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Porphyry Copper Systems*

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Abstract

Porphyry Cu systems host some of the most widely distributed mineralization types at convergent plate boundaries, including porphyry deposits centered on intrusions; skarn, carbonate-replacement, and sediment-hosted Au deposits in increasingly peripheral locations; and superjacent high- and intermediate-sulfidation epithermal deposits. The systems commonly define linear belts, some many hundreds of kilometers long, as well as occurring less commonly in apparent isolation. The systems are closely related to underlying composite plutons, at paleodepths of 5 to 15 km, which represent the supply chambers for the magmas and fluids that formed the vertically elongate (>3 km) stocks or dike swarms and associated mineralization. The plutons may erupt volcanic rocks, but generally prior to initiation of the systems. Commonly, several discrete stocks are emplaced in and above the pluton roof zones, resulting in either clusters or structurally controlled alignments of porphyry Cu systems. The rheology and composition of the host rocks may strongly influence the size, grade, and type of mineralization generated in porphyry Cu systems. Individual systems have life spans of ~100,000 to several million years, whereas deposit clusters or alignments as well as entire belts may remain active for 10 m.y. or longer.

The alteration and mineralization in porphyry Cu systems, occupying many cubic kilometers of rock, are zoned outward from the stocks or dike swarms, which typically comprise several generations of intermediate to felsic porphyry intrusions. Porphyry Cu ± Au ± Mo deposits are centered on the intrusions, whereas carbonate wall rocks commonly host proximal Cu-Au skarns, less common distal Zn-Pb and/or Au skarns, and, beyond the skarn front, carbonate-replacement Cu and/or Zn-Pb-Ag ± Au deposits, and/or sediment-hosted (distal-disseminated) Au deposits. Peripheral mineralization is less conspicuous in noncarbonate wall rocks but may include base metal- or Au-bearing veins and mantos. High-sulfidation epithermal deposits may occur in lithocaps above porphyry Cu deposits, where massive sulfide lodes tend to develop in deeper feeder structures and Au ± Ag-rich, disseminated deposits within the uppermost 500 m or so. Less commonly, intermediate-sulfidation epithermal mineralization, chiefly veins, may develop on the peripheries of the lithocaps. The alteration-mineralization in the porphyry Cu deposits is zoned upward from barren, early sodic-calcic through potentially ore-grade potassic, chlorite-sericite, and sericitic, to advanced argillic, the last of these constituting the lithocaps, which may attain >1 km in thickness if unaffected by significant erosion. Low sulfidation-state chalcopyrite ± bornite assemblages are characteristic of potassic zones, whereas higher sulfidation-state sulfides are generated progressively upward in concert with temperature decline and the concomitant greater degrees of hydrolytic alteration, culminating in pyrite ± enargite ± covellite in the shallow parts of the lithocaps. The porphyry Cu mineralization occurs in a distinctive sequence of quartz-bearing veinlets as well as in disseminated form in the altered rock between them. Magmatic-hydrothermal breccias may form during porphyry intrusion, with some of them containing high-grade mineralization because of their intrinsic permeability. In contrast, most phreatomagmatic breccias, constituting maar-diatreme systems, are poorly mineralized at both the porphyry Cu and lithocap levels, mainly because many of them formed late in the evolution of systems.

Porphyry Cu systems are initiated by injection of oxidized magma saturated with S- and metal-rich, aqueous fluids from cupolas on the tops of the subjacent parental plutons. The sequence of alteration-mineralization events charted above is principally a consequence of progressive rock and fluid cooling, from >700° to <250°C, caused by solidification of the underlying parental plutons and downward propagation of the lithostatic-hydrostatic transition. Once the plutonic magmas stagnate, the high-temperature, generally two-phase hypersaline liquid and vapor responsible for the potassic alteration and contained mineralization at depth and early overlying advanced argillic alteration, respectively, gives way, at <350°C, to a single-phase, low- to moderate-salinity liquid that causes the sericite-chlorite and sericitic alteration and associated mineralization. This same liquid also causes mineralization of the peripheral parts of systems, including the overlying lithocaps. The progressive thermal decline of the systems combined with synmineral paleosurface degradation results in the characteristic overprinting (telescoping) and partial to total reconstitution of older by younger alteration-mineralization types. Meteoric water is not required for formation of this alteration-mineralization sequence although its late ingress is commonplace.

Many features of porphyry Cu systems at all scales need to be taken into account during planning and execution of base and precious metal exploration programs in magmatic arc settings. At the regional and district scales, the occurrence of many deposits in belts, within which clusters and alignments are prominent, is a powerful exploration concept once one or more systems are known. At the deposit scale, particularly in the porphyry Cu environment, early-formed features commonly, but by no means always, give rise to the best orebodies. Late-stage alteration overprints may cause partial depletion or complete removal of Cu and Au, but metal concentration may also result. Recognition of single ore deposit types, whether economic or not, in porphyry Cu systems may be directly employed in combination with alteration and metal zoning concepts to

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search for other related deposit types, although not all those permitted by the model are likely to be present in most systems. Erosion level is a cogent control on the deposit types that may be preserved and, by the same token, on those that may be anticipated at depth. The most distal deposit types at all levels of the systems tend to be visually the most subtle, which may result in their being missed due to overshadowing by more prominent alteration-mineralization.

Introduction

PORPHYRY Cu systems are defined as large volumes (10–>100 km³) of hydrothermally altered rock centered on porphyry Cu stocks that may also contain skarn, carbonate-replacement, sediment-hosted, and high- and intermediate-sulfidation epithermal base and precious metal mineralization. Along with calc-alkaline batholiths and volcanic chains, they are the hallmarks of magmatic arcs constructed above active subduction zones at convergent plate margins (Sillitoe, 1972; Richards, 2003), although a minority of such systems occupies postcollisional and other tectonic settings that develop after subduction ceases (e.g., Richards, 2009). The deeper parts of porphyry Cu systems may contain porphyry Cu ± Mo ± Au deposits of various sizes (<10 million metric tons [Mt]-10 billion metric tons [Gt]) as well as Cu, Au, and/or Zn skarns (<1 Mt–>1 Gt), whereas their shallower parts may host high- and intermediate-sulfidation epithermal Au ± Ag ± Cu orebodies (<1 Mt–>1 Gt). Porphyry Cu systems were generated worldwide since the Archean, although Meso-Cenozoic examples are most abundantly preserved (e.g., Singer et al., 2008; Fig. 1), probably because younger arc terranes are normally the least eroded (e.g., Seedorff et al., 2005; Kesler and Wilkinson, 2006; Wilkinson and Kesler, 2009).

Porphyry Cu systems presently supply nearly three-quarters of the world's Cu, half the Mo, perhaps one-fifth of the Au, most of the Re, and minor amounts of other metals (Ag, Pd, Te, Se, Bi, Zn, and Pb). The systems also contain major resources of these metals as well as including the world's largest known exploitable concentrations of Cu (203 Mt: Los Bronces-Río Blanco, central Chile; A.J. Wilson, writ. commun., 2009) and Mo (2.5 Mt: El Teniente, central Chile; Camus, 2003), and the second largest of Au (129 Moz: Grasberg, including contiguous skarn, Indonesia; J. MacPherson,

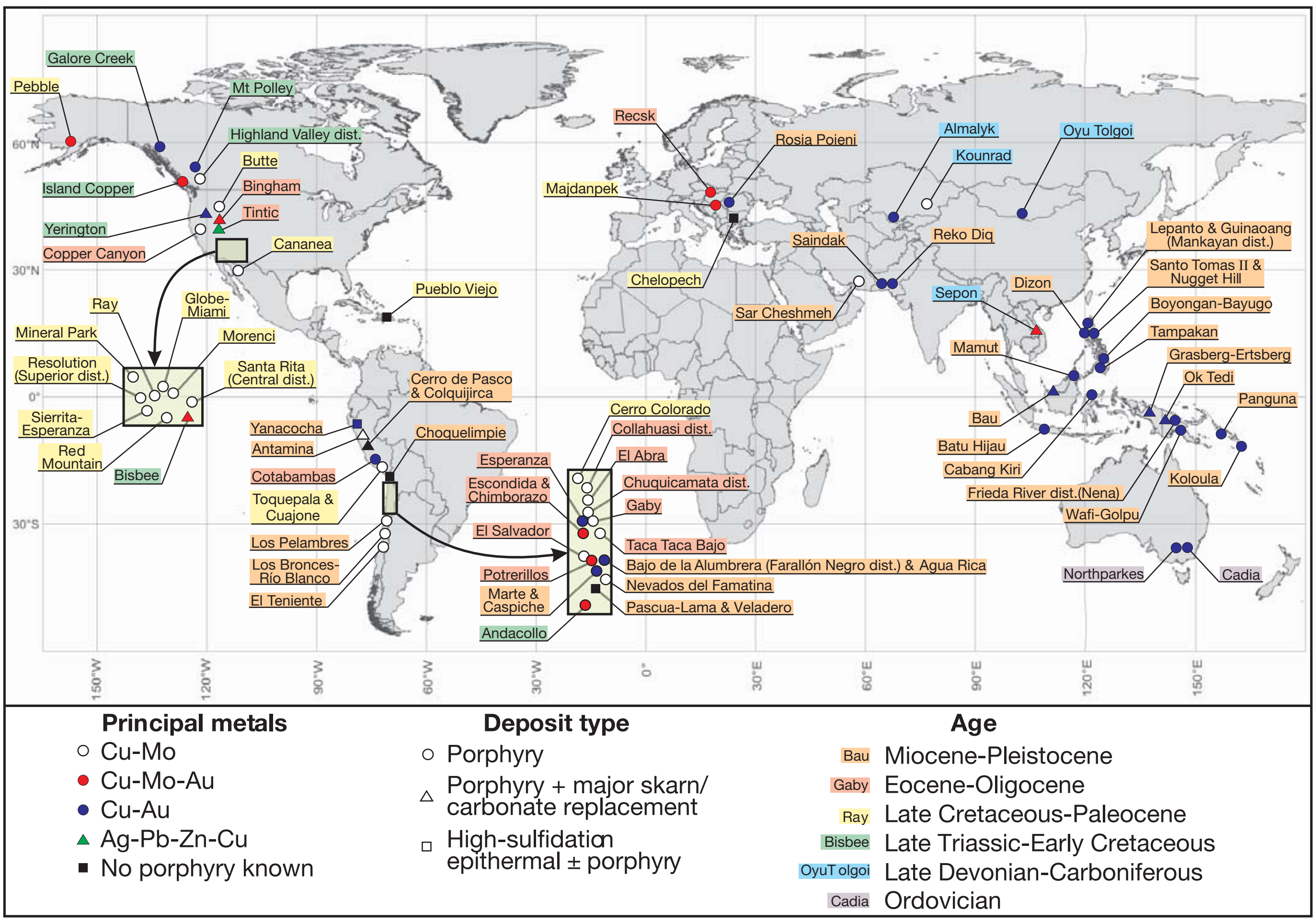

FIG. 1. Worldwide locations of porphyry Cu systems cited as examples of features discussed in the text along with five additional giant examples. The principal deposit type(s), contained metals, and age are also indicated. Data mainly from sources cited in the text.

writ. commun., 2009). Typical hypogene porphyry Cu deposits have average grades of 0.5 to 1.5 percent Cu, <0.01 to 0.04 percent Mo, and 0.0× to 1.5 g/t Au, although a few "Au-only" deposits have Au tenors of 0.9 to 1.5 g/t but little Cu (<0.1 %). The Cu and, in places, Au contents of skarns are typically higher still. In contrast, large high-sulfidation epithermal deposits average 1 to 3 g/t Au but have only minor or no recoverable Cu, commonly as a result of supergene removal.

This field-oriented article reviews the geology of porphyry Cu systems at regional, district, and deposit scales. The resultant geologic model is then used as the basis for a brief synthesis of porphyry Cu genesis and discussion of exploration guidelines. The deposits and prospects used as examples throughout the text are located and further characterized in Figure 1. The economically important results of supergene oxidation and enrichment in porphyry Cu systems have been addressed elsewhere (Sillitoe, 2005, and references therein).

Regional- and District-Scale Characteristics

Belts and provinces

Porphyry Cu systems show a marked tendency to occur in linear, typically orogen-parallel belts, which range from a few tens to hundreds and even thousands of kilometers long, as exemplified by the Andes of western South America (Sillitoe and Perelló, 2005; Fig. 2) and the Apuseni-Banat-Timok-Srednogorie belt of Romania, Serbia, and Bulgaria (Janković, 1977; Popov et al., 2002). Deposit densities commonly attain 15 per 100,000 km^2 of exposed permissive terrane (Singer et al., 2005). Each belt corresponds to a magmatic arc of broadly similar overall dimensions. One or more subparallel belts constitute porphyry Cu or epithermal Au provinces, several of which give rise to global-scale anomalies for Cu (e.g., northern Chile-southern Peru, southwestern North America) or Au (northern Peru; Sillitoe, 2008). Notwithstanding the ubiquity of porphyry Cu belts, major deposits may also occur in isolation or at least as distant outliers of coherent belts and provinces (e.g., Pebble in Alaska, Butte in Montana, and Bingham in Utah; Sillitoe, 2008; Fig. 1). Pueblo Viejo in the Dominican Republic (Fig. 1) is the best example of a major, isolated high-sulfidation epithermal Au deposit, albeit with no currently known porphyry Cu counterpart.

Porphyry Cu belts developed during well-defined metallogenic epochs, which isotopic dating shows to have typical durations of 10 to 20 m.y. Each porphyry Cu epoch is closely linked to a time-equivalent magmatic event. Again, the Andes (Sillitoe and Perelló, 2005), southwestern North America (Titley, 1993; Barra et al., 2005), and Apuseni-Banat-Timok-Srednogorie belt (Zimmerman et al., 2008) provide prime examples. Individual porphyry Cu belts are commonly spatially separate rather than superimposed on one another, reflecting arc migration as a result of steepening or shallowing of subducted slabs between the individual magmatic-metallogenic epochs (e.g., Sillitoe and Perelló, 2005). The processes of subduction erosion and terrane accretion at convergent margins may assist with land- or trenchward migration of the arcs and contained porphyry Cu belts (e.g., von Huene and Scholl, 1991; Kay et al., 2005). Nevertheless, several temporally discrete porphyry Cu-bearing arcs may be superimposed on

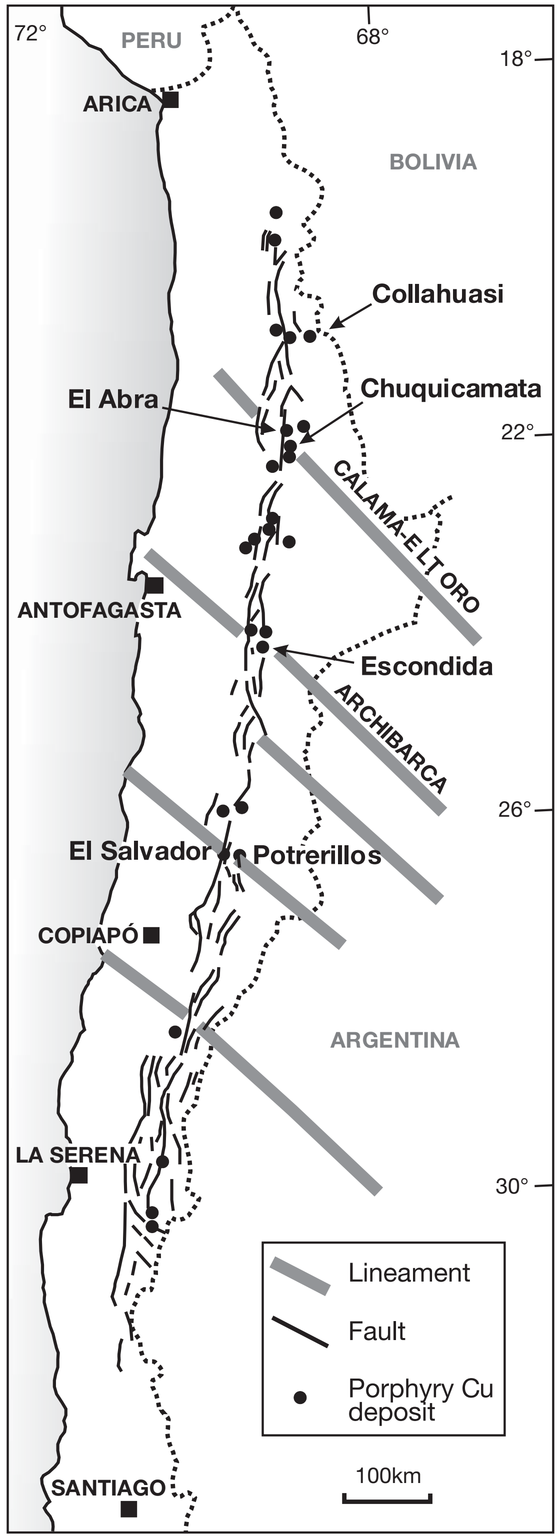

FIG. 2. A preeminent example of spatial and temporal coincidence between a porphyry Cu belt and an intra-arc fault zone: the northern Chile part of the central Andean middle Eocene to early Oligocene porphyry Cu belt and Domeyko fault system (summarized from Sillitoe and Perelló, 2005). The apparent termination of the belt in northernmost Chile is a result of concealment beneath Miocene volcanic rocks. Approximate positions of the main arc-transverse lineaments in northern Chile are also shown (after Salfity, 1985, in Richards et al., 2001).

one another: five since ~45 Ma in the Chagai belt, Pakistan (Perelló et al., 2008).

Tectonic settings

Porphyry Cu systems are generated mainly in magmatic arc (including backarc) environments subjected to a spectrum of regional-scale stress regimes, apparently ranging from moderately extensional through oblique slip to contractional (Tosdal and Richards, 2001). Strongly extensional settings, typified by compositionally bimodal basalt-rhyolite magmatism, lack significant porphyry Cu systems (Sillitoe, 1999a; Tosdal and Richards, 2001). The stress regime depends, among other factors, on whether there is trench advance or rollback and the degree of obliquity of the plate convergence vector (Dewey, 1980).

Nevertheless, there is a prominent empirical relationship between broadly contractional settings, marked by crustal thickening, surface uplift, and rapid exhumation, and large, high-grade hypogene porphyry Cu deposits, as exemplified by the latest Cretaceous to Paleocene (Laramide) province of southwestern North America, middle Eocene to early Oligocene (Fig. 2) and late Miocene to Pliocene belts of the central Andes, mid-Miocene belt of Iran, and Pliocene belts in New Guinea and the Philippines (Fig. 1; Sillitoe, 1998; Hill et al., 2002; Perelló et al., 2003a; Cooke et al., 2005; Rohrlach and Loucks, 2005; Sillitoe and Perelló, 2005; Perelló, 2006). Large, high-sulfidation epithermal Au deposits also form in similar contractional settings at the tops of tectonically thickened crustal sections, albeit not together with giant porphyry Cu deposits (Sillitoe and Hedenquist 2003; Sillitoe, 2008). It may be speculated that crustal compression aids development of large mid- to upper-crustal magma chambers (Takada, 1994) capable of efficient fractionation and magmatic fluid generation and release, especially at times of rapid uplift and erosional unroofing (Sillitoe, 1998), events which may presage initiation of stress relaxation (Tosdal and Richards, 2001; Richards, 2003, 2005; Gow and Walshe, 2005). Changes in crustal stress regime are considered by some as especially favorable times for porphyry Cu and high-sulfidation epithermal Au deposit generation (e.g., Tosdal and Richards, 2001), with Bingham and Bajo de la Alumbrera, Argentina, for example, both apparently occupying such a tectonic niche (Presnell, 1997; Sasso and Clark, 1998; Halter et al., 2004; Sillitoe, 2008).

Faults and fault intersections are invariably involved, to greater or lesser degrees, in determining the formational sites and geometries of porphyry Cu systems and their constituent parts. Intra-arc fault systems, active before as well as during magmatism and porphyry Cu generation, are particularly important localizers, as exemplified by the Domeyko fault system during development of the preeminent middle Eocene to early Oligocene belt of northern Chile (Sillitoe and Perelló, 2005, and references therein; Fig. 2). Some investigators emphasize the importance of intersections between continent-scale transverse fault zones or lineaments and arc-parallel structures for porphyry Cu formation, with the Archibarca and Calama-El Toro lineaments of northern Chile (Richards et al., 2001; Fig. 2), the Lachlan Transverse Zone of New South Wales (Glen and Walshe, 1999), comparable features in New Guinea (Corbett, 1994; Hill et al., 2002), and the much wider (160 km) Texas lineament of southwestern North America (Schmitt, 1966) being oft-quoted examples. These transverse features, possibly reflecting underlying basement structures, may facilitate ascent of the relatively small magma volumes involved in porphyry Cu systems (e.g., Clark, 1993; Richards, 2000).

Deposit clusters and alignments

At the district scale, porphyry Cu systems and their contained deposits tend to occur as clusters or alignments that may attain 5 to 30 km across or in length, respectively. Clusters are broadly equidimensional groupings of deposits (e.g., Globe-Miami district, Arizona; Fig. 3a), whereas alignments are linear deposit arrays oriented either parallel or transverse to the magmatic arcs and their coincident porphyry Cu belts. Arc-parallel alignments may occur along intra-arc fault zones, as exemplified by the Chuquicamata district, northern Chile (Fig. 3b) whereas cross-arc fault zones or lineaments control arc-transverse alignments, as in the Cadia, New South Wales (Fig. 3c) and Oyu Tolgoi, Mongolia districts (Fig. 3d).

Irrespective of whether the porphyry Cu systems and contained deposits define clusters or alignments, their surface distributions are taken to reflect the areal extents of either underlying parental plutons or cupolas on their roofs. Within the clusters and alignments, the distance (100s–1,000s m) between individual deposits (e.g., Sillitoe and Gappe, 1984) and even their footprint shapes can vary greatly, as observed in the Chuquicamata and Cadia districts (Fig. 3b, c).

Clusters or alignments of porphyry Cu systems can display a spread of formational ages, which attain as much as 5 m.y. in the Chuquicamata (Ballard et al., 2001; Rivera and Pardo, 2004; Campbell et al., 2006) and Yanacocha districts (Longo and Teal, 2005) but could be as much as ~18 m.y. in the Cadia district (Wilson et al., 2007). This situation implies that the underlying parental plutons have protracted life spans, albeit intermittent in some cases, with porphyry Cu formation taking place above them at different places over time.

Pluton-porphyry relationships

Varied relationships are observed between porphyry Cu systems and precursor plutons, which are typically multiphase, equigranular intrusions, commonly of batholithic dimensions and dioritic to granitic compositions; they are not only spatially, but also temporally and probably genetically related to porphyry Cu and superjacent epithermal Au formation (Fig. 4). The precursor plutons may act as hosts to a single deposit, as at Mount Polley, British Columbia (Fraser et al., 1995); an alignment of coalesced deposits, as in the Los Bronces-Río Blanco district (Fig. 5a); or clusters of two or more discrete deposits, as in the El Abra intrusive complex, northern Chile (Fig. 5b) and Guichon Creek batholith, Highland Valley district, British Columbia (Fig. 5c). The precursor plutons and porphyry Cu stocks are typically separated by time gaps of 1 to 2 m.y. or less (e.g., Dilles and Wright, 1988; Casselman et al., 1995; Mortensen et al., 1995; Dilles et al., 1997; Deckart et al., 2005; Campbell et al., 2006). Many porphyry Cu systems, particularly those that are only shallowly exposed, lack known precursor plutons, probably because they lie at inaccessible depths (Fig. 4).

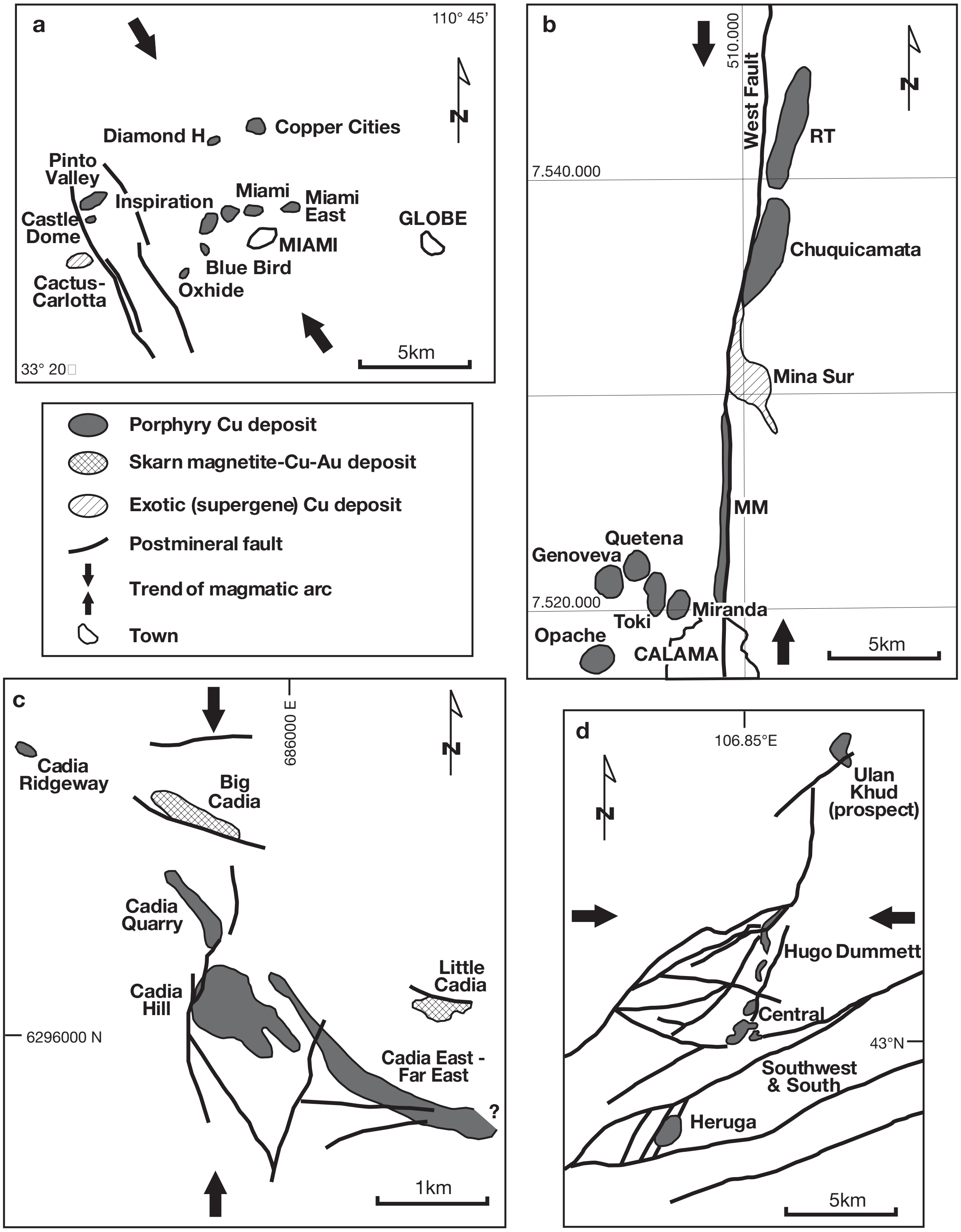

FIG. 3. Examples of porphyry Cu clusters and alignments of various sizes and at different orientations with respect to the axes of contemporaneous magmatic arcs. a. Globe-Miami district cluster, Arizona within the Late Cretaceous-early Tertiary (Laramide) arc (after Creasey, 1980), with the spatial distribution partially the result of mid-Tertiary extensional tectonism (Wilkins and Heidrick, 1995; Seedorff et al., 2008). b. Chuquicamata district, northern Chile aligned parallel to the middle Eocene-early Oligocene arc axis (after Rivera and Pardo, 2004; S. Rivera, writ. commun., 2009), with the spatial distribution possibly partly the result of postmineral sinistral strike-slip faulting (Brimhall et al., 2006). c. Cadia district, New South Wales, Australia, aligned oblique to the Ordovician arc axis (after Holliday et al., 2002). d. Oyu Tolgoi district, Mongolia aligned nearly perpendicular to the Late Devonian arc axis (after Khashgerel et al., 2008). Porphyry Cu and other deposit outlines projected to surface where unexposed. Note scale difference between c and a, b, and d.

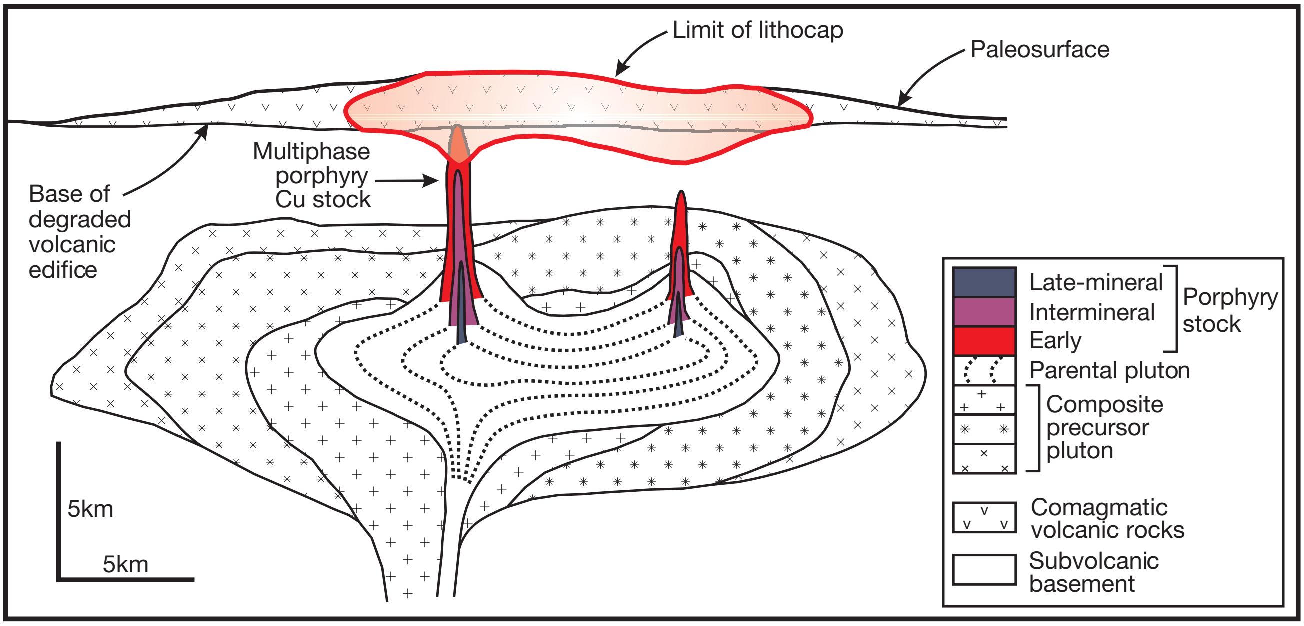

FIG. 4. Spatial relationships between porphyry Cu stocks, underlying pluton, overlying comagmatic volcanic rocks, and the lithocap. The precursor pluton is multiphase, whereas the parental pluton is shown as a single body in which the concentric dotted lines mark its progressive inward consolidation. The early, intermineral, and late-mineral phases of the porphyry Cu stocks, which span the interval during which the porphyry Cu deposits formed, originate from increasingly greater depths in the progressively crystallizing parental chamber. The volcanic sequence is a stratovolcano (but could just as readily be a dome complex; Fig. 6) and has been partly eroded prior to porphyry Cu formation. The lithocap affects the volcanic pile as well as uppermost parts of the underlying rocks. Note that subvolcanic basement rocks host much of the porphyry Cu deposit on the left, whereas that on the right is mainly enclosed by two phases of the precursor pluton. Inspired by Sillitoe (1973), Dilles (1987), Tosdal and Richards (2001), Casselman et al. (1995), and Dilles and Proffett (1995).

The precursor plutons are considered as the mid- to upper-crustal crystallization sites of mafic to felsic magmas that ascended from deeper reservoirs before porphyry Cu systems were developed (see Richards, 2003). Outcropping precursor plutons normally represent the shallower, earlier consolidated parts rather than the magma volumes from which the fluids for porphyry Cu generation were derived (Fig. 4). These parental magma chambers, also represented by similar equigranular to weakly porphyritic plutons, are not exposed in porphyry Cu systems unless postmineralization extensional tectonism caused profound tilting and dismemberment of the systems, as reconstructed in the Yerington district, Nevada (Dilles, 1987; Dilles and Proffett, 1995) and elsewhere (Seedorff et al., 2008).

Volcanic connections

Porphyry Cu systems may be spatially associated with comagmatic, calc-alkaline or, less commonly, alkaline volcanic rocks, typically of intermediate to felsic composition (Sillitoe, 1973; Fig. 4), which are generally erupted subaerially 0.5 to 3 m.y. prior to stock intrusion and mineralization, as well documented in the Bingham (Waite et al., 1997), Farallón Negro, Argentina (Sasso and Clark, 1998; Halter et al., 2004), Yerington (Dilles and Wright, 1988; Dilles and Proffett, 1995), Tampakan, Philippines (Rohrlach and Loucks, 2005), and Yanacocha (Longo and Teal, 2005) districts. However, the erosion involved in the unroofing of porphyry Cu deposits also severely degrades volcanic landforms (e.g., Farallón Negro district) and, commonly, entirely removes the eruptive products, at least in the general vicinities of the deposits themselves. Nevertheless, at a few localities, including the shallowly formed Marte porphyry Au deposit, northern Chile (Vila et al., 1991), a comagmatic andesitic stratovolcano is still partially preserved, including parts of its unmodified lower depositional slopes (or planèze). Notwithstanding their lower preservation potential, smaller volume volcanic centers—flow-dome complexes and maar-diatreme systems (e.g., Mankayan district, Philippines and Grasberg; Sillitoe and Angeles, 1985; MacDonald and Arnold, 1994; I. Kavalieris, pers. commun., 1999) —may still also be recognizable in the shallow parts of porphyry Cu systems. Volcanic landforms are obviously even better preserved in the shallower high-sulfidation epithermal environment above porphyry Cu deposits (e.g., flow-dome complexes at Yanacocha; Turner, 1999; Longo and Teal, 2005; e.g., Fig. 6).

Catastrophically explosive volcanism, particularly ash-flow caldera formation, is normally incompatible with synchronous porphyry Cu and superjacent epithermal Au deposit formation, because magmatic volatiles are dissipated during the voluminous pyroclastic eruptions rather than being retained and focused in a manner conducive to ore formation (Sillitoe, 1980; Pasteris, 1996; Cloos, 2001; Richards, 2005). Nevertheless, calderas may influence the localization of later, genetically unrelated porphyry Cu systems (e.g., El Salvador, northern Chile; Cornejo et al., 1997).

There is a strong suggestion that comagmatic volcanism may be inhibited in some major porphyry Cu belts as a result of their characteristic contractional tectonic settings, as in the

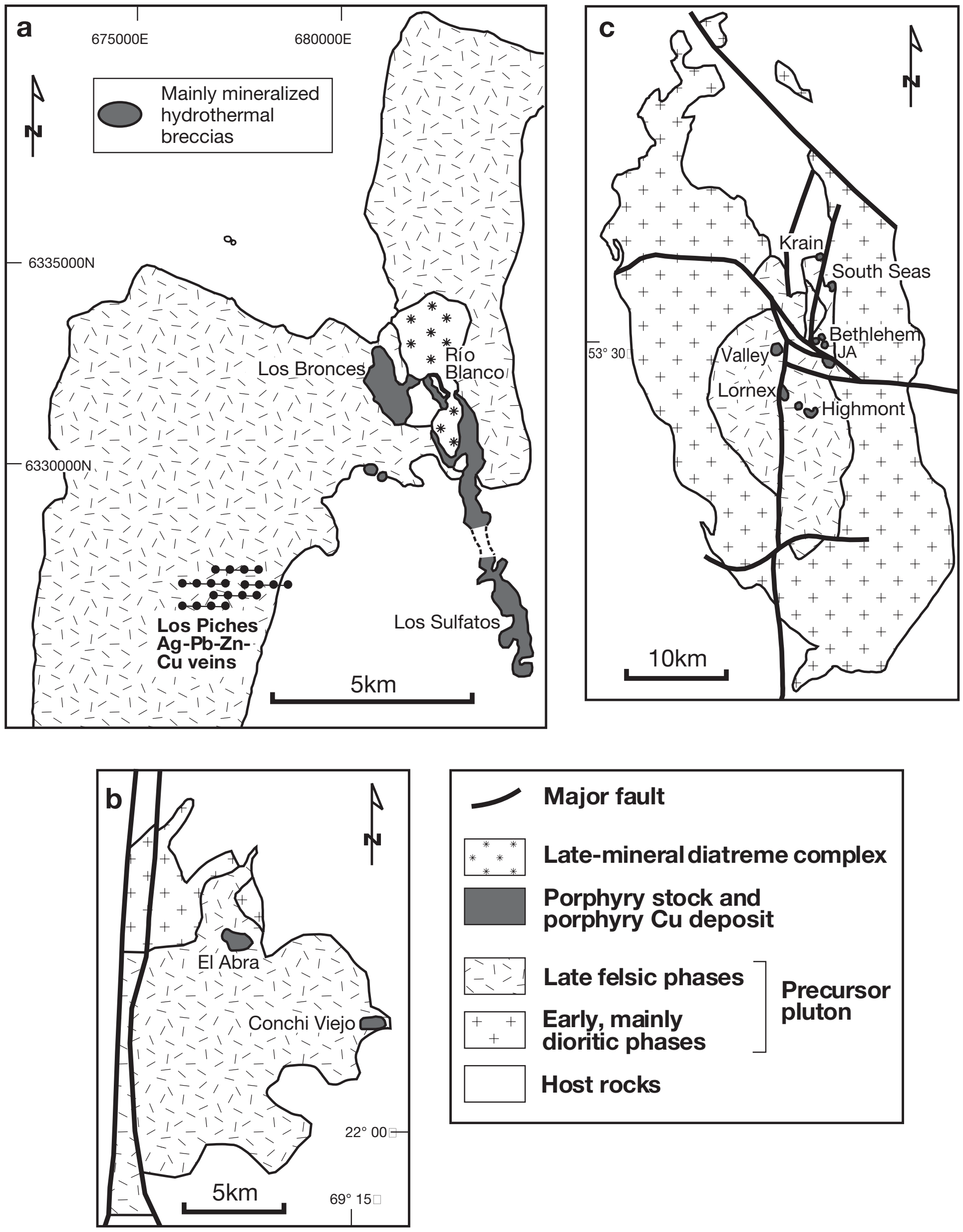

FIG. 5. Examples of porphyry Cu deposits within and near precursor plutons. a. Los Bronces-Río Blanco breccia-dominated deposit trending across the San Francisco batholith, central Chile (after Serrano et al., 1996; J.C. Toro, writ. commun., 2007). b. El Abra and Conchi Viejo deposits in the El Abra intrusive complex, northern Chile (after Dilles et al., 1997). c. Highland Valley deposit cluster in the Guichon Creek batholith, British Columbia (after Casselman et al., 1995). Note the variable positions of the deposits with respect to the exposed plutons, but their confinement to late felsic phases. Scales are different.

middle Eocene to early Oligocene belt of northern Chile, because of the tendency for subsurface magma accumulation in the absence of widely developed extensional faulting (Mpodozis and Ramos, 1990). The same situation is also apparent in several giant high-sulfidation epithermal Au deposits generated in thickened crust during tectonic uplift, such as Pascua-Lama and Veladero, northern Chile-Argentina, where the near absence of contemporaneous volcanism is more certain (Bissig et al., 2001; Charchaflié et al., 2007) given the much shallower erosion level, including partial paleosurface preservation (see below).

Wall-rock influences

Porphyry Cu systems are hosted by a variety of igneous, sedimentary, and metamorphic rocks (e.g., Titley, 1993), giving the initial impression of wall rocks playing a noninfluential role. It is becoming increasingly clear, however, that certain lithologic units may enhance grade development in both porphyry Cu and related deposit types.

Massive carbonate sequences, particularly where marble is developed near intrusive contacts, and other poorly fractured, fine-grained rocks have the capacity to act as relatively

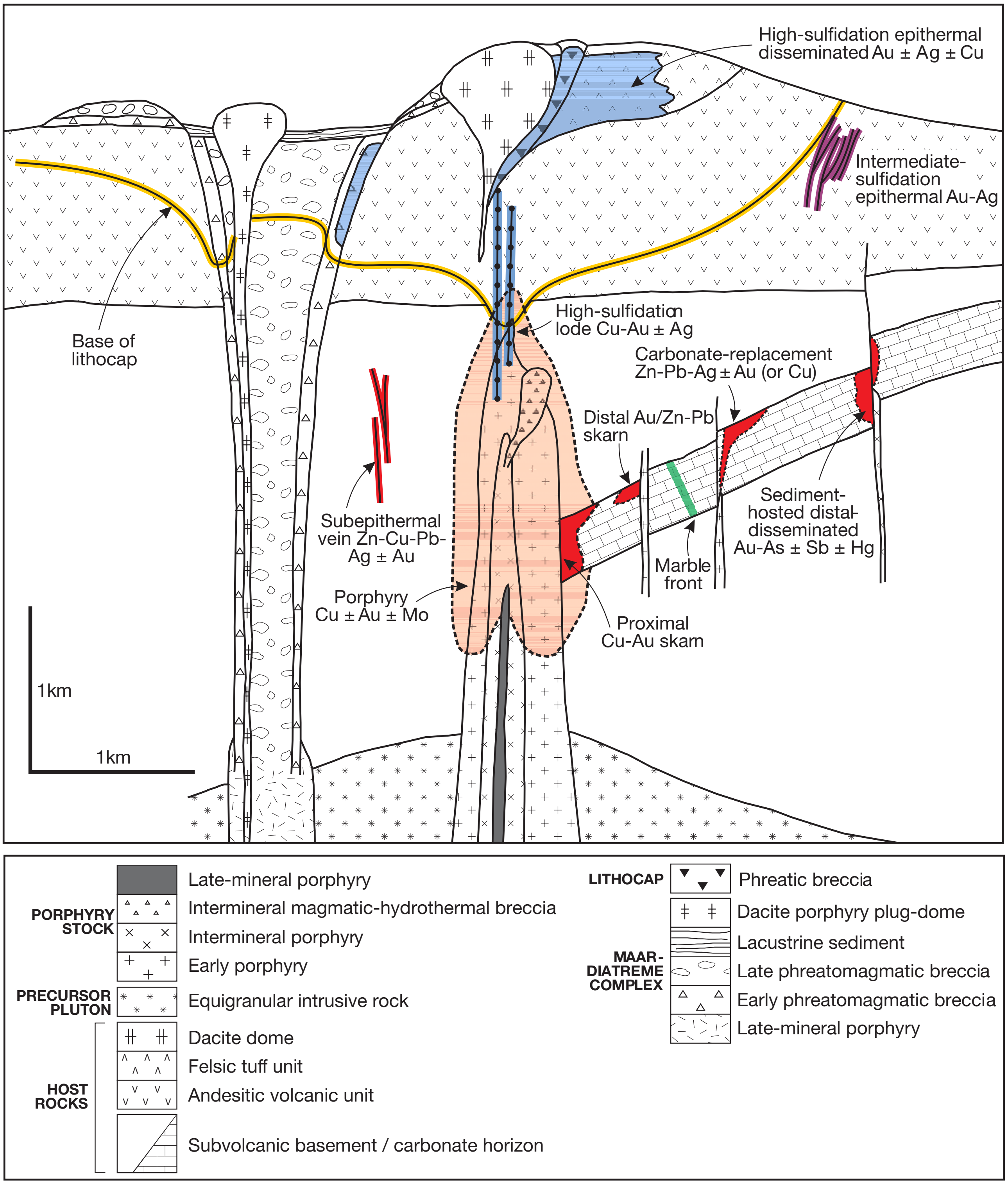

FIG. 6. Anatomy of a telescoped porphyry Cu system showing spatial interrelationships of a centrally located porphyry Cu ± Au ± Mo deposit in a multiphase porphyry stock and its immediate host rocks; peripheral proximal and distal skarn, carbonate-replacement (chimney-manto), and sediment-hosted (distal-disseminated) deposits in a carbonate unit and subepithermal veins in noncarbonate rocks; and overlying high- and intermediate-sulfidation epithermal deposits in and alongside the lithocap environment. The legend explains the temporal sequence of rock types, with the porphyry stock predating maar-diatreme emplacement, which in turn overlaps lithocap development and phreatic brecciation. Only uncommonly do individual systems contain several of the deposit types illustrated, as discussed in the text (see Table 3). Notwithstanding the assertion that cartoons of this sort (including Fig. 10) add little to the understanding of porphyry Cu genesis (Seedorff and Einaudi, 2004), they embody the relationships observed in the field and, hence, aid the explorationist. Modified from Sillitoe (1995b, 1999b, 2000).

impermeable seals around and/or above porphyry Cu deposits, resulting in high-grade ore formation (e.g., Grasberg; Sillitoe, 1997). Elsewhere, small-volume porphyry intrusions and the associated magmatic fluids fail to effectively penetrate low-permeability rock packages, leading to the apparently uncommon development of blind, high-grade deposits, as at Hugo Dummett in the Oyu Tolgoi district (Kirwin et al., 2003, 2005) and Ridgeway in the Cadia district (Wilson et al., 2003). High-sulfidation epithermal deposits may be similarly blind, beneath a thick limestone sequence in the case of Pueblo Viejo (Sillitoe et al., 2006).

Ferrous Fe-rich lithologic units also appear to favor high-grade porphyry Cu mineralization (e.g., Ray and Mineral Park, Arizona; Phillips et al., 1974; Wilkinson et al., 1982), presumably because of their capacity to effectively precipitate Cu transported in oxidized magmatic fluids (see below). It is unlikely coincidental that at least half the ore at three of the highest grade hypogene porphyry Cu deposits is hosted by such rocks: a gabbro-diabase-basalt complex at El Teniente (Skewes et al., 2002), a Proterozoic diabase sill complex at Resolution, Arizona (Ballantyne et al., 2003), and a tholeiitic basalt sequence in the Oyu Tolgoi district (Kirwin et al., 2005).

Mineralization elsewhere in porphyry Cu systems may be even more profoundly influenced by rock type. Proximal and distal skarn, carbonate-replacement, and sediment-hosted mineralization types are obviously dependent on the presence of reactive carbonate rocks, particularly thinly bedded, silty units. Large-tonnage, high-sulfidation epithermal deposits are favored by permeable rock packages, commonly pyroclastic or epiclastic in origin (e.g., Yanacocha; Longo and Teal, 2005), although disparate lithologic units can also prove receptive where extensively fractured (e.g., granitoid at Pascua-Lama; Chouinard et al., 2005).

Deposit-Scale Characteristics

Porphyry stocks and dikes

Porphyry Cu deposits are centered on porphyry intrusions that range from vertical, pluglike stocks (Fig. 6), circular to elongate in plan, through dike arrays to small, irregular bodies. The stocks and dikes commonly have diameters and lengths, respectively, of ≤1 km. However, much larger porphyry intrusions act as hosts in places, such as the elongate, 14-km-long stock at Chuquicamata-Radomiro Tomic (e.g., Ossandón et al., 2001; Fig. 3b) and the 4-km-long, <50-m-wide dike at Hugo Dummett (Khashgerel et al., 2008; Fig. 3d). Mining and deep drilling in a few large porphyry Cu deposits show that mineralized intrusions have vertical extents of >2 km (e.g., Chuquicamata and Escondida, northern Chile, and Grasberg) and, based on evidence from the steeply tilted systems, perhaps ≥4 km (Dilles, 1987; Seedorff et al., 2008; Fig. 6). The size of the stocks does not appear to bear any obvious relationship to the size of the associated porphyry Cu deposits and their Cu contents (cf. Seedorff et al., 2005). For example, the 12.5-Gt resource at Chuquicamata-Radomiro Tomic is confined to the 14-km-long stock referred to above (Ossandón et al., 2001; Camus, 2003), whereas perhaps only roughly 20 percent of the similarly sized El Teniente deposit and <10 percent of the 1.5-Gt El Abra deposit are hosted by the porphyry intrusions (Camus, 1975, 2003; Ambrus, 1977). The distal parts of porphyry Cu systems, beyond the porphyry Cu deposits, either lack porphyry intrusions or contain only relatively minor dikes (e.g., Virgin dike in the skarn-dominated Copper Canyon district, Nevada, and Yerington district skarn Cu occurrences; Wotruba et al., 1988; Dilles and Proffett, 1995).

The porphyry Cu-related intrusions comprise multiple phases (Kirkham, 1971; Gustafson, 1978), which were emplaced immediately before (early porphyries), during (intermineral porphyries), near the end of (late mineral porphyries), and after (postmineral porphyries) the alteration and mineralization events (Fig. 6). For example, seven phases are mapped at Bajo de la Alumbrera (Proffett, 2003), five at Yerington (Proffett, 2009), and four at Bingham (Redmond et al., 2001). The immediately premineral, early porphyries and their contiguous host rocks contain the highest grade mineralization in most deposits although, exceptionally, the earliest phase can be poorly mineralized (e.g., Grasberg; MacDonald and Arnold, 1994). Intermineral porphyries are typically less well mineralized as they become progressively younger, and late- and postmineral phases are barren. The earlier porphyry bodies are not destroyed when intruded by later phases but merely split apart, causing overall inflation of the rock package as would occur during ordinary dike emplacement. Several criteria, in addition to metal contents and ratios (Cu/Au/Mo) and intensity of veining, alteration, and mineralization, are used to distinguish the relative ages of porphyry intrusions: younger phases truncate veinlets in, are chilled against, and contain xenoliths of older phases (Fig. 7; Sillitoe, 2000). Commonly, the xenoliths are largely assimilated by the younger phases, leaving only the contained quartz veinlets, chemically more refractory than the host porphyry, as “floating” pieces (Fig. 7). Wall-rock xenoliths in the marginal parts of some porphyry intrusions may be sufficiently abundant to constitute intrusion breccias. The upper contacts of a few porphyry Cu intrusions are characterized by unidirectional solidification textures (USTs): alternating, crenulate layers of quartz and aplite and/or aplite porphyry produced as a result of pressure fluctuations at the transition from magmatic to hydrothermal conditions (e.g., Kirkham and Sinclair, 1988; Garwin, 2002; Lickfold et al., 2003; Cannell et al., 2005; Kirwin, 2005). However, USTs are not consistently developed and, hence, do not provide a reliable means of subdividing porphyry Cu intrusion phases.

The porphyry intrusions in porphyry Cu deposits are exclusively of I-type and magnetite-series affiliation (Ishihara, 1981), and typically metaluminous and medium K calc-alkaline, but may also fall into the high K calc-alkaline (shoshonitic) or alkaline fields (see Seedorff et al., 2005, for further details). They span a range of compositions from calc-alkaline diorite and quartz diorite through granodiorite to quartz monzonite (monzogranite), and alkaline diorite through monzonite to, uncommonly, syenite (e.g., Galore Creek, British Columbia; Enns et al., 1995). Mo-rich porphyry Cu deposits are normally tied to the more felsic intrusions, whereas Au-rich porphyry Cu deposits tend to be related to the more mafic end members, although intrusions as felsic as quartz monzonite may also host Au-rich examples (e.g., Mamut, East Malaysia; Kósaka and Wakita, 1978). However, Cu-poor porphyry Au

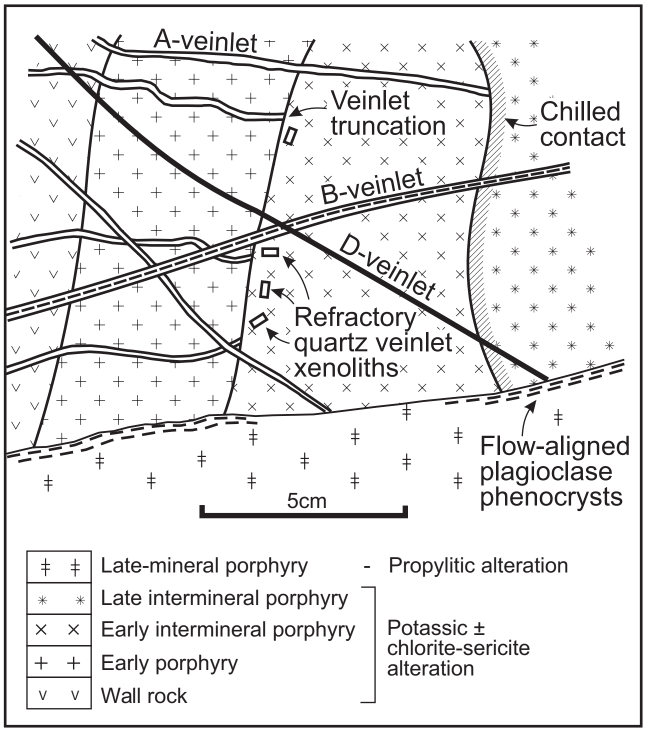

FIG. 7. Schematic crosscutting relationships between early (immediately premineral), intermineral, and late-mineral porphyry phases in porphyry Cu stocks and their wall rocks. Veinlet truncation, quartz veinlet xenoliths, chilled contacts, and flow-aligned phenocrysts as well as textural, grade, and metal-ratio variations may denote the porphyry contacts, albeit generally not all present at the same contact. Early A, B, and late D veinlets are explained in the text and Figure 13. Note that early A veinlets are more abundant in the early porphyry, less abundant in the early intermineral porphyry, and absent from the two later porphyry phases. The late-mineral porphyry lacks veinlets and displays only propylitic alteration. Modified from Sillitoe (2000).

deposits appear to occur exclusively in association with calc-alkaline diorite and quartz diorite porphyries (e.g., Vila and Sillitoe, 1991). The porphyry intrusions contain variable amounts of phenocrysts, typically including hornblende and/or biotite, and fine-grained, commonly aplitic groundmass, resulting in open to crowded textures. The distinctive aplitic groundmass texture is ascribed to pressure quenching during rapid ascent and consequent volatile loss (Burnham, 1967). The porphyry phases in some individual deposits may have clear compositional differences (e.g., Bajo de la Alumbrera; Proffett, 2003) and/or characteristic igneous textures (e.g., El Salvador; Gustafson and Hunt, 1975); however, particularly in many porphyry Au and Cu-Au deposits, the different phases are commonly only subtly different or nearly identical. Furthermore, textural obliteration is commonplace in the highly altered, early porphyry phases, thereby rendering them difficult to distinguish from volcanic wall rocks in some deposits (e.g., Galore Creek and Hugo Dummett).

Isotopic dating, using the U-Pb zircon method, suggests that the multiphase porphyry intrusions in porphyry Cu systems can be assembled in as little as 80,000 years (Batu Hijau, Indonesia; Garwin, 2002), but the process commonly takes much longer. Emplacement of the porphyry stocks in many central Andean deposits took from 2 to 5 m.y., implying that appreciable time (0.5–1.5 m.y.) elapsed between emplacement of the component phases (e.g., Ballard et al., 2001; Maksaev et al., 2004; Padilla-Garza et al., 2004; Jones et al., 2007; Perelló et al., 2007; Harris et al., 2008). Furthermore, there seems to be no obvious relationship between the size of porphyry Cu deposits and the duration of the intrusive activity, the latter seemingly being the main parameter defining the total hydrothermal life spans of porphyry Cu systems. Detailed geochronology of the high-sulfidation parts of some porphyry Cu systems also suggests extended life spans, 1 to >1.5 m.y. at Cerro de Pasco and Colquijirca, central Peru (Bendezú et al., 2008; Baumgartner et al., 2009). However, these life spans are orders of magnitude longer than the theoretically modeled times required for consolidation of individual porphyry intrusions (<40,000 yr; Cathles, 1977; Cathles et al., 1997), porphyry Cu ore formation (<100,000 yr; McInnes et al., 2005), or major potassic alteration events (<2,000 yr; Shinohara and Hedenquist, 1997; Cathles and Shannon, 2007).

Diatremes

Diatremes, upward-flared volcanic vents generated mainly by phreatomagmatic eruptive activity, are widespread in porphyry Cu systems (Sillitoe, 1985), including their shallow, epithermal parts (e.g., Yanacocha; Turner, 1999; Fig. 6). The diatremes, commonly ≥1 km in near-surface diameter and up to at least 2 km in vertical extent (e.g., >1.8 km preserved at the Braden diatreme, El Teniente; Howell and Molloy, 1960; Camus, 2003), can be manifested at the paleosurface by maar volcanoes: ephemerally lake-filled craters encircled by tuff rings (Fig. 6). Diatreme breccias have a distinctive texture, in which widely separated, typically centimeter-sized clasts are dominated by rock-flour matrix containing an andesitic to dacitic tuffaceous component (Table 1), the latter commonly difficult to recognize where alteration is intense. The poorly lithified, friable nature and clay-rich matrix of many diatreme breccias give rise to recessive topography and little, if any, surface exposure. A positive topographic expression results only where barren, late- to postmineral porphyry plugs intrude the diatreme breccias (e.g., Dizon and Guinaoang, Philippines and Batu Hijau; Sillitoe and Gappe, 1984; Sillitoe and Angeles, 1985; Garwin, 2002; Fig. 6).

Many diatremes are late-stage additions to porphyry Cu systems, in which they commonly postdate and either cut or occur alongside porphyry Cu mineralization at depth (Howell and Molloy, 1960; Sillitoe and Gappe, 1984; Perelló et al., 1998; Garwin, 2002) but overlap with high-sulfidation events at shallower epithermal levels (e.g., Dizon; Fig. 6). The diatremes, particularly their contact zones, may localize part of the high-sulfidation Au mineralization (e.g., Wafi-Golpu, Papua New Guinea; Fig. 6). In a minority of cases, however, diatremes (e.g., Grasberg, Galore Creek, and Boyongan-Bayugo, Philippines; MacDonald and Arnold, 1994; Enns et al., 1995; Braxton et al., 2008) or tuff-filled depressions presumably fed by one or more subjacent diatremes (e.g., Resolution) are early features that act as receptive wall rocks to the main alteration and mineralization.

Magmatic-hydrothermal and phreatic breccias

The most common hydrothermal breccias in the deeper parts of porphyry Cu systems are of magmatic-hydrothermal type, the products of release of overpressured magmatic

TABLE 1. Features of Principal Hydrothermal Breccia Types in Porphyry Cu Systems

Type	Position in system (abundance)	Form	Relative timing	Clast features	Matrix/cement	Clast/matrix proportions	Alteration types (Table 2)	Main Cu-bearing mineral(s)	Economic potential
Magmatic hydrothermal	Within porphyry Cu deposits, locally around them (ubiquitous)	Irregular, pipe-like bodies (10s–100s m in diam)	Typically intermineral	Commonly monomict, angular to subrounded	Quartz-magnetite-biotite-sulfides/quartz-muscovite-tourmaline-sulfides ± rock flour ± igneous rock (i.e., igneous breccia)	Clast or matrix supported	Potassic ± chlorite-sericite ± sericitic; uncommonly advanced argillic	Chalcopyrite, uncommonly bornite	May constitute ore, commonly high grade
Phreatic (porphyry Cu level)	Within and around porphyry Cu deposits (relatively common)	Dikes, uncommonly sills and irregular bodies	Late	Polymict, rounded to subrounded	Muddy rock flour	Matrix supported	Sericitic, advanced argillic, or none	Generally none	Barren unless rich in pre-existing mineralization (e.g., Bisbee; Bryant, 1987)
Phreatic (epithermal level)	Within lithocaps; local surface manifestations as eruption breccia (relatively common)	Irregular bodies (10s–100s m in diam)	Typically intermineral relative to lithocap development	Commonly silicified, angular to subrounded	Chalcedony, quartz, alunite, barite, sulfides, native S	Clast or matrix supported	Advanced argillic	Enargite, luzonite	May constitute high-sulfidation Cu/Au/Ag ore
Phreatomagmatic	Diatremes span porphyry Cu and epithermal environments; surface manifestations as maar volcanoes (present in ~20% of systems)	Kilometer-scale, downward-narrowing conduits	Commonly late, but early examples known	Polymict, centimeter-sized, rounded, and polished; juvenile (magma blob, pumice) clasts locally	Rock flour with juvenile tuff or magma blob component; early examples cut by porphyry Cu mineralization	Matrix dominated; accretionary lapilli in matrix-dominated layers	None or advanced argillic, but early examples with any alteration type depending on exposure level	Locally enargite	Commonly barren, but may host porphyry Cu or high-sulfidation ore types

fluids (Sillitoe, 1985). Many porphyry Cu deposits contain minor volumes (5–10%) of magmatic-hydrothermal breccia (Fig. 6); however, even major deposits can be either breccia free, as at Chuquicamata (Ossandón et al., 2001), or breccia dominated, as exemplified by >5 Gt of ore-grade breccia at Los Bronces-Río Blanco (Warnaars et al., 1985; Serrano et al., 1996; Fig. 5a). Magmatic-hydrothermal breccias display a variety of textures (Table 1), which are mainly dependent on clast form and composition, clast/matrix ratio, matrix/cement constitution, and alteration type. They are distinguished from the phreatomagmatic diatreme breccias by several features (Table 1), particularly the absence of tuffaceous material. The breccia clasts may be set in rock-flour matrix, hydrothermal cement, fine-grained igneous material, or some combination of the three. Igneous matrices tend to be more common at depth near the magmatic source, where the term igneous breccia is appropriately applied (e.g., Hunt et al., 1983; Fig. 8).

Magmatic-hydrothermal breccias, generally occupying steep, pipelike to irregular bodies, are commonly intermineral in timing as a result of being generated in close association with intermineral porphyry phases (Figs. 6, 8). Hence, many of the breccias overprint preexisting alteration-mineralization patterns and veinlet types (e.g., Red Mountain, Arizona; Quinlan, 1981), which are incorporated as clasts. Early breccias may display potassic alteration and have biotite, magnetite, and chalcopyrite cements, whereas later ones are commonly sericitized and contain prominent quartz, tourmaline, specularite, chalcopyrite, and/or pyrite as cementing minerals. Sericitized breccia may change downward to potassic-altered breccia (e.g., Los Bronces-Río Blanco; Vargas et al., 1999; Frikken et al., 2005; Fig. 8). The metal contents of some magmatic-hydrothermal breccias may be higher than those of surrounding porphyry Cu stockwork mineralization, reflecting their high intrinsic permeability. In contrast to diatremes, magmatic-hydrothermal breccias are normally blind and do not penetrate the overlying epithermal environment, whereas downward they gradually fade away as a result of increased clast/matrix ± cement ratios, in places accompanied by pods of coarse-grained, pegmatoidal, potassic minerals representing former sites of vapor accumulation (e.g., Los Pelambres, central Chile; Perelló et al., 2007; Fig. 8; see below).

Several types of phreatic (meteoric-hydrothermal) breccia are widely observed in porphyry Cu systems; they may be simply subdivided into pebble dikes and, uncommonly, larger bodies resulting from flashing of relatively cool ground water on approach to magma, typically late-mineral porphyry dikes; and steep, tabular to irregular bodies triggered by vapor-pressure buildup beneath impermeable layers, commonly resulting from self sealing by silicification (Sillitoe, 1985; Table 1). Hence, pebble dikes display downward transitions to porphyry intrusions (e.g., Tintic, Utah and Toquepala, southern Peru; Farmin, 1934; Zweng and Clark, 1995), whereas the breccias triggered by fluid confinement do not normally form in close proximity to intrusive bodies. The pebble dikes and related breccias are chiefly restricted to porphyry Cu deposits, including their marginal parts, whereas brecciation induced by the fluid confinement typifies the overlying high-sulfidation epithermal environment (Fig. 6). There, distinction from phreatomagmatic diatreme breccias may be difficult because of texture obliteration caused by intense advanced argillic alteration (e.g., Pascua-Lama).

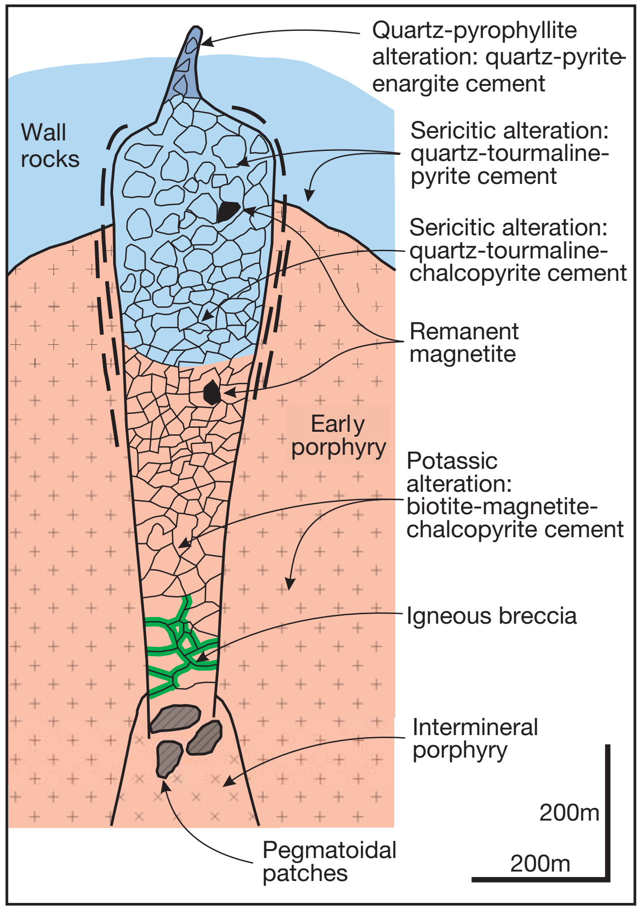

FIG. 8. Schematic depiction of a large magmatic-hydrothermal breccia body genetically linked to the apex of an intermineral porphyry intrusion. The alteration-mineralization is zoned from advanced argillic (with pyrite-enargite) at the top through sericitic (with shallow pyrite and deep chalcopyrite) to potassic (with magnetite-chalcopyrite ± bornite) at the bottom, where the breccia texture may be almost imperceptible and pegmatoidal pods are commonplace. Injection of fine-grained, igneous matrix defines the igneous breccia near the base of the body. The entire breccia body originally underwent potassic alteration prior to partial overprinting by sericitic and, subsequently, advanced argillic assemblages, as documented by localized occurrence of remanent magnetite and muscovite after coarse-grained biotite in the cement to the sericitized breccia.

Phreatic breccias, as exemplified by pebble dikes, normally contain polymictic clast populations set in muddy, rock-flour matrices (Table 1). Vertical clast transport may be appreciable (e.g., >1 km at Tintic; Morris and Lovering, 1979). The breccias are typically late-stage features and, hence, unaltered and barren. In contrast, clast transport in the phreatic breccias produced by fluid confinement in the high-sulfidation environment is more restricted, with many of the clasts being locally derived from the seals themselves and, hence, commonly composed of silicified rock (Table 1). Although rock flour may occur between the clasts, quartz, chalcedony, alunite, barite, pyrite, and enargite are also widely observed as cementing minerals. These phreatic breccias host ore in many high-sulfidation Au ± Ag ± Cu deposits (e.g., Choquelimpie, northern Chile; Gröpper et al., 1991). In contrast to the magmatic-hydrothermal breccias and pebble dikes at the porphyry Cu level, these phreatic breccias in the high-sulfidation environment may attain the paleosurface, where hydrothermal eruptions result in subaerial breccia accumulation as aprons around the eruptive vents (e.g., Hedenquist and Henley, 1985; Fig. 6).

Orebody types and geometries

The deeper, central cores of porphyry Cu systems are occupied by porphyry Cu deposits, in which ore-zone geometries depend mainly on the overall form of the host stock or dike complex, the depositional sites of the Cu-bearing sulfides, and the positions of any late, low- and subore-grade porphyry intrusions and diatremes. The forms of many porphyry Cu deposits mimic those of their host intrusions; thus, cylindrical stocks typically host cylindrical orebodies (Fig. 6), whereas laterally extensive dikes give rise to orebodies with similar narrow, elongate shapes (e.g., Hugo Dummett; Khashgerel et al., 2008). Many porphyry Cu deposits are formed as vertically extensive bodies, which become progressively lower grade both outward and downward, whereas others have a bell- or cap-like form because little Cu was precipitated internally at depth (e.g., Resolution; Ballantyne et al., 2003). The tops of the orebodies tend to be relatively abrupt and controlled by the apices of quartz veinlet stockworks (see below). The shape of any porphyry Cu orebody may undergo significant modification as a result of emplacement of late- to postmineral rock volumes (e.g., Fig. 5a), as exemplified by the low-grade cores caused by internal emplacement of late-mineral porphyry phases (e.g., Santo Tomas II, Philippines; Sillitoe and Gappe, 1984) and, much less commonly, late-stage diatremes (e.g., El Teniente; Howell and Molloy, 1960; Camus, 2003). A few deposits, instead of dying out either gradually (e.g., El Salvador; Gustafson and Quiroga, 1995) or fairly abruptly (e.g., H14-H15 at Reko Diq, Pakistan) at depth, have knife-sharp bases as a result of truncation by late-mineral intrusions (e.g., Santo Tomas II; Sillitoe and Gappe, 1984). Coalescence of closely spaced porphyry Cu deposits enhances size potential (e.g., H14-H15 at Reko Diq; Perelló et al., 2008)

Development of wall rock-hosted orebodies alongside porphyry Cu deposits is most common where receptive carbonate rocks are present (Fig. 6). Deposit types include proximal Cu ± Au and, less commonly, distal Au and/or Zn-Pb skarns (e.g., Meinert, 2000; Meinert et al., 2005); more distal, carbonate-replacement (chimney-manto), massive sulfide bodies dominated by either Cu (e.g., Superior district, Arizona and Sepon district, Laos [Fig. 9c]; Paul and Knight, 1995; Loader, 1999) or, more commonly, Zn, Pb, Ag ± Au (e.g., Recsk, Hungary; Kisvarsanyi, 1988) beyond the skarn front (Fig. 6); and, uncommonly, sediment-hosted (distal-disseminated; Cox and Singer, 1990) Au concentrations on the fringes of the systems (e.g., Barneys Canyon and Melco, Bingham district; Babcock et al., 1995; Gunter and Austin, 1997; Cunningham et al., 2004; Fig. 9a). Continuity between some of these carbonate rock-hosted deposits is possible; for example, transitions from proximal Cu-Au to distal Au skarn in the Copper Canyon district (Cary et al., 2000) and distal Zn-Pb-Cu-Ag skarn to carbonate-replacement Zn-Pb-Ag at Groundhog, Central district, New Mexico (Meinert, 1987). All these carbonate rock-hosted ore types are replacements of receptive beds, commonly beneath relatively impermeable rock units (e.g., Titley, 1996) and, hence, tend to be strata bound, although high- and low-angle fault control is also widely emphasized (e.g., proximal skarns at Ok Tedi, Papua New Guinea, and Antamina, central Peru; Rush and Seegers, 1990; Love et al., 2004).

Distal ore formation in porphyry Cu systems is less common in igneous or siliciclastic wall rocks, within propylitic halos, where fault- and fracture-controlled, subepithermal Zn-Pb-Cu-Ag ± Au veins of currently limited economic importance tend to be developed (e.g., Mineral Park; Eidel et al., 1968 and Los Bronces-Río Blanco; Figs. 5a, 6). Nevertheless, larger tonnage orebodies may occur where permeable wall rocks exist, as exemplified by the stacked, Au-bearing mantos in amygdaloidal and brecciated andesitic flow tops at Andacollo, Chile (Reyes, 1991).

In the lithocap environment—typically located above, areally more extensive than, and commonly overprinting porphyry Cu deposits (Fig. 6; see below)—high-sulfidation epithermal Au, Ag, and/or Cu deposits are characteristic; nevertheless, the preserved parts of many lithocaps are essentially barren. The deeper level high-sulfidation deposits, the Cordilleran base metal lodes of Einaudi (1982), tend to be characterized by massive sulfides, commonly rich in the Cu-bearing sulfosalts (enargite, luzonite, and/or famatinite). They commonly occur as tabular veins overprinting porphyry Cu deposits, like those at Butte (Meyer et al., 1968), Escondida (Ojeda, 1986), Chuquicamata (Ossandón et al., 2001), and Collahuasi, northern Chile (Masterman et al., 2005; Fig. 6). Alternatively, for up to several kilometers beyond porphyry Cu deposits, they comprise structurally controlled replacements and hydrothermal breccias, either in volcanic rocks as at Lepanto in the Mankayan district (Garcia, 1991; Hedenquist et al., 1998), Nena in the Frieda River district, Papua New Guinea (Espi, 1999), and Chelopech, Bulgaria (Chambefort and Moritz, 2006) or, where lithocaps impinge on carbonate rocks, as deposits like Smelter in the Marcapunta sector at Colquijirca (Vidal and Ligarda, 2004; Bendezú and Fontboté, 2009). In contrast, much larger tonnage, disseminated Au ± Ag orebodies are more typical of the shallower (<500 m) parts of lithocaps (Sillitoe, 1999b), as exemplified by Yanacocha (Harvey et al., 1999) and Pascua-Lama (Chouinard et al., 2005), although much deeper development of disseminated Cu-Au deposits is also relatively common (e.g., Tampakan; Rohrlach et al., 1999).

Intermediate-sulfidation epithermal precious metal deposits, containing Zn-Pb-Ag ± Cu ± Au as well as Mn-bearing carbonates, rhodonite, and quartz, occur alongside lithocaps but typically spatially separate from the high-sulfidation orebodies, as observed in the case of the Victoria and Teresa vein systems at Lepanto (Claveria, 2001; Hedenquist et al., 2001) and in the Collahuasi district (Masterman et al., 2005; Fig. 6). Locally, however, the intermediate-sulfidation and both Cordilleran lode and shallow, disseminated high-sulfidation mineralization types display transitional mineralogic relationships, as exemplified by the so-called Main Stage veins at Butte (Meyer et al., 1968) and the disseminated A and Link Au zones at Wafi-Golpu (Leach, 1999; Ryan and Vigar, 1999). The intermediate-sulfidation epithermal veins

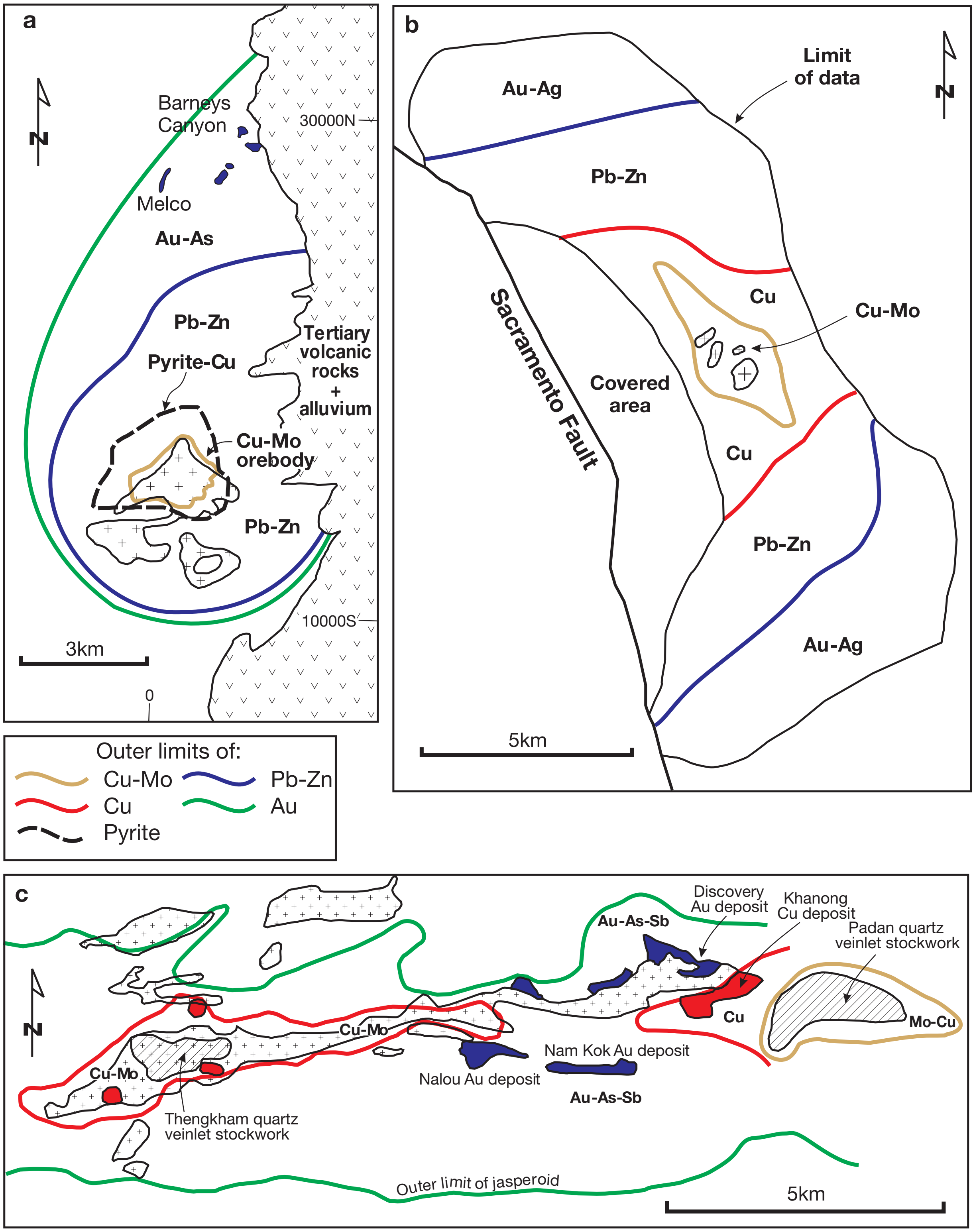

FIG. 9. Examples of well-developed metal zoning centered on porphyry Cu deposits. a. Bingham, Utah, where the porphyry Cu-Au-Mo deposit is followed successively outward by Cu-Au skarn, carbonate-replacement Zn-Pb-Ag-Au, and distal sediment-hosted Au deposits, the latter formerly exploited at Barneys Canyon and Melco (after Babcock et al., 1995). b. Mineral Park, Arizona, where the northwest-striking vein system centered on the porphyry Cu-Mo deposit is zoned outward from Cu through Pb-Zn to Au-Ag (after Lang and Eastoe, 1988). c. Sepon, Laos, where two subeconomic porphyry Mo-Cu centers marked by quartz veinlet stockworks are zoned outward through carbonate-replacement Cu to sediment-hosted Au deposits without any intervening Zn-Pb-Ag zone (summarized from R.H. Sillitoe, unpub. report, 1999). Note the large radii (up to 8 km) of some systems. Scales are different.

are the shallow-level (<1 km paleodepth) counterparts of the Zn-Pb-Cu-Ag ± Au veins located alongside porphyry Cu deposits, but no direct connection between the two types is evident (Fig. 6). The massive, high-sulfidation pyrite-enargite replacements in carbonate rocks are also locally transitional outward through high- to intermediate-sulfidation Zn-Pb-Ag ore, a continuum observed in the Tintic and Colquijirca districts (Lindgren and Loughlin, 1919; Bendezú et al., 2003; Bendezú and Fontboté, 2009).

Alteration-mineralization zoning in porphyry Cu deposits

Porphyry Cu deposits display a consistent, broad-scale alteration-mineralization zoning pattern that comprises, centrally from the bottom upward, several of sodic-calcic, potassic, chlorite-sericite, sericitic, and advanced argillic (cf. Meyer and Hemley, 1967; Table 2; Figs. 10, 11). Chloritic and propylitic alteration develop distally at shallow and deeper levels, respectively (Fig. 10). Equating chlorite-sericite alteration—the abbreviated name used by Hedenquist et al. (1998) for the sericite-clay-chlorite (SCC) of Sillitoe and Gappe (1984)—with Meyer and Hemley's (1967) low-temperature intermediate argillic type (e.g., Hedenquist et al., 1998; Sillitoe, 2000; Seedorff et al., 2005; Bouzari and Clark, 2006) causes confusion and should probably be discontinued. Phyllic (Lowell and Guilbert, 1970) and sericitic are synonyms.

The alteration-mineralization zoning sequence typically affects several cubic kilometers of rock (e.g., Lowell and Guilbert, 1970; Beane and Titley, 1981), although sericitic and,

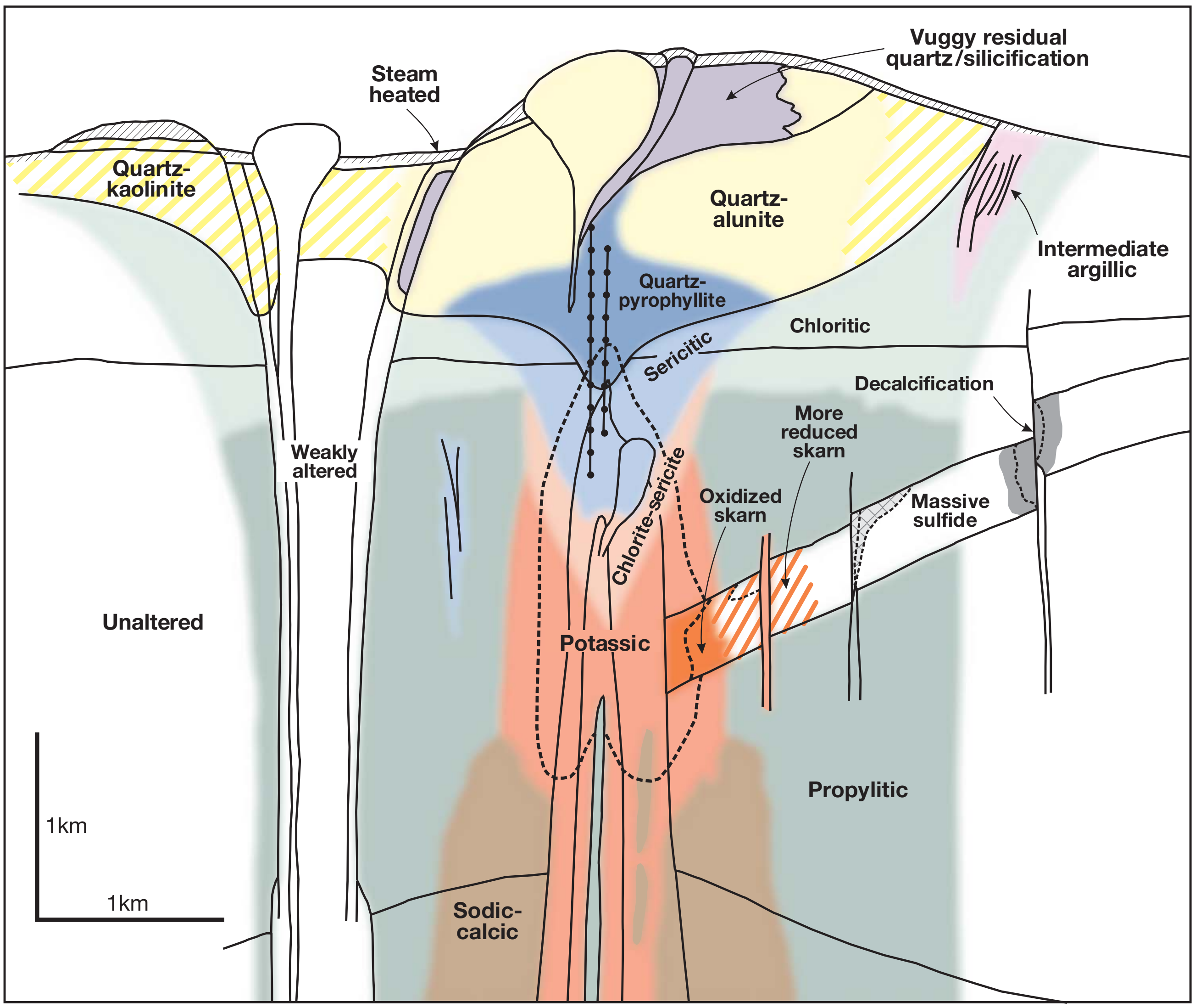

FIG. 10. Generalized alteration-mineralization zoning pattern for telescoped porphyry Cu deposits, based on the geologic and deposit-type template presented as Figure 6. Note that shallow alteration-mineralization types consistently overprint deeper ones. Volumes of the different alteration types vary markedly from deposit to deposit. Sericitic alteration may project vertically downward as an annulus separating the potassic and propylitic zones as well as cutting the potassic zone centrally as shown. Sericitic alteration tends to be more abundant in porphyry Cu-Mo deposits, whereas chlorite-sericite alteration develops preferentially in porphyry Cu-Au deposits. Alteration-mineralization in the lithocap is commonly far more complex than shown, particularly where structural control is paramount. See text for further details and Table 2 for alteration-mineralization details. Modified from Sillitoe (1999b, 2000).

TABLE 2. Characteristics of Principal Alteration-Mineralization Types in Porphyry Cu Systems[1]

Alteration type[2] (alternative name)	Position in system (abundance)	Key minerals	Possible ancillary minerals	Principal sulfide assemblages (minor)	Contemporaneous veinlets[3] (designation)	Veinlet selvages	Economic potential
Sodic-calcic	Deep, including below porphyry Cu deposits (uncommon)	Albite/oligoclase, actinolite, magnetite	Diopside, epidote, garnet	Typically absent	Magnetite ± actinolite (M-type)	Albite/oligoclase	Normally barren, but locally ore bearing
Potassic (K-silicate)	Core zones of porphyry Cu deposits (ubiquitous)	Biotite, K-feldspar	Actinolite, epidote, sericite, andalusite, albite, carbonate, tourmaline, magnetite	Pyrite-chalcopyrite, chalcopyrite ± bornite, bornite ± digenite ± chalcocite	Biotite (EB-type), K-feldspar, quartz-biotite-sericite-K-feldspar-andalusite-sulfides (EDM/T4-type), quartz-sulfides ± magnetite (A-type), quartz-molybdenite ± pyrite ± chalcopyrite (central suture; B-type)	EDM-type with sericite ± biotite ± K-feldspar ± andalusite + disseminated chalcopyrite ± bornite; others none, except locally K-feldspar around A- and B-types	Main ore contributor
Propylitic	Marginal parts of systems, below lithocaps (ubiquitous)	Chlorite, epidote, albite, carbonate	Actinolite, hematite, magnetite	Pyrite (± sphalerite, galena)	Pyrite, epidote		Barren, except for subepithermal veins
Chlorite-sericite (sericite-clay-chlorite [SCC])	Upper parts of porphyry Cu core zones (common, particularly in Au-rich deposits)	Chlorite, sericite/illite, hematite (martite, specularite)	Carbonate, epidote, smectite	Pyrite-chalcopyrite	Chlorite ± sericite ± sulfides	Chlorite, sericite/illite	Common ore contributor
Sericitic (phyllic)	Upper parts of porphyry Cu deposits (ubiquitous, except with alkaline intrusions)	Quartz, sericite	Pyrophyllite, carbonate, tourmaline, specularite	Pyrite ± chalcopyrite (pyrite-enargite ± tennantite, pyrite-bornite ± chalcocite, pyrite-sphalerite)	Quartz-pyrite ± other sulfides (D-type)	Quartz-sericite	Commonly barren, but may constitute ore
Advanced argillic (secondary quartzite in Russian terminology)	Above porphyry Cu deposits, constitutes lithocaps (common)	Quartz (partly residual, vuggy), alunite,[4] pyrophyllite, dickite, kaolinite	Diaspore, andalusite, zunyite, corundum, dumortierite, topaz, specularite	Pyrite-enargite, pyrite-chalcocite, pyrite-covellite	Pyrite-enargite ± Cu sulfides (includes veins)	Quartz-alunite, quartz-pyrophyllite/dickite, quartz-kaolinite	Locally constitutes ore in lithocaps and their roots

[1] Excluding those developed in carbonate-rich rocks
[2] Arranged from probable oldest (top) to youngest (bottom), except for propylitic that is lateral equivalent of potassic; advanced argillic also forms above potassic early in systems (Fig. 10)
[3] Many veinlets in potassic, chlorite-sericite, and sericitic alteration contain anhydrite, which also occurs as late, largely monomineralic veinlets
[4] Alunite commonly intergrown with aluminum-phosphate-sulfate (APS) minerals (see Stoffregen and Alpers, 1987)

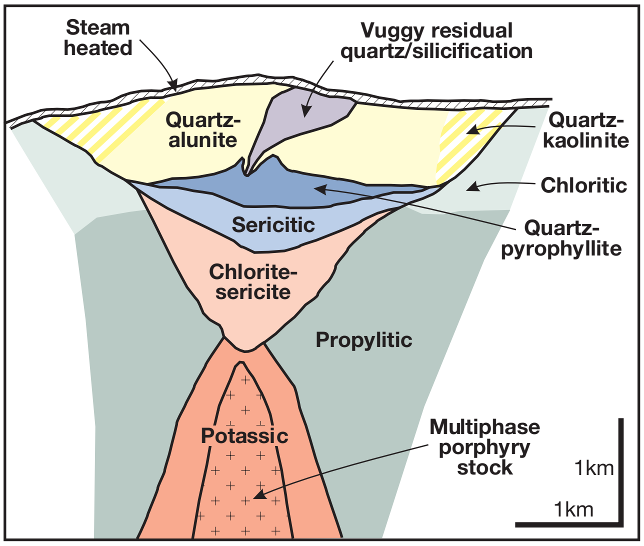

FIG. 11. Generalized alteration-mineralization zoning pattern for a nontelescoped porphyry Cu system, emphasizing the appreciable, commonly barren gap that exists between the lithocap and underlying porphyry stock. Legend as in Figure 10.

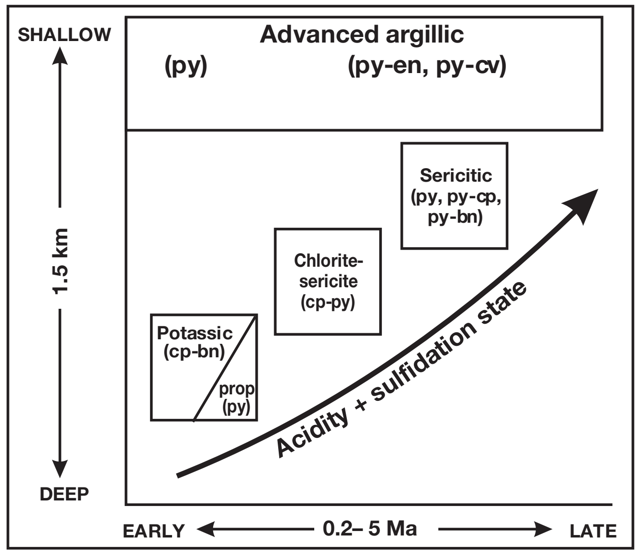

FIG. 12. Schematic representation of generalized alteration-mineralization sequence in porphyry Cu systems in relation to paleodepth and system life span. The sequence, from potassic with peripheral propylitic (prop) through chlorite-sericite and sericitic to advanced argillic, is the result of increasing acidity consequent upon the declining temperature of the hydrothermal fluids. A broadly parallel increase in sulfidation state of the fluids results in changes in the sulfide assemblage from chalcopyrite (cp)-bornite (bn), through chalcopyrite-pyrite (py) and pyrite-bornite, to pyrite-enargite (en) or pyrite-covellite (cv), as charted for several deposits by Einaudi et al. (2003). Note the absence of Cu-bearing sulfides from the early, high-temperature advanced argillic zone. Modified from Sillitoe (2000).

particularly, advanced argillic alteration are much less well developed in porphyry Cu deposits associated with alkaline than with calc-alkaline intrusions (Lang et al., 1995; Sillitoe, 2002; Holliday and Cooke, 2007), reflecting control of the K^+/H^+ ratio by magma chemistry (e.g., Burnham, 1979). Specific opaque mineral assemblages are intrinsic parts of each alteration type (Table 2; Fig. 12) because of the direct linkage between sulfidation state, the chief control on sulfide assemblages, and solution pH, a principal control of alteration type (Barton and Skinner, 1967; Meyer and Hemley, 1967; Einaudi et al., 2003; Fig. 12). Sulfidation state, a function of S fugacity and temperature, changes from low through intermediate to high as temperature declines (Barton and Skinner, 1967; Einaudi et al., 2003). In general, the alteration-mineralization types become progressively younger upward (Fig. 12), with the result that the shallower alteration-mineralization zones invariably overprint and at least partly reconstitute deeper ones.

Sodic-calcic alteration, commonly magnetite bearing (Table 2), is normally rather poorly preserved at depth in some porphyry Cu deposits, commonly in the immediate wall rocks to the porphyry intrusions (e.g., Panguna, Papua New Guinea and El Teniente; Ford, 1978; Cannell et al., 2005), a position that may give rise to confusion with propylitic alteration (Fig. 10). Nevertheless, it also characterizes the centrally located zones of some porphyry Cu stocks (e.g., Koloula, Solomon Islands and Island Copper, British Columbia; Chivas, 1978; Perelló et al., 1995; Arancibia and Clark, 1996). Sodic-calcic alteration is typically sulfide and metal poor (except for Fe as magnetite) but can host mineralization in Au-rich porphyry Cu deposits (e.g., Nugget Hill, Philippines), in some of which hybrid potassic-calcic (biotite-actinolite-magnetite) assemblages are also commonplace (e.g., Santo Tomas II, Ridgeway, and Cotabambas, southern Peru; Sillitoe and Gappe, 1984; Wilson et al., 2003; Perelló et al., 2004a).

Large parts of many porphyry Cu deposits (e.g., Lowell and Guilbert, 1970; Titley, 1982), especially deeply formed (e.g., Butte; Rusk et al., 2004, 2008a) or relatively deeply eroded examples like El Abra (Ambrus, 1977; Dean et al., 1996) and Gaby (Gabriela Mistral), northern Chile (Camus, 2001, 2003), are made up predominantly of potassic alteration, which grades marginally into generally weakly developed propylitic zones (Fig. 10). Biotite is the predominant alteration mineral in relatively mafic porphyry intrusions and host rocks, whereas K-feldspar increases in abundance in more felsic, granodioritic to quartz monzonitic settings. Sodic plagioclase may be an accompanying alteration mineral in both settings. Locally, texture-destructive quartz-K ± Na-feldspar flooding overprints and destroys the more typical potassic assemblages (e.g., Chuquicamata; Ossandón et al., 2001). The chalcopyrite ± bornite ore in many porphyry Cu deposits is largely confined to potassic zones (Table 2; Fig. 12), with one or more bornite-rich centers characterizing the deeper, central parts of many deposits. In some bornite-rich centers, the sulfidation state is low enough to stabilize digenite ± chalcocite (Einaudi et al., 2003; Table 2). Chalcopyrite-bornite cores are transitional outward to chalcopyrite-pyrite annuli, which, with increasing sulfide contents, grade into pyrite halos, typically parts of the surrounding propylitic zones (Table 2). Pyrrhotite may accompany the pyrite where reduced host rocks are present (e.g., Kósaka and Wakita, 1978; Perelló et al., 2003b). The potassic alteration affects the early and intermineral porphyry generations (Fig. 7) and many intermineral magmatic-hydrothermal breccias as well as variable volumes of wall rocks. The potassic-altered wall rocks range from restricted veneers near the stock or dike contacts to kilometer-scale zones, such as those in the mafic host lithologic units mentioned previously at El Teniente, Resolution, and Oyu Tolgoi. The potassic alteration generally becomes less intense from the older to younger porphyry phases, although the late-mineral intrusions postdate it and display a propylitic assemblage (Fig. 7), albeit of later timing than the propylitic halos developed peripheral to potassic zones.

Chlorite-sericite alteration (Table 2), giving rise to distinctive, pale-green rocks, is widespread in the shallower parts of some porphyry Cu deposits, particularly Au-rich examples, where it overprints preexisting potassic assemblages (Figs. 10, 11). The alteration is typified by partial to complete transformation of mafic minerals to chlorite, plagioclase to sericite (fine-grained muscovite) and/or illite, and magmatic and any hydrothermal magnetite to hematite (martite and/or specularite), along with deposition of pyrite and chalcopyrite. Although Cu and/or Au tenors of the former potassic zones may undergo depletion during the chlorite-sericite overprints (e.g., Esperanza, northern Chile; Perelló et al., 2004b), metal introduction is also widely recognized (e.g., Leach, 1999; Padilla Garza et al., 2001; Harris et al., 2005; Masterman et al., 2005) and, at a few deposits, is considered to account for much of the contained Cu (e.g., Cerro Colorado, northern Chile; Bouzari and Clark, 2006).

Sericitic alteration (Table 2) in porphyry Cu deposits normally overprints and wholly or partially destroys the potassic and chlorite-sericite assemblages (Figs. 10–12), although sericitic veinlet halos are zoned outward to chlorite-sericite alteration in places (e.g., Dilles and Einaudi, 1992). The degree of overprinting is perhaps best appreciated in some magmatic-hydrothermal breccia bodies in which isolated magnetite aggregates occur as stranded remnants in sericitic or chlorite-sericite zones up to 1 km above the magnetite-cemented, potassic parts (e.g., Chimborazo, northern Chile; Fig. 8). Sericitic alteration may be subdivided into two different types: an uncommon, early variety that is greenish to greenish-gray in color and a later, far more common and widespread, white variety. In the few deposits where it is recognized, the early, greenish sericitic alteration is centrally located and hosts a low sulfidation-state chalcopyrite-bornite assemblage, which is commonly ore grade (e.g., Chuquicamata; Ossandón et al., 2001). The late, white sericitic alteration has varied distribution patterns in porphyry Cu deposits. It may constitute annular zones separating the potassic cores from propylitic halos, as emphasized in early porphyry Cu models (Jerome, 1966; Lowell and Guilbert, 1970; Rose, 1970), but is perhaps more common as structurally controlled or apparently irregular replacements within the upper parts of chlorite-sericite and/or potassic zones (Fig. 10). The sericitic alteration is commonly pyrite dominated, implying effective removal of the Cu (± Au) present in the former chlorite-sericite and/or potassic assemblages. However, sericitic alteration may also constitute ore where appreciable Cu remains with the pyrite, either in the form of chalcopyrite or as high sulfidation-state assemblages (typically, pyrite-bornite, pyrite-chalcocite, pyrite-covellite, pyrite-tennantite, and pyrite-enargite; Table 2; cf. Einaudi et al., 2003).

The main development of these bornite-, chalcocite-, and covellite-bearing, high-sulfidation assemblages is largely confined to white sericitic alteration that overprints now-barren quartz veinlet stockworks (see below). These high-sulfidation assemblages commonly have higher Cu contents than the former potassic alteration, resulting in hypogene enrichment (Brimhall, 1979), although any Au may be depleted (e.g., Wafi-Golpu; Sillitoe, 1999b). The Cu-bearing sulfides typically occur as fine-grained coatings on disseminated pyrite grains, which leads to recognition difficulties in deposits that also underwent supergene Cu sulfide enrichment (e.g., Chuquicamata, Ossandón et al., 2001); indeed, the hypogene contribution is commonly not distinguished from the supergene enrichment products (e.g., Taca Taca Bajo, Argentina; Rojas et al., 1999).

The root zones of advanced argillic lithocaps, commonly at least partly structurally controlled, may overprint the upper parts of porphyry Cu deposits, where the sericitic alteration is commonly transitional upward to quartz-pyrophyllite (Fig. 10), an assemblage widespread in the deep, higher temperature parts of many lithocaps (e.g., El Salvador; Gustafson and Hunt, 1975; Watanabe and Hedenquist, 2001). Elsewhere, however, lower temperature quartz-kaolinite is the dominant overprint assemblage (e.g., Caspiche, northern Chile). The advanced argillic alteration preferentially affects lithologic units with low (e.g., quartz sandstone, felsic igneous rocks) rather than high (mafic igneous rocks) acid-buffering capacities. At several localities, the advanced argillic alteration at the bottoms of lithocaps displays a characteristic patchy texture, commonly defined by amoeboid pyrophyllite patches embedded in silicified rock (e.g., Escondida and Yanacocha; Padilla Garza et al., 2001; Gustafson et al., 2004). However, the patches may also comprise alunite or kaolinite, suggesting that the texture may result by either preferential nucleation of any common advanced argillic mineral or even advanced argillic overprinting of a nucleation texture developed during earlier potassic or chlorite-sericite alteration of fragmental rocks (e.g., Hugo Dummett; Khashgerel et al., 2008, and Caspiche).

The vertical distribution of the alteration-mineralization types in porphyry Cu deposits depends on the degree of overprinting or telescoping, the causes of which are addressed further below. In highly telescoped systems, the advanced argillic lithocaps impinge on the upper parts of porphyry stocks (Fig. 10) and their roots may penetrate downward for >1 km. In such situations, the advanced argillic alteration may be 1 to >2 m.y. younger than the potassic zone that it overprints (e.g., Chuquicamata and Escondida; Ossandón et al., 2001; Padilla-Garza et al., 2004), reflecting the time needed for the telescoping to take place. Where telescoping is limited, however, the lithocaps and potassic-altered porphyry stocks may be separated by 0.5 to 1 km (Sillitoe, 1999b), a gap typically occupied by pyritic chlorite-sericite alteration (Fig. 11).

Where carbonate (limestone and dolomite) instead of igneous or siliciclastic rocks host porphyry Cu deposits, calcic or magnesian exoskarn is generated in proximity to the porphyry intrusions, whereas marble is produced beyond the skarn front (Fig. 10). In the case of limestone protoliths, anhydrous, prograde andraditic garnet-diopsidic pyroxene skarn forms contemporaneously with the potassic alteration of noncarbonate lithologic units, whereas hydrous, retrograde skarn, commonly containing important amounts of magnetite, actinolite, epidote, chlorite, smectite, quartz, carbonate, and iron sulfides, is the equivalent of the chlorite-sericite and sericitic assemblages (Einaudi et al., 1981; Meinert et al., 2003). Distal Au skarns are typically more reduced (pyroxene rich) than their proximal counterparts (Fortitude, Copper Canyon district; Myers and Meinert, 1991; Fig. 10). A quartz-pyrite assemblage replaces any carbonate rocks incorporated in advanced argillic lithocaps (e.g., Bisbee, Arizona; Einaudi, 1982). Endoskarn tends to be volumetrically minor (Beane and Titley, 1981; Meinert et al., 2005). The massive sulfide carbonate-replacement deposits are normally enveloped by marble. Any sediment-hosted Au mineralization on the fringes of carbonate rock-hosted porphyry Cu systems forms where rock permeability is enhanced by decalcification (Fig. 10), including sanding of dolomite, but also locally occluded by Au-related jasperoid formation (e.g., Bingham and Sepon districts; Babcock et al., 1995; Smith et al., 2005; Fig. 9a, d).

Porphyry Cu veinlet relationships

The veinlet sequence in porphyry Cu deposits, first elaborated by Gustafson and Hunt (1975) at El Salvador and widely studied since (e.g., Hunt et al., 1983; Dilles and Einaudi, 1992; Gustafson and Quiroga, 1995; Redmond et al., 2001; Pollard and Taylor, 2002; Cannell et al., 2005; Masterman et al., 2005), is highly distinctive. In a general way, the veinlets may be subdivided into three groups (Table 2; Fig. 13): (1) early, quartz- and sulfide-free veinlets containing one or more of actinolite, magnetite (M type), (early) biotite (EB type), and K-feldspar, and typically lacking alteration selvages; (2) sulfide-bearing, granular quartz-dominated veinlets with either narrow or no readily recognizable alteration selvages (A and B types); and (3) late, crystalline quartz-sulfide veins and veinlets with prominent, feldspar-destructive alteration selvages (including D type). Group 1 and 2 veinlets are mainly emplaced during potassic alteration, whereas group 3 accompanies the chlorite-sericite, sericitic, and deep advanced argillic overprints. Narrow, mineralogically complex quartz-sericite-K-feldspar-biotite veinlets with centimeter-scale halos defined by the same minerals (± andalusite ± corundum) along with abundant, finely disseminated chalcopyrite ± bornite characterize the changeover from group 1 to 2 veinlets in a few deposits, although they may have been confused elsewhere with D-type veinlets because of their eye-catching halos; they are termed early dark micaceous (EDM) halo veinlets at Butte (Meyer, 1965; Brimhall, 1977; Rusk et al., 2008a) and Bingham (Redmond et al., 2004), and type 4 (T4) veinlets at Los Pelambres (Atkinson et al., 1996; Perelló et al., 2007). Group 3 also includes uncommon, but economically important massive chalcopyrite ± bornite ± chalcocite veinlets at the high-grade Grasberg (Pollard and Taylor, 2002; I. Kavalieris, pers. commun., 1999), Hugo Dummett (Khashgerel et al., 2008), and Resolution deposits as well as elsewhere.

Many porphyry Cu deposits display single veinlet sequences that comply with the generalizations summarized above and

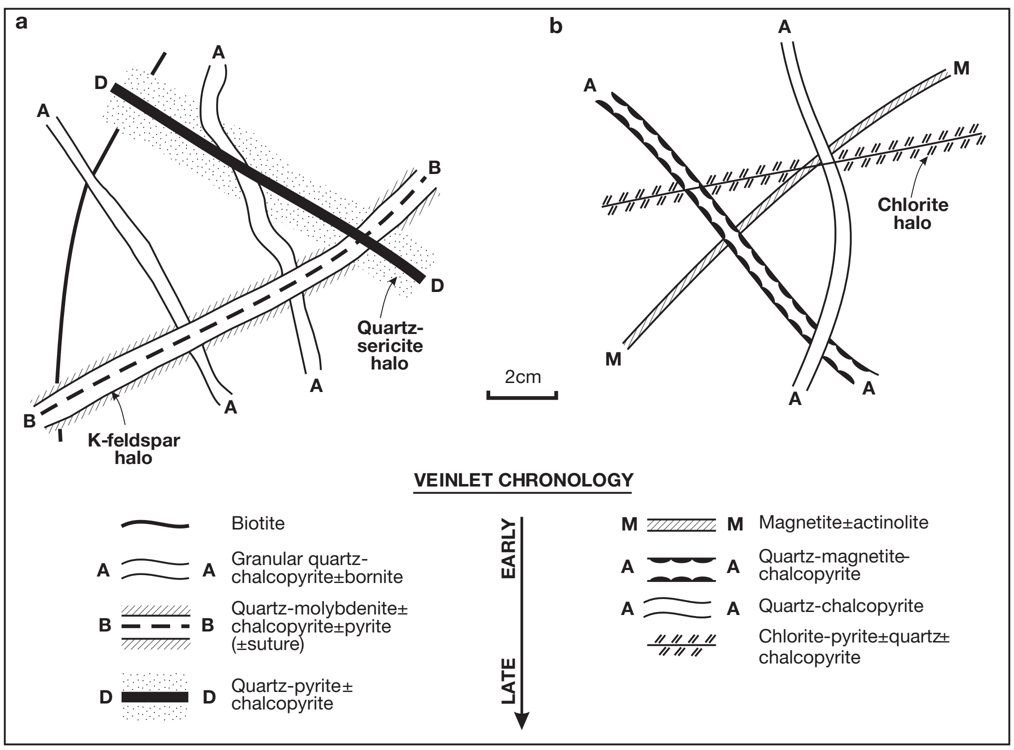

FIG. 13. Schematic chronology of typical veinlet sequences in a. porphyry Cu-Mo deposits and b. porphyry Cu-Au deposits associated with calc-alkaline intrusions. Porphyry Cu-Au deposits hosted by alkaline intrusions are typically veinlet poor (Barr et al., 1976; Lang et al., 1995; Sillitoe, 2000, 2002). Background alteration between veinlets is mainly potassic, which is likely to contain more K-feldspar in the Mo-rich than the Au-rich porphyry Cu stockworks. Note the common absence of B- and D-type veinlets from Au-rich porphyry Cu stockworks and M-, magnetite-bearing A-, and chlorite-rich veinlets from Mo-rich porphyry Cu stockworks. Veinlet nomenclature follows Gustafson and Hunt (1975; A, B, and D types) and Arancibia and Clark (1996; M type).

in Figure 13 and Table 2, but repetitions of group 1 and 2 veinlets, for example, early biotite, EDM halo, and A types cut by lesser numbers of later EDM halo and A types (e.g., Bingham; Redmond et al., 2001), occur where time gaps between porphyry phases are sufficiently large; however, group 2 and 3 veinlets are only uncommonly repeated. Additional complications are widely introduced by repetitive veinlet reopening during subsequent veining events. Much of the metal in many porphyry Cu deposits is contained in the quartz-dominated, group 2 veinlets and as disseminated grains in the intervening potassic-altered rocks, although some of the late, group 3 quartz-sulfide veins and their wall rocks may also be important contributors. Irrespective of whether the Cu-bearing sulfide minerals are coprecipitated with veinlet quartz or, as generally seems to be the case, introduced paragenetically later (e.g., Redmond et al., 2001, 2004), a particularly strong correlation exists between quartz veinlet intensity and metal content in many porphyry Cu deposits, particularly in Au-rich examples (Sillitoe, 2000). However, the porphyry Cu-Au deposits associated with alkaline rocks, particularly those in British Columbia, are largely devoid of veinlets (Barr et al., 1976). Once formed, the quartz-bearing veinlets are permanent features that are not erased during subsequent alteration overprinting, although their metal contents may be wholly or partially removed (see above). Therefore recognition of A- and B-type veinlets in sericitic or advanced argillic zones testifies unambiguously to the former presence of potassic alteration.

The A-type veinlets range from stockworks to subparallel, sheeted arrays, the latter particularly common in Au-rich porphyry deposits (Sillitoe, 2000). Few, if any, stockworks are truly multidirectional and one or more preferred veinlet orientations are the norm. These may reflect district-scale tectonic patterns (e.g., Heidrick and Titley, 1982; Lindsay et al., 1995) or, where concentric and radial arrays predominate, control by magma ascent and/or withdrawal in the subjacent parental chambers (e.g., El Teniente; Cannell et al., 2005). The quartz veinlet stockworks are most intense in and around the early porphyry intrusions, where they may constitute as much as 90 to 100 percent of the rock (e.g., Ok Tedi and Hugo Dummett; Rush and Seegers, 1990; Khashgeral et al., 2006), and die out gradually both laterally into the wall rocks (e.g., Sierrita-Esperanza, Arizona; Titley et al., 1986) and downward (e.g., El Salvador; Gustafson and Quiroga, 1995); however, they tend to have more clearcut upper limits, just a few tens of meters above the apices of the porphyry intrusions, in the few deposits where relevant data are available (e.g., Guinaoang, Wafi-Golpu, and Hugo Dummett; Sillitoe and Angeles, 1985; Sillitoe, 1999b; Khashgeral et al., 2006). The quartz veinlets commonly cut proximal prograde exoskarn (Einaudi, 1982) but do not extend into the more distal carbonate rock-hosted ore types. Locally, early A-type veinlets displaying aplitic centers or along-strike transitions to aplite and/or aplite porphyry (vein dikes) are observed (e.g., Gustafson and Hunt, 1975; Heithersay et al., 1990; Lickfold et al., 2003; Rusk et al., 2008a). The earliest A-type veinlets may be sinuous and have nonmatching margins, features ascribed to formation under high-temperature, overall ductile conditions, whereas later veinlets are more planar.

Much of the Mo in many porphyry Cu-Mo deposits occurs in the B-type veinlets, in marked contrast to the Cu dominance of the A-type generations, but D-type veinlets may also contain appreciable amounts of molybdenite in some deposits. The B-type veinlets are typically absent from Au-rich, Mo-poor porphyry Cu deposits (Fig. 13b). The D-type veinlets, far more abundant in porphyry Cu-Mo than Cu-Au deposits (Fig. 13a), may also occur as structurally controlled swarms (e.g., El Abra; Dean et al., 1996), a characteristic particularly evident in the case of the late-stage, meter-scale, enargite-bearing, massive sulfide veins spanning the upper parts of porphyry Cu deposits and lower parts of overlying lithocaps (Fig. 6; see above).

Magnetite ± actinolite (M-type) and quartz-magnetite (A-type) veinlets are far less common in Mo- than Au-rich porphyry Cu deposits (Fig. 13), the latter typified by particularly elevated hydrothermal magnetite contents, commonly attaining 5 to 10 vol percent (Sillitoe, 1979, 2000; MacDonald and Arnold, 1994; Proffett, 2003). The dominant veinlets in most Au-only porphyry deposits, as documented in the Maricunga belt, are distinctly banded and comprise layers of both translucent and dark-gray quartz (Vila and Sillitoe, 1991), the color of the latter commonly caused by abundant vapor-rich fluid inclusions (Muntean and Einaudi, 2000). These banded veinlets are ascribed to the shallowness of porphyry Au formation (<1 km; Vila and Sillitoe, 1991; Muntean and Einaudi, 2000).

Anhydrite and tourmaline are prominent veinlet, breccia-filling, and alteration minerals in many porphyry Cu deposits (Table 2), including associated skarns. The anhydrite, attaining 5 to 15 percent of rock volumes, occurs in small amounts in most group 2 and 3 veinlet types as well as in the form of disseminated grains in the intervening altered rocks but commonly also constitutes end-stage, nearly monomineralic veinlets. Absence of anhydrite to depths of several hundred meters beneath the current surface in many porphyry Cu systems is normally due to supergene dissolution (see Sillitoe, 2005). Tourmaline may occur in minor amounts in several veinlet types, even those formed early in deposit histories (e.g., T4 veinlets al Los Pelambres; Perelló et al., 2007), but it is most abundant with quartz and/or pyrite in D-type veinlet generations and any associated magmatic-hydrothermal breccias affected by sericitic alteration (Fig. 8).

Advanced argillic lithocaps

The upper parts of porphyry Cu systems, mainly at shallower levels than their porphyry intrusions, are characterized by lithocaps: lithologically controlled zones of pervasive advanced argillic alteration with structurally controlled components, including their subvertical root zones (Figs. 4, 6, 10; Table 2; Sillitoe, 1995a). Original lithocaps have areal extents of several to >10 and, locally, up to 100 km^2 and thicknesses of >1 km, and hence are much more extensive than the underlying porphyry Cu deposits. Indeed, two or more porphyry Cu deposits may underlie some large, coalesced lithocaps (Fig. 4), which, as noted above, may have formed progressively over periods of up to several million years (e.g., Yanacocha; Gustafson et al., 2004; Longo and Teal, 2005). Most observed lithocaps are only erosional remnants, which may either wholly or partially overlie and conceal porphyry Cu deposits (e.g., Wafi-Golpu; Sillitoe, 1999b) or occur alongside them and, hence, above propylitic rock (e.g., Nevados del Famatina, Argentina, Batu Hijau, and Rosia Poieni, Romania; Lozada-

Calderón and McPhail, 1996; Clode et al., 1999; Milu et al., 2004; Figs. 6, 10). Many lithocaps are vertically zoned, from the previously described quartz-pyrophyllite at depth to predominant quartz-alunite and residual quartz—the residue of extreme base leaching (Stoffregen, 1987) with a vuggy appearance that reflects the original rock texture—at shallower levels where the causative fluid was cooler and, hence, more acidic (Giggenbach, 1997; Fig. 10). The roots of lithocaps may also contain the relatively high temperature species, andalusite and corundum (>~370°C; Hemley et al., 1980), as accompaniments to pyrophyllite and/or muscovite (e.g., Cabang Kiri, Indonesia, El Salvador, and Cerro Colorado; Lowder and Dow, 1978; Watanabe and Hedenquist, 2001; Bouzari and Clark, 2006). Where the fluids that cause advanced argillic alteration are F rich, topaz, zunyite, and fluorite are lithocap minerals (e.g., Hugo Dummett; Perelló et al., 2001; Khashgerel et al., 2006, 2008, and Resolution). The principal borosilicate mineral in lithocaps is dumortierite rather than tourmaline. The more structurally and lithologically confined components of lithocaps, termed ledges rather than veins because they are mainly the products of rock replacement rather than incremental open-space filling, display well-developed alteration zoning (e.g., Steven and Ratté, 1960; Stoffregen, 1987), with cores of vuggy, residual quartz, and associated silicification rimmed outward (and downward) by consecutive bands of quartz-alunite, quartz-pyrophyllite/dickite/kaolinite (pyrophyllite and dickite at hotter, deeper levels), and chlorite-illite/smectite.

Although all these alteration zones are pyritic, the Au-, Ag-, and Cu-bearing, high sulfidation-state assemblages (commonly pyrite-enargite and pyrite-covellite; Table 2; Fig. 12) tend to be confined to the vuggy, residual quartz and silicified rock, the latter normally better mineralized where phreatic breccias are present (see above). Apart from the massive, commonly enargite-bearing sulfide veins and replacement bodies in the deeper parts of some lithocaps (see above), veins and veinlets are generally only poorly developed, with much of the pyrite and any associated sulfides being in disseminated form. Open-space filling is also uncommon, except in phreatic breccias and unusual, isolated veins (e.g., La Mejicana alunite-pyrite-famatinite vein at Nevados del Famatina; Lozada-Calderón and McPhail, 1996). Barite and native S are common late-stage components of many ledges.

These advanced argillic alteration zones extend upward to the sites of paleowater tables, which may be defined, if suitable aquifers (e.g., fragmental volcanic rocks) were present, by subhorizontal, tabular bodies of massive opaline or chalcedonic silicification, up to 10 m or so thick; the low crystallinity is caused by the low temperature (~100°C) of silica deposition. The overlying vadose zones are marked by easily recognizable, steam-heated alteration rich in fine-grained, powdery cristobalite, alunite, and kaolinite (Sillitoe, 1993, 1999b; Fig. 10).

Metal zoning

Metal zoning in porphyry Cu systems is well documented, particularly at the deeper, porphyry Cu levels (e.g., Jerome, 1966; Titley, 1993). There, Cu ± Mo ± Au characterize the potassic, chlorite-sericite, and sericitic cores of systems. However, in Au-rich porphyry Cu deposits, the Au, as small (<20 μm) grains of high (>900) fineness native metal and in solid solution in bornite and, to a lesser degree, chalcopyrite (e.g., Arif and Baker, 2002), and Cu are introduced together as components of centrally located potassic zones; hence, the two metals normally correlate closely (Sillitoe, 2000; Ulrich and Heinrich, 2001; Perelló et al., 2004b). Gold grades may be up to ~50 percent higher in bornite-rich than chalcopyrite-dominated potassic assemblages, which has been explained by the experimental observation that bornite solid solution is capable of holding up to one order of magnitude more Au than intermediate solid solution (ISS), the high-temperature precursors of bornite and chalcopyrite, respectively (Simon et al., 2000; Kesler et al., 2002). The Au grains in some deposits contain minor amounts of PGE minerals, particularly Pd tellurides (Tarkian and Stribrny, 1999). In contrast, Cu and Mo correlate less well, with spatial separation of the two metals commonly resulting from the different timing of their introduction (e.g., Los Pelambres; Atkinson et al., 1996). In many Au-rich porphyry Cu deposits, Mo tends to be concentrated as external annuli partly overlapping the Cu-Au cores (e.g., Saindak, Pakistan, Cabang Kiri, Batu Hijau, Bajo de la Alumbrera, and Esperanza; Sillitoe and Khan, 1977; Lowder and Dow, 1978; Ulrich and Heinrich, 2001; Garwin, 2002; Proffett, 2003; Perelló et al., 2004b). The Bingham, Island Copper, and Agua Rica, Argentina, porphyry Cu-Au-Mo deposits are exceptions to this generalization because of their deep, centrally located molybdenite zones (John, 1978; Perelló et al., 1995, 1998).

The Cu ± Mo ± Au cores typically have kilometer-scale halos defined by anomalous Zn, Pb, and Ag values that reflect lower temperature, hydrothermal conditions (Fig. 9a, b). In some systems, Mn (±Ag) is also markedly enriched in the outermost parts of the halos (e.g., Butte; Meyer et al., 1968). These Zn-Pb-Ag ± Mn halos commonly coincide spatially with propylitic alteration zones but are invariably best defined in the distal skarn environment (e.g., Meinert, 1987; Meinert et al., 2005), beyond which even more distal Au-As ± Sb zones may be developed (e.g., Bingham and Sepon districts; Babcock et al., 1995; Cunningham et al., 2004; Smith et al., 2005; Fig. 9a, c). Peripheral veins cutting propylitic halos may also be Au rich, and at Mineral Park an outward zoning from Pb-Zn to Au-Ag is evident (Eidel et al., 1968; Lang and Eastoe, 1988; Fig. 9b). Nevertheless, in some porphyry Cu deposits, these halo metals, particularly Zn, occur as late-stage veinlet arrays overprinting the Cu-dominated cores rather than peripherally (e.g., Chuquicamata; Ossandón et al., 2001).

In a general sense, the broad-scale zoning pattern developed in the deeper parts of porphyry Cu systems persists into the overlying lithocap environment where any Cu and Au (±Ag) commonly occur approximately above the underlying porphyry Cu deposits, albeit commonly areally more extensively, particularly where structural control is prevalent. The main geochemical difference between the Cu-Au zones in porphyry Cu deposits and those in the overlying lithocaps is the elevated As (±Sb) contents consequent upon the abundance of the Cu sulfosalts in the latter. Nevertheless, the lithocap mineralization also contains greater albeit trace amounts of Bi, W, Sn, and/or Te (e.g., Einaudi, 1982) as well as appreciable Mo. The Cu/Au ratios of lithocap-hosted, high-sulfidation mineralization tend to decrease upward, with the result that most major high-sulfidation Au (±Ag) deposits occur in the shallow parts of lithocaps, commonly—but not always—with their tops immediately below the former paleowater table positions (Sillitoe, 1999b). Nevertheless, supergene leaching commonly masks the original Cu distribution pattern. Any intermediate-sulfidation precious metal mineralization developed alongside the lithocaps contains much higher contents of Zn, Pb, Ag, and Mn than the high-sulfidation orebodies, in keeping with the situation described above from the porphyry Cu level. The shallow-level, steam-heated and paleowater-table zones are typically devoid of precious and base metals and As and Sb, unless telescoped onto the underlying mineralization as a result of water-table descent, but commonly have elevated Hg contents (e.g., Pascua-Lama; Chouinard et al., 2005).

Genetic Model

Magma and fluid production

Porphyry Cu systems typically span the upper 4 km or so of the crust (Singer et al., 2008; Figs. 6, 10), with their centrally located stocks being connected downward to parental magma chambers at depths of perhaps 5 to 15 km (Cloos, 2001; Richards, 2005; Fig. 4). The parental chambers, tending to be localized at sites of neutral buoyancy (Cloos, 2001; Richards, 2005), are the sources of both magmas and high-temperature, high-pressure metalliferous fluids throughout system development.

Field observations and theoretical calculations suggest that parental chambers with volumes on the order of 50 km^3 may be capable of liberating enough fluid to form porphyry Cu deposits, but chambers at least an order of magnitude larger are needed to produce giant systems, particularly where deposit clusters or alignments exist (Dilles, 1987; Cline and Bodnar, 1991; Shinohara and Hedenquist, 1997; Cloos, 2001; Cathles and Shannon, 2007). The metal-charged aqueous phase is released from the cooling and fractionating parental chambers during open-system magma convection as well as later stagnant magma crystallization (Shinohara and Hedenquist, 1997). Convection provides an efficient mechanism for delivery of copious amounts of the aqueous phase, in the form of bubble-rich magma, from throughout the parental chambers to the basal parts of porphyry stocks or dike swarms (Candela, 1991; Shinohara et al., 1995; Cloos, 2001; Richards, 2005). In most systems, any volcanism ceases before porphyry Cu system formation is initiated, although relatively minor eruptive activity, such as dome emplacement, may be either interspersed with or perhaps even accompany ascent of the magmatic aqueous phase (e.g., Bingham and Yanacocha; Deino and Keith, 1997; Longo and Teal, 2005).

The shallow-level porphyry stocks do not themselves generate the bulk of the magmatic fluid volume, but simply act as "exhaust valves," conduits for its upward transmission from the parental chambers, perhaps via cupolas on their roofs (Fig. 4). This scenario implies episodic but focused magma and fluid ascent for as long as ~5 m.y. in the case of long-lived porphyry Cu systems, whereas elsewhere the loci of intrusive and hydrothermal activity migrate, either systematically or randomly, to give rise to the porphyry Cu and epithermal Au deposit clusters and alignments discussed above.

The parental magmas need to be water rich (>~4 wt %) and oxidized in order to maximize the metal contents of the resultant aqueous phase (Burnham and Ohmoto, 1980; Candela and Holland, 1986; Dilles, 1987; Cline and Bodnar, 1991; Candela, 1992; Candela and Piccoli, 2005; Richards, 2005). High water contents result in magmas becoming saturated with the aqueous phase, into which the ore metals can partition efficiently; and high oxidation state suppresses magmatic sulfide, such as pyrrhotite, precipitation, a process that may cause sequestration of metals before they can partition into the aqueous phase. Nevertheless, resorption of any sulfide melt during ascent of oxidized magmatic fluids could make a major contribution to metal budgets (Keith et al., 1997; Halter et al., 2005). The magmas are also exceptionally S rich, as emphasized by recognition of anhydrite as a magmatic mineral in some porphyry stocks (Lickfold et al., 2003; Audétat et al., 2004; Stern et al., 2007; Chambefort et al., 2008). Addition of mafic melt to the parental chambers could be an effective means of augmenting S and metal budgets (Keith et al., 1997; Hattori and Keith, 2001; Maughan et al., 2002; Halter et al., 2005; Zajacz and Halter, 2009).

Early porphyry Cu system evolution

Porphyry Cu mineralization in the deeply formed (up to 9 km) potassic alteration zones at Butte and elsewhere took place directly from a single-phase, relatively low salinity (2–10 wt % NaCl equiv), aqueous liquid (Rusk et al., 2004, 2008a); such a phase may contain several thousand ppm to several percent of base metals and several ppm Au, based on thermodynamic (Heinrich, 2005) and analytical (Audétat et al., 2008) observations. However, at the shallower depths typical of most deposits (<~4 km), the mineralization is introduced by a two-phase fluid, comprising a small fraction of hypersaline liquid (brine) and a much larger volume of low-density vapor (Fournier, 1999), produced by either direct exsolution from the melt (Shinohara, 1994) or, more typically, as the single-phase liquid decompresses, cools, and intersects its solvus (e.g., Henley and McNabb, 1978; Burnham, 1979; Cline and Bodnar, 1991; Webster, 1992; Bodnar, 1995; Cline, 1995). Coexistence of immiscible hypersaline liquid and vapor has been ubiquitously demonstrated in numerous fluid inclusion studies (Roedder, 1984), which also show that the liquid phase is enriched in Na, K, and Fe chlorides, giving rise to salinities of 35 to 70 wt percent NaCl equiv (e.g., Roedder, 1971; Nash, 1976; Eastoe, 1978; Bodnar, 1995), whereas the vapor phase contains acidic volatile species, preeminently SO_2, H_2S, CO_2, HCl, and any HF (e.g., Giggenbach, 1992, 1997). Fluid inclusion microanalysis and experimental studies reveal that, during phase separation, specific element suites selectively fractionate between the vapor and hypersaline liquid. In many cases, vapor can contain an appreciable amount of the Cu, Au, Ag, and S, plus much of the As, Sb, Te, and B, whereas Fe, Zn, Pb, Mn, and possibly Mo preferentially partition into the hypersaline liquid (Heinrich et al., 1999; Heinrich, 2005; Pokrovski et al., 2005, 2008, 2009; Williams-Jones and Heinrich, 2005; Simon et al., 2007; Audétat et al., 2008; Nagaseki and Hayashi, 2008; Wilkinson et al., 2008; Pudack et al., 2009; Seo et al., 2009).

Transport of Cu and probably also Au was for decades tacitly assumed to be in the form of chloride complexes in the hypersaline liquid phase (e.g., Holland, 1972; Burnham, 1967, 1997; Burnham and Ohmoto, 1980; Candela and Holland, 1986), but recent experimental work and fluid inclusion S analysis show that volatile S ligands ($H_2S \pm SO_2$) in the vapor phase can also act as major Cu- and Au-transporting agents (Nagaseki and Hayashi, 2008; Pokrovski et al., 2008, 2009; Seo et al., 2009; Zajacz and Halter, 2009). In contrast, Mo may be transported as different, possibly oxochloride complexes in the hypersaline liquid phase (Ulrich and Mavrogenes, 2008).

Current orthodoxy maintains that the early sodic-calcic alteration observed in some porphyry Cu deposits is a product of inflowing brine sourced from host-rock sequences (Carten, 1986; Dilles and Einaudi, 1992; Dilles et al., 1995; Seedorff et al., 2005, 2008), in keeping with theoretical predictions for fluids following heating paths under silicate-rock–buffered conditions (e.g., Giggenbach, 1984, 1997). Light stable isotope studies of sodic-calcic alteration in the Yerington district support the involvement of externally derived brine from the host sedimentary sequence (Dilles et al., 1992, 1995), although the albite-actinolite alteration there is magnetite destructive (Carten, 1986; Dilles et al., 1995). In other cases, however, there is evidence for an origin from hypersaline magmatic liquids, with the paucity of contained sulfide mineralization being due to excessively high temperatures and oxygen fugacities and the consequent deficiency of reduced S (John, 1989; Clark and Arancibia, 1995; Lang et al., 1995). A magmatic source would certainly be favored where sodic-calcic zones are metal bearing (see above).

As porphyry Cu systems cool through the 700° to 550°C temperature range, the single-phase liquid or, more commonly, coexisting hypersaline liquid and vapor initiate potassic alteration and perhaps the first metal precipitation in and around the early porphyry intrusions (e.g., Eastoe, 1978; Bodnar, 1995; Frei, 1995; Ulrich et al., 2001). Nevertheless, in many porphyry Cu deposits, it is fluid cooling over the ~550° to 350°C range, assisted by fluid-rock interaction, that is largely responsible for precipitation of the Cu, in low sulfidation-state Cu-Fe sulfide assemblages, plus any Au (e.g., Ulrich et al., 2001; Redmond et al., 2004; Landtwing et al., 2005; Klemm et al., 2007; Rusk et al., 2008a). In addition, upward decompression and expansion of the vapor phase causes rapidly decreasing solubility of the vapor-transported metals (Williams-Jones et al., 2002), as confirmed by their very low contents in high-temperature but atmospheric-pressure fumaroles (Hedenquist, 1995). Such a decrease in solubility leads to wholesale precipitation of the Cu-Fe sulfides together with Au, thereby potentially accounting for the typically shallow formation (Cox and Singer, 1992; Sillitoe, 2000) of Au-rich porphyry Cu deposits (Williams-Jones and Heinrich, 2005). The different Mo complexing (see above), probably assisted by progressive increase of the Mo/Cu ratio in the residual parental melt as crystallization proceeds (Candela and Holland, 1986), results in much of the molybdenite being precipitated not only later than but also spatially separate from the bulk of the Cu ± Au (see above).

Potassic alteration and associated metal deposition are initiated under near-lithostatic conditions and involve extensive hydraulic fracturing of the ductile rock at high strain rates (Fournier, 1999) to generate the pervasive stockwork veining (Burnham, 1979): a process that may give rise to large increases in rock volume (Cathles and Shannon, 2007). The single-phase liquid, the mineralizer in deeply formed porphyry Cu deposits, may generate the relatively uncommon EDM halo veinlets (Rusk et al., 2008a; Proffett, 2009), whereas the two-phase fluid produces the more common A- and B-type quartz veinlets (e.g., Roedder, 1984, and references therein). The local occurrence of vein dikes (see above), as well as recognition of coexisting melt and aqueous fluid inclusions in early quartz veinlets (Harris et al., 2003), confirms that magma and mineralizing fluid commonly coexist, although markedly different densities dictate that they typically separate. The stockwork veinlets control and focus continued fluid ascent, with partial dissolution of quartz during cooling through its retrograde solubility field (<~550-400°C at pressures <~900 b; Fournier, 1999) enhancing the permeability of the A-type quartz veinlets during at least some of the Cu-Fe sulfide precipitation (Rusk and Reed, 2002; Redmond et al., 2004; Landtwing et al., 2005); synmineral faulting and fracturing may play a similar role. The quartz-veined cores of potassic zones remain barren where temperatures are too high to permit appreciable Cu-Fe sulfide and associated Au deposition, potentially giving rise to the bell- and cap-shaped ore zones described above (e.g., Bingham, Resolution, and Batu Hijau; Babcock et al., 1995; Ballantyne et al., 2003; Setyandhaka et al., 2008). Fluid pressures may fluctuate from lithostatic to hydrostatic during porphyry Cu formation (e.g., Ulrich et al., 2001), as a result of both repetitive fracture propagation and sealing and reductions in confining pressure consequent upon surface degradation (see below). These pressure variations may induce changes in the fluid phases present and consequent remobilization as well as precipitation of metals (e.g., Klemm et al., 2007; Rusk et al., 2008a). Magmatic-hydrothermal brecciation may be triggered by sudden release of fluid overpressures caused by roof failure above large, expanding vapor bubbles (Norton and Cathles, 1973; Burnham, 1985), particularly near the ductile-brittle transition (Fournier, 1999).

During the protracted potassic alteration event(s) that affect the early and intermineral porphyries and their immediate wall rocks, heated external water, largely meteoric but possibly containing a connate component (e.g., Bingham; Bowman et al., 1987), generates the peripheral propylitic alteration, mainly by moderate-temperature hydration reactions (Meyer and Hemley, 1967). Convective circulation of the external water takes place where rock permeabilities are adequate (Fig. 14): a process that acts as a potent cooling mechanism for porphyry Cu systems (Cathles, 1977), particularly after parental intrusions have crystallized and no longer exsolve magmatic fluid.

The voluminous vapor readily separates from the coexisting hypersaline liquid and, because of its lower density, ascends buoyantly into the 1- to 2-km-thick rock column above the porphyry intrusions (e.g., Henley and McNabb, 1978; Hedenquist et al., 1998; Fig. 14). Progressive disproportionation of the contained SO_2 (to H_2SO_4 and H_2S) once it and HCl (plus any HF) condense into ground water (Giggenbach, 1992; Rye, 1993) generates the extremely low pH fluid responsible for the high degrees of base leaching involved in advanced argillic lithocap formation (e.g., Meyer and Hemley,

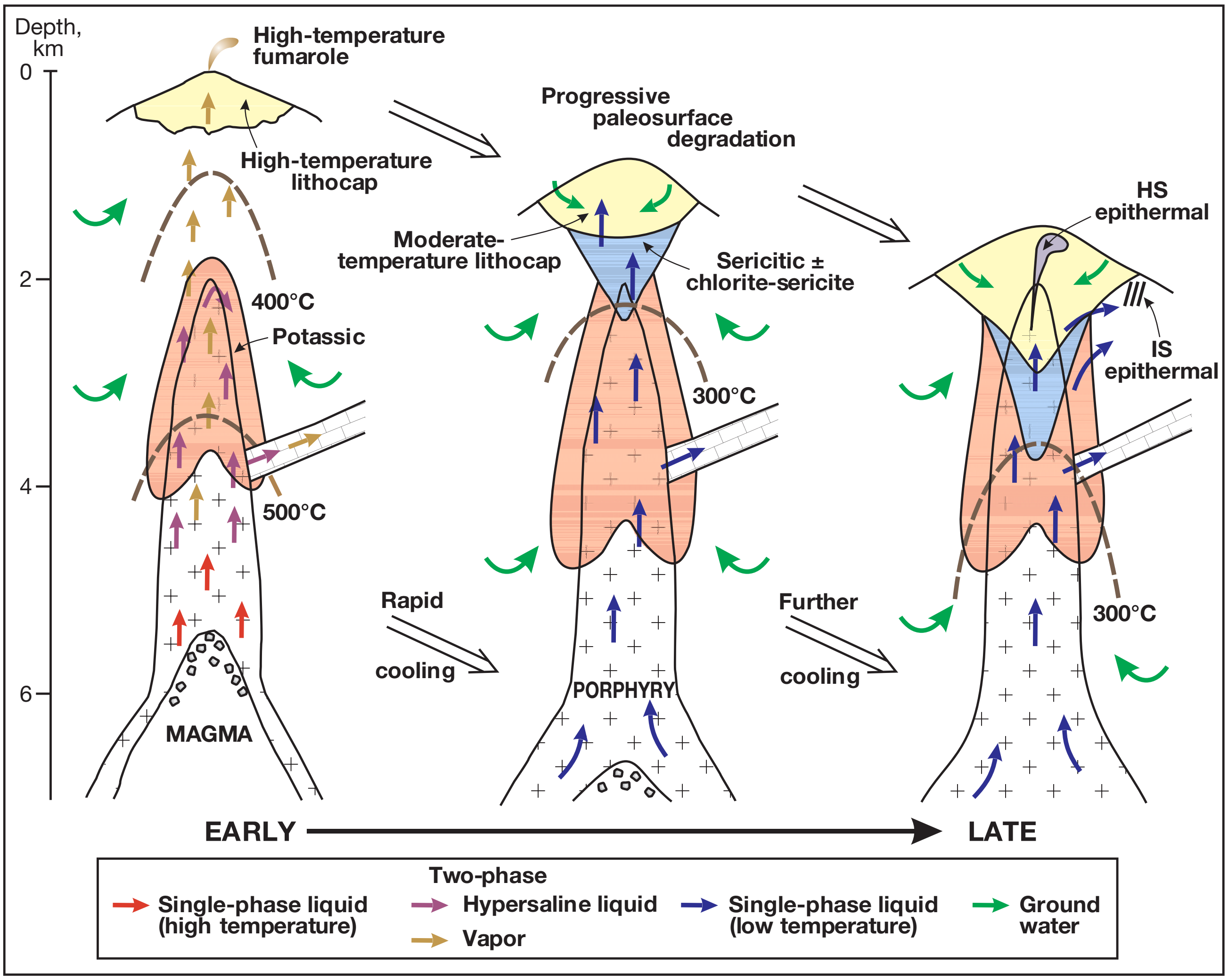

FIG. 14. Schematic time slices through the telescoped porphyry Cu system illustrated in Figures 6 and 10 to show the evolution of the main magmatic fluid and alteration-mineralization types in concert with progressive downward magma solidification, cooling, and paleosurface degradation. At the early stage (left side), magma is present at the top of the parental chamber, a single-phase, low- to moderate-salinity liquid exits the magma and undergoes phase separation during ascent to generate immiscible hypersaline liquid and vapor, which generate potassic alteration plus contained low sulfidation-state porphyry Cu ± Au mineralization. The upward-escaping, low-pressure vapor that does not attain the paleosurface as high-temperature fumaroles (e.g., Hedenquist, 1995; Hedenquist et al., 1993) forms acidic condensate to produce generally barren advanced argillic alteration. As magma solidification advances downward (middle), the entire system progressively cools, and the rock can fracture in a brittle fashion on cooling below ~400°C (Fournier, 1999); at this stage, lithostatic gives way to hydrostatic pressure, and erosion (or some other mechanism) progressively degrades the paleosurface. Under these lower temperature conditions, sericitic ± chlorite-sericite alteration zones begin to form from a deeply derived, single-phase aqueous liquid generated by one or both of the methods (see text) postulated by Hedenquist et al. (1998) and Heinrich et al. (2004). Eventually (right side), the sericitic ± chlorite-sericite alteration may cause variable degrees of Cu ± Au removal, but hypogene Cu enrichment is also possible in the former. The same liquid continues upward into the lithocap where, upon cooling in an unbuffered environment, it evolves into a high sulfidation-state liquid; if properly focused, it may generate high-sulfidation (HS) epithermal deposits. Renewed neutralization of this same liquid on exiting the lithocap and/or aliquots of the deep liquid that bypass the lithocap entirely may give rise to peripheral intermediate-sulfidation (IS) epithermal mineralization. Based on modeling by Hedenquist et al. (1998), Sillitoe and Hedenquist (2003), and Heinrich (2005).

1967). Focused ascent of the reactive fluid through fault and other permeable conduits leads to generation of the vuggy, residual quartz cores (if pH is <2; Stoffregen, 1987), flanked by zoned advanced argillic halos (Table 2; see above) indicative of partial outward fluid penetration, neutralization, and cooling. However, because of the low pressure of the lithocap environment and, hence, low metal-transporting capability of the absorbed vapor (see above), the resultant acidic fluid is unlikely to produce much mineralization, thereby possibly accounting for the barren status of many lithocaps (e.g., Hedenquist et al., 1998, 2000; Heinrich et al., 2004; Heinrich, 2005).

Late porphyry Cu system evolution

As the underlying parental magma chambers progressively solidify and magma convection ceases, there are marked reductions in both the heat flux and aqueous fluid supply to the overlying porphyry Cu systems (Dilles, 1987; Shinohara and Hedenquist, 1997), effects that are accompanied by downward propagation of the lithostatic-hydrostatic transition (Fournier, 1999). Under these lower temperature conditions, the aqueous liquid phase exsolves more slowly from the still crystallizing magma and, in turn, advects more slowly and cools, such that it may not intersect its solvus. If this scenario is correct, a single-phase, low- to moderate-salinity (5–20 wt % NaCl equiv) liquid in the 350° to 250°C temperature range ascends directly from the parental chambers into overlying porphyry Cu systems (Shinohara and Hedenquist, 1997; Hedenquist et al. 1998; Fig. 14). Alternatively, a single-phase liquid may form, possibly after separation of some brine, by subsequent contraction of vapor of the same composition as it cools at elevated pressures above the critical curve of the fluid system (Heinrich et al., 2004; Heinrich, 2005). The low-salinity liquid, whose ascent is controlled by the preexisting quartz veinlet stockworks, synmineral faults, and permeability contrasts provided by steep intrusive contacts, appears to be responsible for the progressive formation of the chlorite-sericite and sericitic alteration, as well as continued advanced argillic alteration and the principal Cu and Au mineralization in the overlying lithocaps (Hedenquist et al., 1998; Heinrich et al., 2004; Rusk et al., 2008b).

Admixture of magmatic and meteoric fluids, with the latter dominant, was long considered necessary to produce sericitic alteration and the attendant low- to moderate-salinity liquid, i.e., 5 to 10× dilution of the hypersaline liquid (e.g., Sheppard et al., 1971; Taylor, 1974), but recent interpretations of stable O and H isotope data reveal that an exclusively magmatic fluid is quite capable of producing the chlorite-sericite and sericitic assemblages (Kusakabe et al., 1990; Hedenquist and Richards, 1998; Hedenquist et al., 1998; Watanabe and Hedenquist, 2001; Harris and Golding, 2002; Skewes et al., 2003; Rusk et al., 2004; Khashgerel et al., 2006). However, meteoric water involvement in late sericitic alteration is by no means precluded (e.g., Hedenquist et al., 1998; Harris et al., 2005), particularly on the margins of systems where the advecting magmatic liquid may entrain convecting meteoric water, although its formerly preeminent role in the porphyry Cu genetic model (e.g., Beane and Titley, 1981; Hunt, 1991) is now greatly diminished. Since chlorite-sericite alteration partially or totally reconstitutes potassic assemblages, and sericitic alteration does the same to potassic and/or chlorite-sericite assemblages, it is generally impossible to determine if the contained metals are inherited from the former sulfide assemblage(s) (e.g., Brimhall, 1979) or newly introduced in the ascendant, still magmatic-sourced aqueous liquid. However, apparent confinement of hypogene Cu enrichment (see above) to sericitic alteration overprinting rocks cut by quartz veinlet stockworks that formerly contained chalcopyrite ± bornite may suggest that a large component of the Cu in the newly generated high sulfidation-state assemblages is derived by relatively localized remobilization (Sillitoe, 1999b).

The base and precious metal deposit types in both carbonate and noncarbonate wall-rock lithologic units likely form from the same aqueous magmatic fluids that are involved in porphyry Cu alteration and mineralization, wherever there is provision of lateral fluid access from the porphyry stock or dikes via lithologic, structural, and/or hydrothermally induced permeability (Fig. 14). In the skarn environment, the early two-phase hypersaline liquid plus vapor is likely to be followed under declining temperature conditions by the single-phase liquid (e.g., Meinert et al., 1997, 2003; Fig. 14), from which the retrograde skarn Cu ± Au ± Zn, carbonate-replacement Cu or Zn-Pb-Ag-(Au), and sediment-hosted Au-(As-Sb) deposits are formed (e.g., Meinert et al., 1997, 2003; Heinrich, 2005).

High Zn, Pb, Ag, and Mn contents are recorded in hypersaline liquid inclusions from quartz veinlets formed during potassic alteration (Bodnar, 1995; Heinrich et al., 1999; Ulrich et al., 1999; Wilkinson et al., 2008), but these chloride-complexed metals (see above) remain in solution because they are not appreciably concentrated in the sulfides present in the main porphyry Cu orebodies. Cooling of the hypersaline liquid in contact with external wall rocks and dilution with meteoric water in the propylitic halos may be the main causes of Zn, Pb, Ag, and Mn precipitation (Hemley and Hunt, 1992), giving rise to the geochemical halos of these metals and, in some systems, localized vein concentrations (Jerome, 1966; Figs. 6, 10). The largest concentrations of peripheral Zn, Pb, and Ag are confined to systems hosted by receptive carbonate rocks, where fluid neutralization induces the precipitation of these metals in skarn and carbonate-replacement deposits (Seward and Barnes, 1997).

The fluid most likely to lead to appreciable high-sulfidation Au ± Ag ± Cu mineralization in the relatively barren, early-formed lithocaps is the low- to moderate-salinity, H_2S-rich, aqueous liquid that produces the underlying sericitic zones (Hedenquist et al., 1998; Heinrich et al., 2004; Heinrich, 2005; Pudack et al., 2009; Fig. 14). On entering the lithocap environment, this intermediate sulfidation-state liquid (forming chalcopyrite and tennantite at depth) becomes unbuffered and easily evolves to a higher sulfidation state on cooling (Einaudi et al., 2003; Sillitoe and Hedenquist, 2003). The Cordilleran massive sulfide lodes are localized where the liquid follows pronounced structural permeability spanning the sericitic to advanced argillic transition (Figs. 6, 10) or, less commonly, encounters reactive carbonate rocks (e.g., Baumgartner et al., 2008; Bendezú and Fontboté, 2009). However, much of the Au precipitates in the shallower parts of lithocaps because of the greater likelihood of sharp drops in Au solubility caused by either intense boiling in upflow conduits or admixture of the ascendant liquid with cool, inflowing ground water; in some cases, the latter appears to originate from the vadose zone (see below) where it was steam heated (Hedenquist et al., 1998; Heinrich, 2005, and references therein; Figs. 6, 14). These shallow Au precipitation processes may be particularly effective in permeable phreatic breccias created by boiling of the ascendant liquid, vapor buildup beneath silicified seals, and eventual catastrophic release, perhaps assisted by external triggers (faulting, seismic shaking, and/or deep intrusion contributing gases; e.g., Nairn et al., 2005).

The low- to moderate-salinity liquids responsible for high-sulfidation deposits in lithocaps may, under appropriate structural and hydrologic conditions, pass into adjoining, less-altered rocks and undergo sufficient neutralization and reduction during outward flow and wall-rock reaction to produce liquids appropriate for formation of intermediate-sulfidation epithermal deposits (Sillitoe, 1999b; Einaudi et al., 2003; Sillitoe and Hedenquist, 2003; Fig. 14). The abovecited examples of mineralogic transitions between high- and intermediate-sulfidation mineralization provide support for this mechanism. Alternatively, the deeply derived intermediate sulfidation-state liquids may bypass the lithocaps entirely and still produce intermediate-sulfidation mineralization at shallow epithermal levels (Sillitoe and Hedenquist, 2003, Fig. 14).

At paleowater tables, near the tops of the lithocaps and nearby areas, the liquid portion of the boiling high- and intermediate-sulfidation fluids follows hydrologic gradients, whereas the H_2S-bearing vapor (with H_2S contributed by the magma as well as SO_2 disproportionation) continues its ascent into the overlying vadose zones. There, it condenses into ground water to oxidize and produce the low-temperature, acidic fluid responsible for the blanketlike, advanced argillic alteration zones characteristic of the steam-heated environment (Sillitoe, 1993, 1999b; Fig. 10).

As the thermal regimes of porphyry Cu systems decay, shallowly generated alteration-mineralization types become telescoped over more deeply formed ones (e.g., Gustafson, 1978; Fournier, 1999; Heinrich et al., 2004; Williams-Jones and Heinrich, 2005; Rusk et al., 2008a), thereby causing the sequence of metal remobilization and reprecipitation events emphasized above. Indeed, the tops of porphyry intrusions may be subjected to at least four distinct alteration-mineralization events, commencing with potassic and ending with advanced argillic, as temperature fronts retreat downward (Fig. 14). The resultant telescoping is potentially more extreme, giving rise to deep penetration of advanced argillic alteration into porphyry stocks, where porphyry Cu systems undergo either rapid, synhydrothermal erosion under high uplift, pluvial or glacial conditions (Fig. 14) or, perhaps less commonly, gravity-induced sector collapse of any overlying volcanic edifices (Sillitoe, 1994; Perelló et al., 1998; Landtwing et al., 2002; Carman, 2003; Heinrich, 2005; Masterman et al., 2005; Rohrlach and Loucks, 2005; Pudack et al., 2009).

By the time that the late-mineral porphyry phases are added to porphyry Cu stocks or dike swarms, fluid ascent from the parental magma chambers has all but ceased, and K and metal availability is too limited to generate appreciable potassic alteration and mineralization. The only fluid present is of external origin and produces propylitic alteration similar to that in the earlier formed propylitic halos. Diatreme breccias are preferentially emplaced at this time because external water access to late-mineral magma bodies, a requirement for phreatomagmatic activity, is facilitated. End-stage, groundwater incursion into the hot porphyry Cu deposits leads to anhydrite veinlet formation, in conformity with the mineral's retrograde solubility (e.g., Rimstidt, 1997).

Exploration Implications

Target selection

When planning exploration programs for porphyry Cu ± Mo ± Au, skarn Cu ± Au, or high-sulfidation epithermal Au deposits, the preeminent ore types hosted by porphyry Cu systems, the choice is between selection of (1) mature, well-endowed Cu or Au belts, (2) emerging belts with less obvious metallogenic credentials but having at least one important deposit of the type that is sought, or (3) frontier terranes with geologic conditions that are perceived to imply potential. On the basis of recent exploration successes, the first choice has been shown to be a wise one, as witnessed by the strings of high-sulfidation Au and Au-rich porphyry discoveries that now define the El Indio-Maricunga belt in northern Chile and Cajamarca-Huaraz belt in northern Peru (Sillitoe, 2008), as well as discovery of the Resolution porphyry Cu-Mo deposit in the southwestern North American Cu province after a 100-year exploration history (Manske and Paul, 2002). To date, the second choice could be taken to have been less successful, as shown by the lack of economically significant discoveries in the vicinities of the major but isolated Bingham, Butte, Pebble, and Oyu Tolgoi districts, although greenfield exploration is in its infancy in the still poorly defined magmatic arcs that host the last two of these. However, the El Indio and Yanacocha high-sulfidation Au deposits were initially the isolated orebodies that led to eventual definition of the El Indio-Maricunga and Cajamarca-Huaraz belts, respectively. The third choice, frontier terranes, obviously involves higher risk but resulted in the recent discoveries of Pebble, Oyu Tolgoi, and Reko Diq, for example (Bouley et al., 1995; Perelló et al., 2001, 2008; Kirwin et al., 2003).

The empirical relationship between well-established magmatic (including postcollisional) arcs containing major, high-grade hypogene porphyry Cu and high-sulfidation Au deposits and contractional tectonic settings characterized by high surface uplift and denudation rates (see above) may prove to be a useful criterion for selection of underexplored arc segments with incompletely tested potential. Contractional settings are strongly suggested where entire arc segments possess only minor volcanic rock volumes contemporaneous with the development of porphyry Cu systems, particularly where lithocaps are widely preserved as evidence for shallow erosion. Contractional settings are also likely in belts or districts where porphyry Cu stocks or dike swarms are overprinted on precursor plutons or, in island-arc settings, where marine sedimentary rocks only slightly older than the porphyry Cu systems have been uplifted to ~1 km or more above sea level (Sillitoe, 1998). In arcs where volcanic rocks are abundant, large-volume ignimbrites, indicative of caldera formation, are taken to seriously downgrade porphyry Cu and related epithermal Au potential for the reason given above.

The clustering or alignment of both porphyry Cu and high-sulfidation Au deposits has been shown time and again to be a highly effective exploration concept. The recent major porphyry Cu-Mo ± Au discoveries in the productive Collahuasi (Rosario Oeste), Chuquicamata (Toki cluster; Rivera and Pardo, 2004; Fig. 3b), Escondida (Pampa Escondida), and Los Bronces-Río Blanco (Los Sulfatos; Fig. 5a) districts of Chile are all within <1 to 3 km of the previously known deposits, as are the several porphyry Cu-Au discoveries in the Cadia district (Holliday et al., 1999) and high-sulfidation Au discoveries in the Yanacocha district (Harvey et al., 1999) that were made since mining commenced. In any deposit cluster or alignment, the best deposit may be found first or only after several lesser discoveries have already been made (e.g., Hugo Dummett; Kirwin et al., 2003). Whether or not these and other deposit clusters and alignments owe their existence to fundamental faults or lineaments (see above; Richards, 2000), it is often not obvious if exploration—commonly focused on areas of pre- or postmineral cover—should target broadly equidimensional deposit arrays or arc-parallel or arc-transverse alignments, particularly when only one or two deposits have been defined. However, local structural observations, perhaps interpreted from geophysical responses, may assist in this regard, although this approach played no part in the recent discoveries noted above in the Chilean porphyry Cu districts.

Even cursory inspection of Figure 6 shows clearly that erosion level is a fundamental control on the mineralization types that may be anticipated to occur in porphyry Cu systems. If porphyry Cu deposits concealed beneath advanced argillic alteration are the principal target, then deeply eroded lithocaps in which quartz-pyrophyllite ± muscovite ± andalusite alteration is prominent are best selected (e.g., El Salvador). Any exposed A-type quartz veinlet stockworks overprinted by sericitic and/or advanced argillic assemblages immediately pinpoint the spots for initial scout drilling (Sillitoe, 1995a). In contrast, D-type veinlets may be up to 1 km laterally away from the target. Nevertheless, bearing in mind that most observed lithocaps are only erosional remnants, exploration should focus first around their peripheries in case a porphyry Cu deposit has already been exposed. However, should the search be for high-sulfidation Au deposits, the shallow parts of lithocaps may have the best potential for the discovery of large, albeit commonly low-grade orebodies. The existence of even minor erosional remnants of steam-heated horizons and their chalcedonic bases, generated above and at paleowater tables, respectively, guarantees that the appropriate near-surface level is preserved (Sillitoe, 1999b).

Assessment of the likely host-rock lithologic units is also important during initial appraisals of porphyry Cu belts and districts. Obviously, major skarn, carbonate-replacement, and sediment-hosted Au deposits can only be expected where relatively thinly bedded, commonly silty carbonate rocks are present. Large, high-grade porphyry Cu deposits seem to be favored by the "pressure-cooker effect" provided by impermeable wall rocks, including massive, thickly bedded carbonate sequences (e.g., Grasberg), a situation that can also lead to the formation of blind high-grade deposits overlain by largely unaltered rocks (e.g., Hugo Dummett, Ridgeway, Pueblo Viejo). Exceptionally ferrous Fe-rich rocks, relatively uncommon in most arc terranes, also appear to assist with development of high hypogene Cu grades as well as maximizing the wall rock-hosted component of the deposit (e.g., El Teniente, Resolution, Oyu Tolgoi). Highly permeable, noncarbonate host rocks may promote lateral fluid channeling, which may lead to generation of distal ore types other than structurally controlled veins (e.g., Andacollo). Porous and permeable volcaniclastic and epiclastic sequences also favor large-tonnage orebody development in the lithocap environment, especially where they happen to be shallowly located with respect to paleosurfaces.

The large size of some porphyry Cu systems, with maximum radii of ~8 km (e.g., Fig. 9) and maximum areal extents approaching 100 km^2 (Singer et al., 2008), complicates their effective exploration because attention is unavoidably focused on the more prominently altered parts, such as pyrite-bearing porphyry Cu mineralization, pyrite halos, and pyrite-rich lithocaps. As a consequence, the most distal and subtle ore types, sediment-hosted Au on the fringes of carbonate rock-hosted districts and Zn-Pb-Ag ± Au–bearing intermediate-sulfidation epithermal veins and carbonate-replacement bodies on the fringes of lithocaps attract less attention and may be easily missed.

It also needs to be emphasized that few porphyry Cu systems, whatever their exposure level, contain the full spectrum of potential ore types depicted in Figure 6, although the Bingham district with its porphyry Cu-Au-Mo, Cu-Au skarn, carbonate-replacement Zn-Pb-Ag-Au, and sediment-hosted Au deposits (Babcock et al., 1995) and the more shallowly exposed Lepanto district with its porphyry Cu-Au, high-sulfidation Cu-Au-Ag, and intermediate-sulfidation Au-Ag-Cu deposits (Hedenquist et al., 2001) are exceptionally well endowed in this regard. Nevertheless, many systems contain only one or two deposit types rather than a full zonal array (Table 3), with the presence of the more distal ore types at either porphyry Cu or lithocap levels being independent of the size and grade of the porphyry Cu deposits or prospects. Therefore, recognition of even weakly developed mineralization of a single type may help to direct exploration for potentially higher grade mineralization of other types elsewhere in the system. Furthermore, Mo- as well as Au-rich porphyry Cu deposits may be associated with Au-endowed lithocaps (e.g., Nevados del Famatina district; Lozada-Calderón and McPhail, 1996), although lithocaps above any porphyry Cu deposit may lack appreciable high-sulfidation mineralization (e.g., Red Mountain; Corn, 1975; Quinlan, 1981), at least in their preserved parts.

Target appraisal

Notwithstanding the typical occurrence model depicted in Figure 6 and taking into account the critical importance of erosion level, the innumerable variations on the porphyry Cu genetic theme result in a broad spectrum of three-dimensional intrusion, breccia, alteration, and mineralization geometries (e.g., Gustafson and Hunt, 1975). At first glance, using representative cross sections of alteration at four high-grade hypogene porphyry Cu deposits as examples (Fig. 15), these varied geometries are not easy to relate to a standard geologic model. Each individual deposit or prospect must be carefully constructed using surface mapping and core logging, with particular attention paid to the temporal as well as spatial relationships of its constituent parts. Only then will the positive and negative geologic features and, hence, its overall potential become evident.

In most magmatic arc terranes, it is roughly estimated that >90 percent of explored porphyry systems lack Cu and Au concentrations with foreseeable potential, commonly because the ore-forming processes, from magma generation through to alteration and mineralization, were less than fully optimized (e.g., Richards, 2005). Some critical step in the genetic sequence was either poorly developed or entirely missing. For example, porphyry Cu prospects containing only weakly developed potassic alteration and A-type quartz veinlets in their central parts, indicating a deficiency of early-stage magmatic fluids, are typically subore grade. Similarly, proximal skarns lacking hydrous, retrograde overprints are unlikely to host significant Cu-Au deposits. Lithocaps dominated by quartz-alunite or quartz-pyrophyllite alteration but without

TABLE 3. Representative Examples of Various Mineralization-Type Combinations in Porphyry Cu Systems[1]

District, location	Porphyry deposit	Proximal skarn	Distal skarn	Carbonate-replacement	Sediment-Hosted	Peripheral veins/mantos	High-sulfidation	Intermediate-sulfidation	Reference(s)
Bingham, Utah	Bingham Cu-Au-Mo	Carr Fork and North Shoot Cu-Au		Zn-Pb-Ag-(Au) bodies	Barneys Canyon and Melco Au				Babcock et al. (1995), Cunningham et al. (2004)
Copper Canyon, Nevada	*Copper Canyon stock Cu-Mo-Au*[2]	East and West Zone Cu-Au[2]	Lower Fortitude Au[3]	*Zn-Pb occurrences*					Theodore et al. (1982), Wotruba et al. (1988), Cary et al. (2000)
Superior, Arizona	Resolution Cu-Mo			Superior Cu-Ag					Paul and Knight (1995), Manske and Paul (2002)
Yanacocha, Peru	Kupfertal Cu-Au						Yanacocha Norte Au	*Au occurrences*	Harvey et al. (1999)
Antamina, Peru	*Cu-Mo occurrence*	Antamina Cu-Zn-Mo-Ag-Au				*Zn-Pb-Ag veins*			Love et al. (2004), Redwood (2004)
Potrerillos, Chile	Mina Vieja Cu-Mo-Au	San Antonio Cu			Jerónimo Au		Silica Roja Au	El Hueso Au	Thompson et al. (2004)
Andacollo, Chile	Carmen de Andacollo Cu-Mo-Au					Andacollo Oro Au			Reyes (1991)
Lepanto, Philippines	Far Southeast Cu-Au						Lepanto Cu-Au-Ag	Victoria and Teresa Au-Ag-Cu	Hedenquist et al. (2001)
Wafi-Golpu, Papua New Guinea	Wafi Cu-Au						A and Link Zone Au	Link Zone Au	Ryan and Vigar (1999)
Sepon, Laos	*Padan and Thengkham Mo-Cu occurrences*	*Cu-Au occurrences*		Khanong and Thengkham South Cu-Au	Discovery, Nalou, etc. Au				R.H. Sillitoe (unpub. repts., 1994–1999), Smith et al. (2005)
Bau, Malaysia	*Cu-Au occurrences*	*Cu-Au occurrences*		*Zn-Pb occurrences*	Bau Au				Percival et al. (1990), Sillitoe and Bonham (1990)
Recsk, Hungary	Recsk Deeps Cu-Au-Mo	Recsk Deeps Cu-Au	Recsk Deeps Zn-Cu	Recsk Deeps Zn-Pb			Lahóca Cu-Au	Parád Au-Ag	Kisvarsanyi (1988), Földessy and Szebényi (2008)

[1] Minor occurrences italicized
[2] Porphyry Cu formation likely inhibited by reduced nature of the host porphyry (Meinert, 2000)
[3] Proximal to distal skarn transition spanned by Phoenix and Greater Midas pits (Cary et al., 2000)

appreciable development of vuggy, residual quartz and associated silicification, perhaps because fluid pH was too high or exposure level is too deep, are much less likely to contain major high-sulfidation Au deposits, although Pueblo Viejo provides a salutary exception (Kesler et al., 1981; Sillitoe et al., 2006).

Commonly, the highest grade and most coherent porphyry Cu deposits are those that retain their early porphyry phases and potassic alteration assemblages—with which much of the metal content is initially introduced—in essentially unmodified form. This is particularly the case for Au, which tends to be removed and dissipated during the formation of lower temperature, pyrite-bearing alteration assemblages (Gammons and Williams-Jones, 1997; Sillitoe, 2000; Kesler et al., 2002). Thermal regimes that permit vertically extensive ore zone development in potassic zones commonly have greater size potential than those that were excessively hot internally, thereby inhibiting sulfide precipitation and giving rise to large, low-grade or barren cores; the exception is where the resultant shell-like orebodies are areally extensive and thick (e.g., Bingham and Resolution; Babcock et al., 1995; Ballantyne et al., 2003). The enhanced Cu ± Au tenors of many bornite-rich

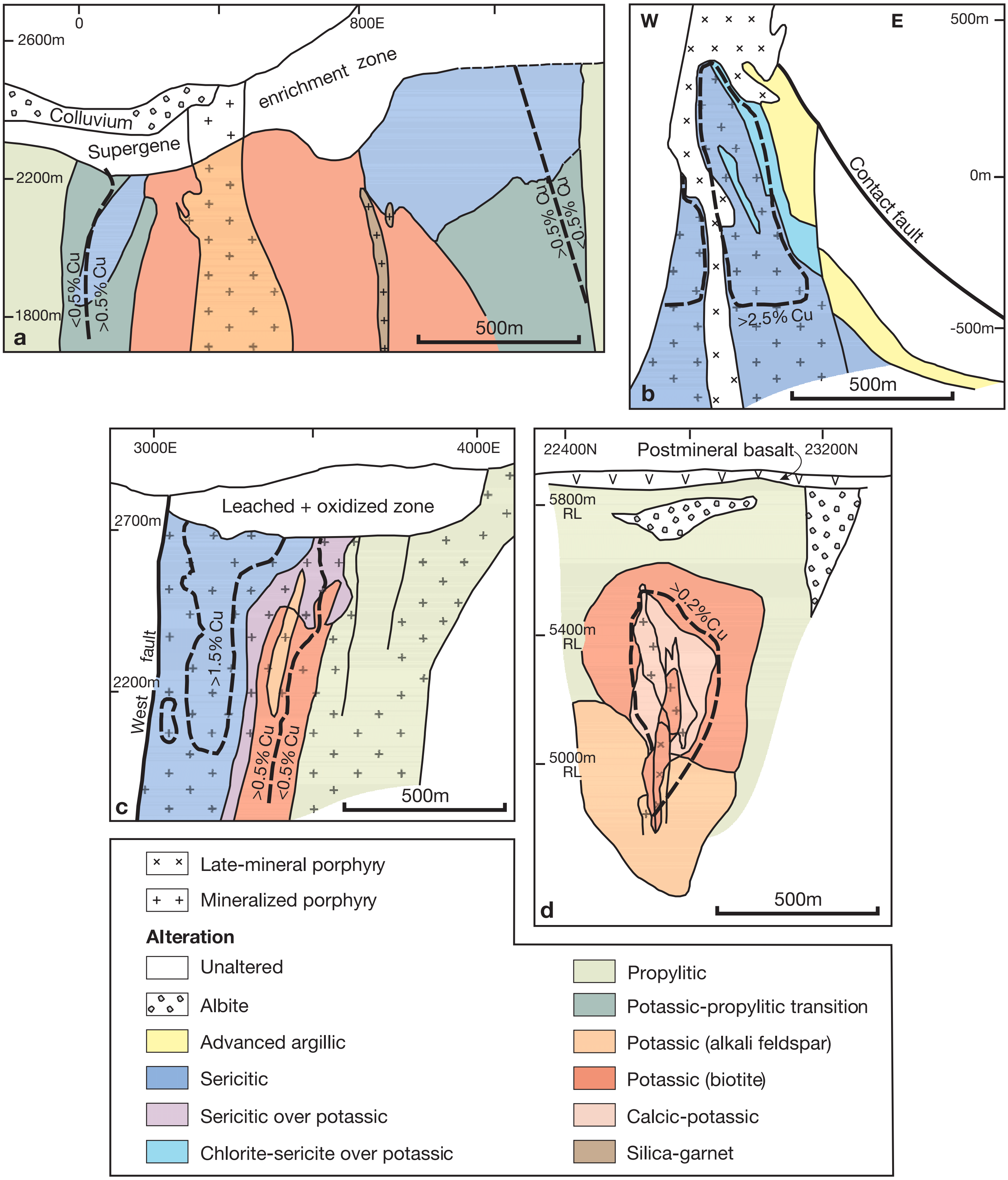

FIG. 15. Simplified sections through high-grade hypogene porphyry Cu deposits to illustrate the wide variation in alteration zoning patterns and their relationship to porphyry intrusions and Cu tenor. a. El Teniente, Chile (from Cannell et al., 2005). b. Hugo Dummett North at Oyu Tolgoi, Mongolia (from Khashgerel et al., 2008). c. Chuquicamata, Chile (from Ossandón et al., 2001). d. Ridgeway, New South Wales, Australia (from Wilson et al, 2003). Note that the high-grade (>1.5% Cu) zone at Chuquicamata includes a substantial, but unquantified, contribution from supergene enrichment.

porphyry Cu deposits provide the justification for deeper drill testing of both bornite-bearing and bornite-free potassic alteration zones that are judged to have been only relatively shallowly explored (e.g., Esperanza; Perelló et al., 2004b). Abundant hydrothermal magnetite is a good indicator of potentially Au-rich porphyry Cu deposits (Sillitoe, 1979), and the presence of banded quartz veinlets may be used to identify most, but not all, Cu-poor porphyry Au deposits (Vila and Sillitoe, 1991).

Where lower grade intermineral intrusions and barren late- and postmineral intrusions or diatremes are volumetrically important, the original mineralized rock volumes may be physically disrupted and ore-zone geometries radically changed and generally rendered less continuous. Where intense chlorite-sericite or sericitic alteration overprints are developed, the reconstitution of potassic alteration may result in either reduction or complete stripping of original metal contents. Furthermore, even where appreciable Au is retained in chlorite-sericite assemblages, flotation recoveries are commonly lower (<60%) than for ore from potassic zones (>80%) because some of the Au originally in solid solution and encapsulated in and attached to chalcopyrite ± bornite becomes linked to introduced pyrite (Sillitoe, 2000).

In the more highly telescoped systems, where sericitic and/or advanced argillic assemblages overprint appreciable volumes of potassic and/or chlorite-sericite alteration within porphyry intrusions, the ensuing effects can be varied. Where the sericitic alteration is superimposed on quartz veinlet stockworks, Cu contents in the form of the high sulfidation-state Cu sulfides may be increased by hypogene enrichment (e.g., Wafi-Golpu). However, if the overprinted high-sulfidation assemblages also contain appreciable arsenical sulfosalts, a situation that becomes increasingly likely upward in most systems, the resultant mineralization is less desirable because it is not only refractory if subjected to bacterial heap leaching but also generates As-rich flotation concentrates that may prove difficult to market.

Although many hydrothermal breccias, like the late diatremes mentioned above, are commonly diluents to ore, some magmatic-hydrothermal breccias give rise to anomalously high-grade rock volumes despite their intermineral timing. Furthermore, magmatic-hydrothermal breccia cemented mainly by quartz, tourmaline, and pyrite may be zoned downward over hundreds of meters to chalcopyrite-rich material, which is likely to persist into any underlying biotite-cemented breccia (e.g., Los Bronces-Río Blanco; Vargas et al., 1999; Fig. 8).

Interrelationships between porphyry intrusions and carbonate host rocks can influence the form and size of skarn deposits, typically with above-average Cu tenors. Where steeply dipping, receptive carbonate rock sequences abut steep porphyry stock contacts, vertically extensive proximal skarn bodies may form (e.g., the >1,600-m extent of the Ertsberg East (Gunung Bijih) Cu-Au deposit, Indonesia; Coutts et al., 1999). Unusually large, laterally extensive proximal skarn bodies may form preferentially where suitable carbonate host rocks abut the tops of porphyry stocks (e.g., Antamina; Redwood, 2004). Structural permeability linking porphyry stocks to the fringes of carbonate rock-dominated districts seems to be a requirement for formation of substantial sediment-hosted Au deposits (e.g., Bingham district; Cunningham et al., 2004).

In lithocaps, permeable lithologic units are an especially important control of the largest high-sulfidation epithermal Au deposits, as mentioned previously, in contrast to tight rocks, such as little-fractured lava domes and flows, which typically host smaller, fault- and fracture-controlled deposits. Any carbonate rocks affected by the lithocap environment may be particularly receptive. However, through-going structural (fault and fracture network) and hydrothermal (phreatic breccia and vuggy, residual quartz) permeability is probably the most critical requirement for the development of important high-sulfidation deposits; otherwise, inadequate late-stage aqueous liquids from the cooling parental magma chambers gain access to the lithocaps. Lateral transfer of such liquids beyond lithocaps to form intermediate-sulfidation epithermal deposits is also dependent on the existence of suitable permeability, which in a few cases is the direct continuation of that utilized by the contiguous high-sulfidation mineralization (e.g., Colquijirca, Tintic).

Concluding Statement

Porphyry Cu deposits are arguably the most studied and potentially best known and understood ore deposit type (e.g., Seedorff et al., 2005), and their relationships with the skarn environment have been appreciated for many years (Einaudi et al., 1981; Einaudi, 1982). Only in the last decade or so, however, have the physicochemical connections with the high- and intermediate-sulfidation epithermal environment within and around overlying lithocaps been clarified (e.g., Hedenquist et al., 1998, 2001). The current state of geologic understanding allows explorationists to use a combination of empirical and genetic models with ever-increasing degrees of confidence (Thompson, 1993; Sillitoe and Thompson, 2006). Furthermore, the current geologic knowledge base permits meaningful deployment of sophisticated geochemical and geophysical techniques in some exploration programs (e.g., Kelley et al., 2006; Holliday and Cooke, 2007).

Nevertheless, there is still a great deal to learn, a fact underscored by the relatively recent appreciation of the contrasting metal contents of coexisting hypersaline liquids and vapors (Heinrich et al., 1999; Ulrich et al., 1999) and experimental determination of volatile S complexes as potentially important Cu- and Au-transporting agents throughout porphyry Cu systems (Williams-Jones et al., 2002; Nagaseki and Hayashi, 2008; Pokrovski et al., 2008, 2009). A short, personalized selection of outstanding questions includes the following: (1) what are the fundamental mantle and/or crustal factors that dictate whether youthful arc segments are endowed with giant porphyry Cu systems (e.g., central Andes), only incipiently developed systems (e.g., Cascades, western United States), or none at all (e.g., Japan)? (2) what are cross-arc lineaments, and can they be demonstrated to play a truly influential role in the localization of porphyry Cu systems? (3) how important are mafic magmas in the development of the parental magma chambers beneath porphyry Cu systems, and what material contributions do they make to the systems themselves? (4) how is the single-phase magmatic liquid transferred from the parental magma chambers to porphyry Cu stocks or dike swarms, and what distance can be travelled by this fluid between exiting the chambers and eventual phase separation? (5) what are the deep processes that result in some porphyry Cu systems being apparently short lived while others may remain at least intermittently active for up to ≥5 m.y.? (6) why do some porphyry Cu deposits develop large and high-grade magmatic-hydrothermal breccias, whereas others have only minor examples or none at all? (7) if externally derived, nonmagmatic brine is responsible for at least some examples of sodic-calcic alteration, how does it access the cores of some porphyry Cu deposits between early porphyry emplacement and magmatic fluid ascent responsible for initiation of potassic alteration (and locally sinuous A-type quartz veining)? (8) what controls metal depletion versus enrichment during chlorite-sericite and sericitic overprints? (9) what are the main mechanisms controlling the bulk Cu/Au/Mo ratios of porphyry Cu deposits? (10) why is Au transported to the distal limits of only a few porphyry Cu systems for concentration in sediment-hosted deposits, and why are most of these apparently small compared to virtually identical Carlin-type deposits (Cline et al., 2005)? (11) what is the fluid regime responsible for metal zoning in lithocaps, and why are so many lithocaps apparently barren? (12) why do only a few lithocaps appear to develop intermediate-sulfidation epithermal precious metal deposits on their peripheries?

Effective study of these and other problems will require field-based geochemical and geophysical work and an array of evermore sophisticated laboratory equipment for high-precision fluid inclusion and trace element analysis, isotopic determinations, isotopic dating, and experimental work on fluid evolution and metal transport. But more fundamentally, however, we require better and more detailed documentation of geologic relationships in porphyry Cu systems worldwide, at all scales from the thin section to the entire system, and with greater emphasis on the regional to district context, particularly the relationship to igneous evolution. And these geologic observations must further emphasize the relative timing of intrusion, brecciation, alteration, and mineralization events because isotopic dating techniques do not and may never have the required resolution. It is acquisition of this geologic detail that is going to enable better application of laboratory techniques and, hopefully, further clarify the localization and evolutionary histories of porphyry Cu systems as well as the fundamental controls on large size and high hypogene grade.

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