of Cu/Mo ratios.

Porphyry Cu deposits form a continuum of igneous compositions (Fig. 8). In current usage, most of the deposits that would be considered "porphyry Cu-Au deposits" are either in the subalkaline tonalitic-granodioritic porphyry Cu-(Au-Mo) subclass or the alkaline monzonitic Cu-(Mo-Au) subclass. Interestingly, these fields are not contiguous on any of the petrologic plots (Fig. 8); instead the two groups are separated from one another by the field occupied by deposits of the quartz monzodioritic-granitic porphyry Cu-(Mo) subclass, which are generally Au poor (Fig. 13). Likewise, deposits from the two Au-rich groups are intermixed in the triangular Cu-Mo-Au plot, with examples from both subclasses spreading across both the Cu-Au-Mo and Cu-Au fields.

Porphyry molybdenum deposits: We recognize six subclasses, several of which are uncommon, within the porphyry Mo class (Table A2). Our geochemical database is inadequate to make many comments based on Ag, Au, Cu, and Mo values other than to state that most of the Mo deposits for which there are data plot in the quadrant of low Au/Ag ratios and low Cu/Mo ratios.

There are few examples of alkaline monzonitic porphyry Mo-(Au) deposits, including Three Rivers, New Mexico (Giles and Thompson, 1972), and Central City, Colorado (Rice et al., 1985). The alkaline syenitic porphyry Mo subclass has a few more examples that are scattered around the globe, including Cave Peak, Texas (Sharp, 1979), and Werner Bjerge, Greenland (Brooks et al., 1982). Few of these deposits have been studied comprehensively, and neither the possible relationships between these two classes nor with their monzonitic and syenitic counterparts among porphyry Cu deposits are well understood.

Deposits of the quartz monzodioritic-granitic porphyry Mo-Cu subclass have Mo = Cu. The best studied examples are in western North America, including Hall (Shaver, 1991), Buckingham, Nevada (Theodore et al., 1992), and Cumobabi, Sonora (Scherkenbach et al., 1985). This is a geologically important subclass and exhibits the most obvious continua to porphyry Cu deposits of the quartz monzodioritic-granitic porphyry Cu-(Mo) subclass. Nonetheless, deposits of this subclass do exhibit certain distinctive igneous and hydrothermal features (Table A2).
The granitic porphyry Mo deposits are Cu-poor deposits in which the mineralizing intrusions are biotite-granite porphyries (Table A2). Less differentiated rocks, such as granodiorite, may be present in the same magmatic center. Some of the best known examples are Endako (Selby et al., 2000), Quartz Hill (Ashleman et al., 1997), and Thompson Creek, Idaho (Schmidt et al., 1983). As a group, this subclas is fairly diverse, particularly with inclusion of the transitional examples.

The trondhjemitic porphyry Mo subclas is highly unusual because the mineralizations intrusions are not particularly silicic and the examples are as old as Archean, such as Setting Net Lake, Ontario (Ayres et al., 1982). Other examples include Trout Lake, British Columbia (Linnen et al., 1995), and probably Kirk, Greece (Ariaka, 1979), and Bald Hill, New Zealand (Bates, 1989). Hydrothermal albite, muscovite, and carbonates are present; sodic alteration or assemblages transitional to sodic are possible, perhaps obscuring the original igneous compositions.

The rhyolitic porphyry Mo subclas (Table A2) includes the classic Climax-type deposits and those that are considered transitional to them by Westra and Keith (1981) and Carten et al. (1993). Representative deposits include Climax (Wallace et al., 1968), Henderson (Carten et al., 1985b), Questa (Cline and Bodnar, 1984), Pine Grove (Keith et al., 1986), and Bear Mountain, Alaska (Barker and Swainbank, 1986).

Porphyry tungsten deposits: This subclas has only one respective subclas: rhyolitic porphyry W-Mo (Table A2). The known examples are limited to New Brunswick, with Mount Pleasant (Koistman et al., 1986) being the only deposit of the class to have been mined by modern methods. It generally has been included among porphyry Mo deposits, but it has considerably more W than any other porphyry Mo deposit, and as discussed above, the sulfide-oxide assemblages indicate conditions that are of substantially lower oxidation and sulfidation state than any porphyry Mo deposit.

Porphyry tin deposits: The porphyry Sn class also has only one subclas: rhyodacitic porphyry Sn (Table A2). All known deposits occur in Bolivia (Grant et al., 1980), but Majuba Hill, Nevada (MacKenzie and Bookstrom, 1976), has many similarities with the exception of its deep zone of low-grade Mo mineralization.

Structural styles

Every class of porphyry deposit and most subclas—including many of the subclas with only a few representatives—have one or more examples with significant breccias (Table A2), suggesting that breccia formation is independent of magma composition. The deposits with abundant breccia typically have multiple varieties of breccia.

The distribution of deposits along the continuum between lode and disseminated styles (Einaudi, 1982a) as a function of subclas is intriguing. First, deposits with an important lode component are most abundant among the two subalkaline subclas of porphyry Cu deposit, the tonalite-granodioritic porphyry Cu-(Au-Mo) and quartz monzodioritic-granitic porphyry Cu-(Mo) deposits. Nonetheless, there are several examples from among the monzonitic porphyry Cu-(Mo-Au) deposits, such as Bisbee, Arizona (Bryant and Metz, 1966), and Robinson (Westra, 1982; Seedorff et al., 1996). There are at least local examples of advanced argillic alteration and/or lodes in the quartz monzonitic-granitic porphyry Mo-Cu subclas at Turnley Ridge (Steele and Atkinson, 1984), the rhyolitic subclas porphyry Mo at multiple sites around Questa (Meyer and Leonardson, 1990), and the rhyolitic porphyry Sn subclas at Cerro Rico (Potosi) (Sillitoe et al., 1998).

As noted in an earlier section, the relative positions of the lode (or advanced argillic alteration with or without high-sulfidation–state mineralization) and disseminated environments are variable (e.g., Sillitoe, 1999). Lodes can be superimposed on, can immediately overlie, or can be restricted to shallow levels, well above the disseminated environment. In cases where the disseminated and lode styles overlap or the relationship between them is clear, their precious metal characteristics are similar. For example, Au-bearing disseminated deposits from the tonalite-granodioritic porphyry Cu-(Au-Mo) subclas are associated with high-sulfidation Au-Cu deposits; such pairs include Far Southeast and Lepanto, the latter enargite-gold mineralization in the epithermal environment (Hedenquist et al., 1998), Frieda River and Nena, Papua New Guinea (Morrison et al., 1999), and the deep and shallow levels of Wafi, Papua New Guinea, with a strong vertical collapse of advanced argillic alteration due to rapid uplift (Sillitoe, 1990). In contrast, disseminated deposits from the Au-poor quartz monzodioritic-granitic porphyry Cu-(Mo) subclas are associated with Au-poor, Ag-bearing Cu lode deposits. Examples of such pairs include the pre-Main Stage and Main Stage lodes at Butte (Meyer et al., 1968), and disseminated and lode mineralization at Chiquicamata (Ossandon et al., 2001), Rosario in the Collluastui district, Chile (Masternak et al., 2004), and Resolution in the Superior district (Manske and Paul, 2002).

For the lodes that occur above any known disseminated mineralization, the nature of the underlying porphyry system, if present, is speculative. The Perruvian Au-poor lode deposits of epithermal affnity at Quinivieca (Bartos, 1987), Jucani (Deen et al., 1994), and Huanzala (Imai et al., 1985) may be related to deep, disseminated mineralization of the quartz monzodioritic-granitic porphyry Cu-(Mo) subclas. For the relatively Cu poor but Ag rich lodes, the associated disseminated deposit might be a quartz monzonitic-granitic porphyry Mo-Cu deposit. Examples include Equity Silver, British Columbia (Cyr et al., 1984; Wojdak and Sinclair, 1984) and Flatthead, Montana (Lange et al., 1994). The Tasna deposit, Bolivia, with lode Bi-W-Sn mineralization and associated advanced argillic alteration (Sillitoe et al., 1998), could be the top of a porphyry Sn system.

Links to other deposit types—continua versus gaps

Epithermal deposits: In the last decade, it has become increasingly clear that high-sulfidation deposits, many of epithermal character, are but one expression of the deeper porphyry environment and that they can be Cu, Au, or Ag rich. Although most recent interpretations generally do not link low-sulfidation deposits with porphyry systems, intermediate-sulfidation deposits are a more complicated issue (Sillitoe and Hedenquist, 2003). Most porphyry deposits have veins that formed late and/or in distal positions that can be termed intermediate sulfidation on the basis of the sulfide assemblage (Einaudi et al., 2003). In a few systems, these occurrences are
major orebodies. The best examples are the Mn-Ag-Zn veins in the peripheral zone of Butte (Meyer et al., 1968), and the carbonate-replacement bodies and monozone-hosted veins of Pb-Zn-Ag at Bingham (Rubright and Hart, 1968). In the case of Butte, high-sulfidation inner zones of Cu-Ag-Au grade outward into low-sulfidation Mn-Ag-Zn veins. There are other moderate-sized occurrences, such as the distal veins at Bajo de la Alumbrena (Proffett, 2003a) and certain Au-Ag orebodies that occur as limestone-replacement and fissure-vein deposits in the Robinson district, although there also are high-sulfidation and Carlin-like occurrences there (James, 1990; Seedorff, 2000). In these four districts, most of the intermediate-sulfidation mineralization is at or below the structural level of the porphyry system; i.e., these deposits are, for the most part, the late, lateral equivalents of the porphyry system—not the tops of the system. Intermediate-sulfidation epithermal base-metal-Au-telluride deposits are also closely associated with alkaline porphyry systems (e.g., Jensen and Barton, 2000).

All of the intermediate-sulfidation occurrences referred to are also right next to broadly coeval, high-temperature hydrothermal features such as mineralized porphyry or skarn. It is also remarkable that, of the systems that have been disseminated and tilted, many have no epithermal occurrences exposed. Relatively few porphyry deposits have intermediate-sulfidation precious metal ore deposits located in the vicinity; the Baguio and Mankayan districts, Philippines, are notable exceptions (Cooke and Bloom, 1990; Clavera et al., 1999). We suggest that, excluding the types of exceptions noted above, most low-sulfidation precious metal deposits do not have a link to porphyry systems and that further research is required to clarify the possible relationship between porphyries and intermediate-sulfidation deposits.

Distal disseminated gold deposits: There are distal disseminated Au-Ag or “Carlin-like” occurrences (Seedorff, 1991a; Cline et al., 2005) in some porphyry Cu districts; among these we include Barneys Canyon, Bingham district (Gunter et al., 1990; Cunningham et al., 2004); Ban, Malaysia (Sillitoe and Bonham, 1990), Purisima Concepcion, Yauricocha district, Peru (Alvarez and Noble, 1985), and Star Point and Knavish, Robinson district (Seedorff, 2000), and we would include them in the continua of features that form in porphyry systems. On the other hand, we interpret a genetic gap between the classic Carlin-type deposits in Nevada, such as Carlin, Getchell, and Jerritt Canyon, and porphyry deposits and related Carlin-like deposits (Cline et al., 2005), as Carlin-type deposits probably are regional—rather than intrusion-centered—hydrothermal systems (cf. Sillitoe and Bonham, 1990).

Challenges and applications

The differences between deposits still remain an important concern, especially for exploration geologists. For example, few major companies today may be interested in any of the known deposits in the quartz monzonitic-granitic porphyry Mo-Cu subclass (Table A2); geologists who are able to distinguish between this subclass and a quartz monzodioriticgranitic porphyry Cu-(Mo) system may have an advantage.

Genetic Understanding

In the first article in the first issue of the journal, Ransome (1905) referred to the latest swing in the pendulum in the battle between the Plutonists and the Neptunists just as porphyry deposits were coming onto the scene as major orebodies. Porphyry deposits came into their own while Lindgren and the Plutonists held sway. The geologists working on porphyry deposits—especially those in industry—have always held strong magmatist views (e.g., Sales, 1954; Westra and Keith, 1981; Sillitoe, 1995b). The outcrop-scale features linking porphyry plutons and hydrothermal features, coupled with the first-order correlation between igneous compositions and predominant metals in igneous-related deposits, kept porphyry geologists firmly in the Plutonist fold for the entire century. Beginning in the 1960s, geochemical studies of fluid inclusion compositions, melt-fluid experiments and theoretical modeling, and metal solubilities in aqueous solutions provided more support of the magmatic-hydrothermal interpretation (e.g., Roedder, 1971; Burnham, 1979; Hemley et al., 1992). For two decades or so beginning in the early 1970s, fluid-flow modelers and geochemists working on stable isotopes and fluid inclusions created a nonmagmatic alternative that drew adherents but has recently seen little support (see summary in Hedenquist and Richards, 1998).

Sources of components

Potassic alteration is produced by a hydrothermal fluid with a magmatic-dominated source, for which there is overwhelming geologic and geochemical evidence (e.g., Ulrich et al., 2001). Hydrolytic or acid alteration, which can form at deep, intermediate, and shallow depths, includes sericitic and local advanced argillic types, requires a magmatic source of fluids as supported by recent isotopic studies but may also include an important external, nonmagmatic source (Westra, 1982; Zaluski et al., 1994; Watanabe and Hedenquist, 2001; Harris and Golding, 2002; Field et al., 2005). In some deposits, even late intermediate argillic alteration and base metal veins can be the product of saline magmatic waters (e.g., Seedorff and Einaudi, 2004b). Although propylitic alteration has long been hypothesized to be the distal equivalent of potassic fluids of magmatic origin (e.g., Gustafson and Hunt, 1975), isotopic data from Bingham (Bowman et al., 1987) and geologic relationships at Bajo de la Alumbrena (Proffett, 2003a) and Henderson (Seedorff and Einaudi, 2004a) indicate that—at least in some deposits—propylitic alteration is formed by an influx of external fluids, though probably with a magmatic volatile fraction. Geologic and isotopic data support production of sodic-calcic and associated propylitic alteration by nonmagmatic brines (Dilles et al., 1992, 1995).

The ultimate source of ore metals, water, sulfur, chlorine, and other ore constituents in the porphyry magmas, however, involves lithospheric-scale processes. The upper mantle, subducting slab, lower to mid-crust, as well as the upper crust contribute components to the magma. The varying proportions of these sources likely lead to the observed compositional diversity of ore-forming magmas. For example, Cu, Au, and platinum-group elements may have predominantly mantle sources, whereas Mo and Pb may be contributed by subducting slab or crust. Certain magmatic components, such as sulfur, likely have diverse sources that vary between deposits (Ohmoto and Rye, 1979), with proposals for both mafic magma sources (Hattori and Keith, 2001) and crustal assimilation sources (Field et al., 2005).
The formation of porphyry deposits begins with construction of an intermediate to silicic magma chamber, the composition of which can vary widely (Fig. 8). This process is initiated with generation of mantle-sourced mafic magmas aided by subduction-derived fluids or asthenospheric rise. Mafic magma rise into the crust, mix and hybridize with crustal partial melts, and undergo fractional crystallization (e.g., Hildreth, 1981). The observation that porphyries result from the late stages of magmatic activity (e.g., Lang and Titley, 1998) in part reflects the likelihood that large, integrated upper crustal magma chambers capable of generating major porphyry deposits (e.g., Dilles, 1987; Cline and Bodnar, 1991) are not constructed until late stages; many arc volcanic systems probably never do so (Hildreth and Lanphere, 1994). Through part of the lifetime of a magmatic center, the chamber operates as an open system—degassing, erupting, recharging, crystallizing, assimilating, and conningling with magmas of various compositions (e.g., Eichelberger et al., 2000). Volumetrically minor magmas can play a key role by contributing trace constituents or triggering the extraction of an aqueous phase from the magma, enabling fluids and magma to ascend from the cupola.

Formation of a large porphyry deposit requires a series of events within an upper-crustal magma chamber. These include magmatic volatile exsolution to produce a separate aqueous phase, which must scavenge metals from a sufficient volume of silicate liquid and magmatic sulfides, and then gather in a magmatic cupola prior to its ascent to the site of ore deposition (cf. Candela and Piccoli, 2005; Fig. 14). The bulk composition of the magma, the water content of the magma, and depth are important variables in this regard (e.g., Cline, 1995) but so are geologic factors such as the geometry of the cupola and the homogeneity (convecting vs. nonconvecting) of the chamber (e.g., Dilles, 1987; Shinohara and Hedenquist, 1997).

Active volcanoes such as Pinatubo have ejected more sulfur than could be dissolved in the volume of erupted magma, and this “excess sulfur” (e.g., Keppler, 1999) is evidence that the eruption entrained a separate, sulfur-rich aqueous phase (Hattori, 1993) that is now argued to have a mafic magma source (Hattori and Keith, 2001). One of the key requirements is to transport the aqueous phase to the site of ore deposition without causing an eruption, but in some cases eruption may create structures favorable for later porphyry ore formation.

Multiple batches of porphyry magma intrude as dikes or plugs into the ore zone of most deposits over a short period of time (<300,000 yr; see above). The relative positions of early mineralized versus intermineral or postmineral intrusions produce the geometry of the orebody and the distribution of grade within it (Fig. 14). The rate and mass of intrusion and flux of various hydrothermal fluids dictate the thermal evolution of the system, i.e., heating versus cooling, which in turn produce the system-scale space-time distribution of ore minerals, veins, and wall-rock alteration features of the deposit. When later batches of magma arrive into a preexisting hydrothermal system, small amounts of various types of fluids can be present in pores and fractures. During magma

![Fig. 14. Schematic cross section of a porphyry deposit emplaced into older granitic rocks, which illustrates geometry and timing of principal intrusions, wall-rock alteration zones, and inferred hydrothermal sources and flow paths. The geometry of the orebody and alteration zones are typical of shallowly emplaced deposits, whereas deeper deposits have more uniform aspect ratios. Sulfide contents are generally highest in the pyrite-rich sericite zone, lower in potassic and ore zones, and lowest in propylitic and sodic-calcic zones. Erosion surface marks typical upper limit of exposures, above which alteration may be weak in most porphyry Mo and W deposits as well as in porphyry Cu deposits that are emplaced deeply or in reactive wall rocks. A volcanic edifice with characteristic intense quartz-albite alteration is illustrated schematically for shallowly emplaced porphyry Cu and Au systems and corresponds to the epithermal high-sulfidation environment.](image-url)
emplacement, magmatic-hydrothermal fluids dominate but also create hydraulic fractures that produce hydros_params compatible with external fluid flow. The external fluid may be formation waters of variable salinity; earlier magmatic-hydrothermal fluids that had cooled, depressurized, and changed composition since their introduction; dilute meteoric water; seawater; or mixtures of the aforementioned waters. Field and theoretical evidence suggests that in the high-temperature environment (~350°C), only dense brines such as formation waters or magmatic brines may gain access via hydrostatic circulation to the center part of the hydrothermal system (cf. Fournier, 1999).

The introduction of a new batch of magma, whether or not it is accompanied by release of magmatic-hydrothermal fluids, could result in formation of breccias and hydraulic fractures (e.g., Burnham, 1985; Fournier, 1999). Brecciation is a strong function of the depth of the system and the degree to which the fluids are confined or effectively open to the surface. These scenarios introduce the likelihood of a continuum of breccia types in the porphyry environments.

Numerous factors produce variability in porphyry hydrothermal systems that result in deposition of characteristic ore minerals, vein patterns, and wall-rock alteration. The characteristic system-scale time-space evolution and zoning of ores, veins, and alteration minerals (Figs. 6, 9, 10, 14, App. 4) are produced by changes along geochronological evolutionary pathways as a function of time, sulfidation state (Fig. 11), acidity, and activity or concentration ratios (Fig. 12). The diverse geochronological pathways of deposits are caused principally by (1) the initial magmatic fluid composition as dictated by magmatic compositions and processes; (2) hydrologic factors related to the effectiveness of magmatic-hydrothermal fracturing, depth of emplacement, permeability of host rocks, concurrent structural events, and the steepness of pressure and temperature gradients; (3) wall-rock composition; (4) cumulative water-rock mass ratios; and (5) degree and timing of interaction of magmatic fluids with external fluids. Despite the advances in understanding, the diverse processes that form economically important porphyry deposits are still incompletely understood today and remain fertile grounds for research.

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REFERENCES


PORPHORY DEPOSITS: CHARACTERISTICS AND ORIGIN OF HYPOGENE FEATURES


—1927, Relations of the disseminated copper ore in porphyry to igneous intrusive; American Institute of Mining and Metallurgical Engineers Transactions, v. 75, p. 797–815.

Porphyry Deposits: Characteristics and Origin of Hypogene Features


Seedorff, E., 1988, Cyclic development of hydrothermal mineral assemblages related to multiple intrusions at the Henderson porphyry molybdenum deposit, Colorado: Canadian Institute of Mining and Metallurgy Special Volume 39, p. 367–383.


—, 1993, Alteration logged in core holes from the Liberty pit, metal and alteration zoning in the Ruth-Ruth Extension are, and implications for structural reconstructions: Geological Society of America Special Publication 17, p. 21–51.


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