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Chapter 6

Magmatic and Structural Controls on the Development of Porphyry Cu ± Mo ± Au Deposits

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Abstract

Porphyry Cu ± Mo ± Au deposits require the coincidence and positive interaction of a series of individually commonplace geological processes. They, and all their genetically associated deposits, are a natural consequence of convergent margin magmatism, and reflect the dynamic interplay between magmatic, hydrothermal, and tectonic processes. Magmas generated during subduction rise into the upper crust, commonly along zones of lithospheric weakness, where they pond in tabular magma chambers at depths of 6 km or deeper. The chambers grow laterally by chamber floor depression (cantilever mechanism) and some roof lifting (piston mechanism). Apophyses rise from the parental magma chamber and intrude to within 1 to 3 km of the surface, where they may undergo volatile exsolution and crystallization as porphyritic stocks. Emplacement of porphyry stocks is facilitated by structural anisotropy in the roof rocks. Ascending hydrothermal fluids exsolved from the porphyry stocks and the underlying parental magma chamber are focused into the cupola, taking advantage of vertical structural and rheological anisotropies introduced either before or during porphyry emplacement.

From a structural standpoint, three recurrent processes enhance permeability in the form of fracture or breccia networks through which hydrothermal fluids pass and precipitate minerals. Fracture-producing events are related to intrusion of pre-, syn-, and post-mineral porphyry stocks or dikes to near-surface depths (1–3 km), phase separation and volume expansion of a hydrothermal fluid through a variety of mechanisms, and tectonically induced failure. Concentric and radial fracture patterns reflect magmatic processes whereas more linear arrays of veins reflect tectonic influences. The resulting different vein arrays are commonly vertically and temporally distributed in the porphyry system; concentric and radial arrays are more common above or in the upper parts of the stocks, whereas linear arrays dominate at depth, forming as the system cools and the pluton solidifies. Orthogonal and conjugate arrays of veins characterize all scales and all parts of porphyry systems. Veins from a particular paragenetic stage do not have unique orientations, but rather occur with all orientations typical of that system. The common conjugate to orthogonal inter-vein relationships in porphyry Cu deposits requires repetitive exchange of principal stress orientations, events that are facilitated by conditions of low differential horizontal stress. Such stress conditions indicate that many porphyry Cu deposits form in specific environments where the magmatic arc is under a near-neutral stress state. These conditions occur either in areas removed from active deformation, or during periods of stress relaxation and low strain in the magmatic arc. Achievement of these conditions in time and space is likely to be infrequent and transitory during the life of a convergent margin, which may explain the spatial and temporal clustering of deposits in large porphyry districts.

Introduction

PORPHYRY Cu ± Mo ± Au deposits, hereafter referred to as porphyry Cu deposits, are associated with shallow-level plutonic complexes emplaced in magmatic arcs at convergent plate margins (Figs. 1, 2A; Sillitoe, 1972, 1997; Sutherland Brown, 1976; Titley, 1982; Sawkins, 1990). Since their first recognition as important sources of metals (Ransome, 1904, 1919; Lindgren, 1905, 1933, 1937; Emmons, 1927), much work has been devoted to understanding the formation of these deposits from a physical and geochemical standpoint (Burnham, 1967, 1979; Gustafson and Hunt, 1975; Cathles, 1977; Burnham and Ohmoto, 1980; Beane and Titley, 1981; Candela, 1989; Cline and Bodnar, 1991; Hedenquist and Richards, 1998). In addition, the question of why there is considerable variation in the size of these types of deposits has been pondered (Clark, 1993). Other studies have focused on structural controls on vein and fracture distributions within porphyry Cu deposits, and their spatial and paragenetic relationships to porphyry stocks (Rehrig and Heidrick, 1972, 1976; Gustafson and Hunt, 1975; Titley and Heidrick, 1978; Heidrick and Titley,

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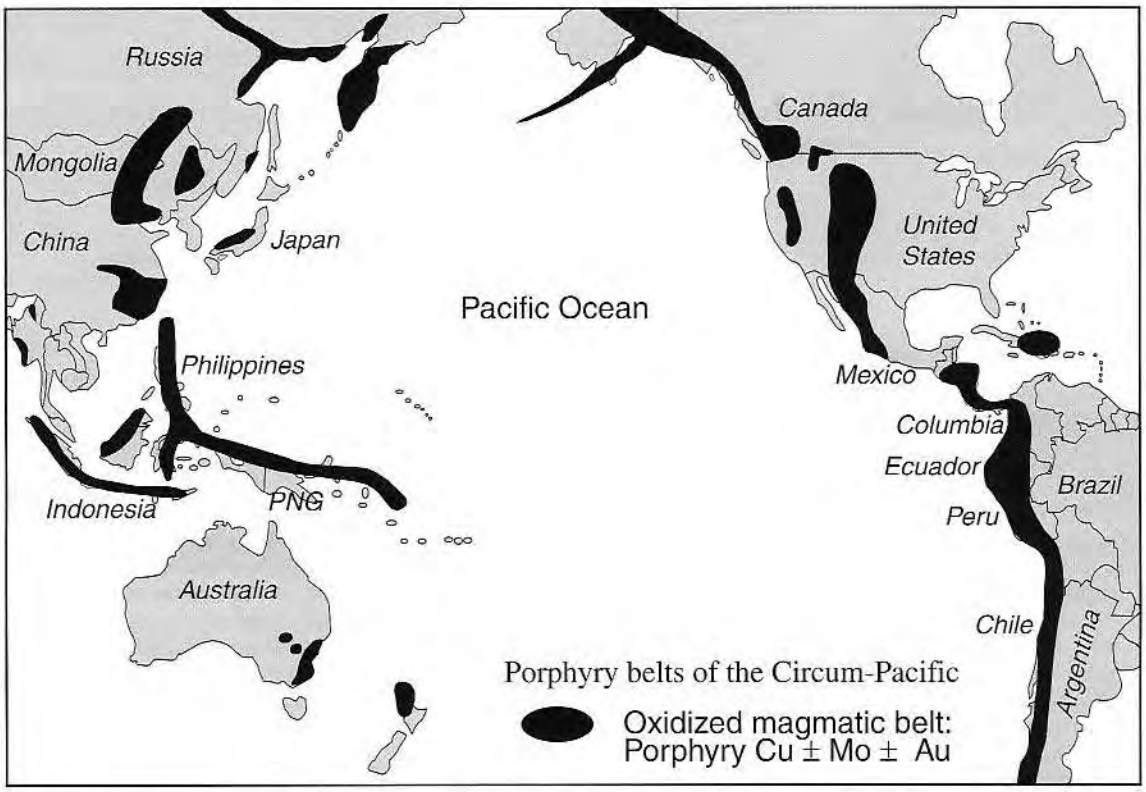

FIG. 1. Distribution of Mesozoic and Cenozoic porphyry Cu ± Mo ± Au deposits along convergent plate margins in the circum-Pacific region. PNG = Papua New Guinea. Adapted from Thompson (1995).

1982; Wilkinson et al., 1982; Lindsay et al., 1995). These studies demonstrate that porphyry Cu deposits are influenced not only by magmatic and hydrothermal processes, but also by regional tectonics at the time of formation. In fact, the prevailing tectonic setting and structural framework may strongly influence not only the size and form of a porphyry system (Sillitoe, 1994, 1997; Skewes and Stern, 1995) but also its location (Rehrig and Heidrick, 1972, 1976; Titley and Heidrick, 1978; Heidrick and Titley, 1982; Richards et al., 2001).

This chapter reviews the interplay between magmatism and dynamic structural environments in the development of porphyry Cu deposits. Critical to this goal is the general model of porphyry Cu deposits, which highlights the need for constructive interaction between magmatic, magmatic-hydrothermal, and tectonic events in the formation of porphyry Cu deposits. It is equally important to consider emplacement mechanisms of shallow-level plutonic complexes in the upper crust because porphyry Cu deposits are restricted to these environments. Throughout this chapter, we draw liberally on well-described examples of porphyry Cu deposits to illustrate various points. Most examples are from the southwestern United States and from the Andes of Perú and Chile. This geographic focus is not intended to indicate that other porphyry Cu provinces are insignificant, but merely reflects past history of geologic investigations, availability of literature, and our personal experiences.

General Porphyry Model

Physicochemical model

Important characteristics of porphyry Cu deposits summarized herein are based on the primary physiochemical studies of Lindgren (1937), Hemley and Jones (1964), Sheppard et al. (1969, 1971), Lowell and Guilbert (1970), Rose (1970), Roedder (1971), Sillitoe (1973), Gustafson and Hunt (1975), Hollister (1975), Sheppard and Gustafson (1976), Sutherland Brown (1976), Eastoe (1978), and Reynolds and Beane (1985), and reviews by Gustafson (1978), Beane and Titley (1981), Titley and Beane (1981), Titley (1982), Hunt (1991), Kirkham and Sinclair (1996), and Hedenquist and Richards (1998).

Porphyry-type systems are so-named because of their spatial and genetic association with porphyritic intrusions (Fig. 2A). Commonly, such intrusions are characterized by phenocryst assemblages consisting of one or more of the minerals quartz, K feldspar, plagioclase, hornblende, and biotite (rarely pyroxenes and olivine) in a fine-grained matrix. Rock compositions vary from granitic to dioritic or rarely gabbroic, and are of calc-alkaline to alkaline parentage (Brown, 1982). Parental magmas are believed to be moderately water-rich as indicated by the presence of hydrous phenocryst phases such as amphibole and biotite, whose stability requires at least 3 wt percent H_2O in the melt (Holland, 1972; Burnham, 1979). Water concentrations are probably

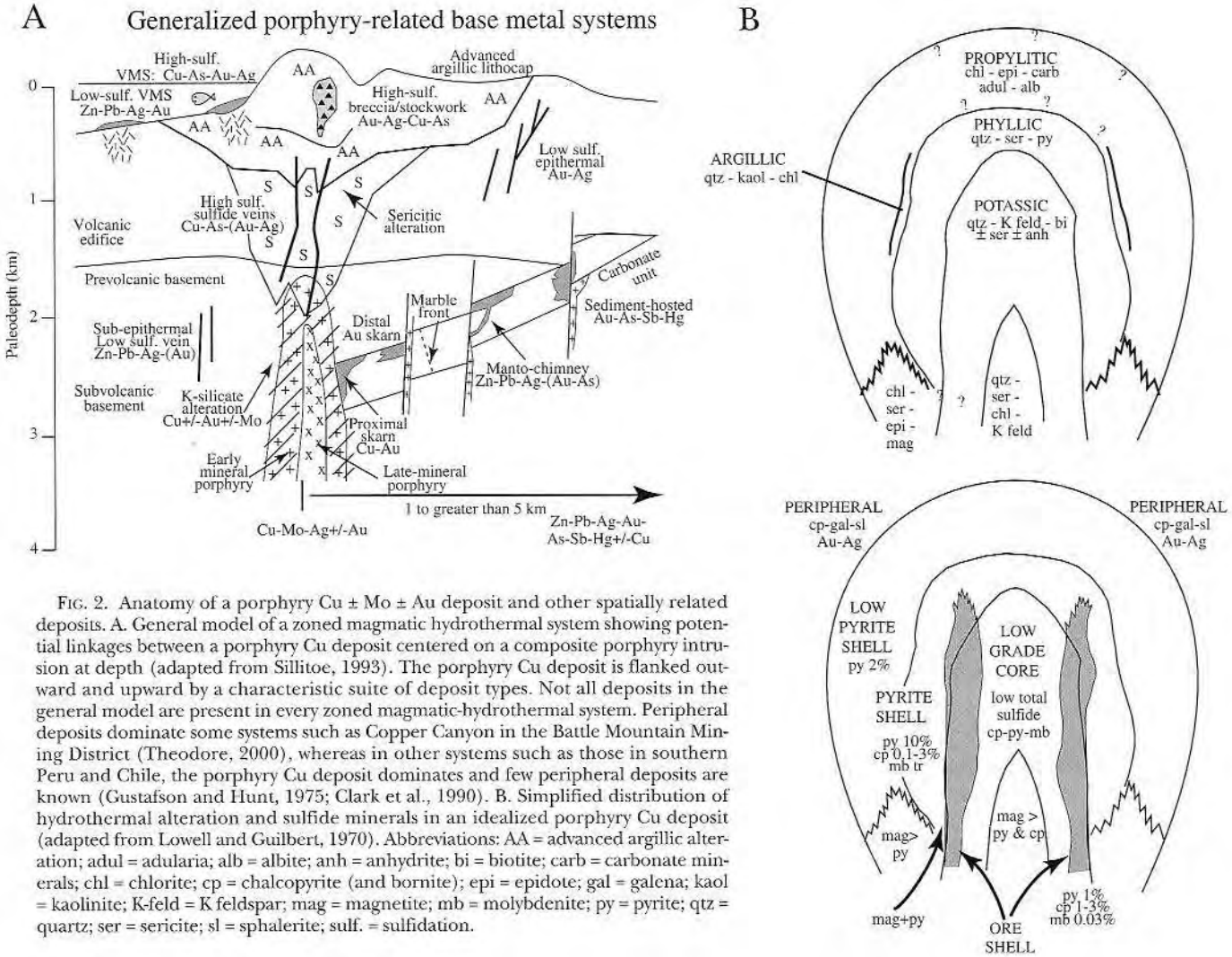

FIG. 2. Anatomy of a porphyry Cu ± Mo ± Au deposit and other spatially related deposits. A. General model of a zoned magmatic hydrothermal system showing potential linkages between a porphyry Cu deposit centered on a composite porphyry intrusion at depth (adapted from Sillitoe, 1993). The porphyry Cu deposit is flanked outward and upward by a characteristic suite of deposit types. Not all deposits in the general model are present in every zoned magmatic-hydrothermal system. Peripheral deposits dominate some systems such as Copper Canyon in the Battle Mountain Mining District (Theodore, 2000), whereas in other systems such as those in southern Peru and Chile, the porphyry Cu deposit dominates and few peripheral deposits are known (Gustafson and Hunt, 1975; Clark et al., 1990). B. Simplified distribution of hydrothermal alteration and sulfide minerals in an idealized porphyry Cu deposit (adapted from Lowell and Guilbert, 1970). Abbreviations: AA = advanced argillic alteration; adul = adularia; alb = albite; anh = anhydrite; bi = biotite; carb = carbonate minerals; chl = chlorite; cp = chalcopyrite (and bornite); epi = epidote; gal = galena; kaol = kaolinite; K-feld = K feldspar; mag = magnetite; mb = molybdenite; py = pyrite; qtz = quartz; ser = sericite; sl = sphalerite; sulf. = sulfidation.

not higher than 5 wt percent, however, because of the common presence of plagioclase phenocrysts (Gill, 1981). Fluid exsolution from the melt upon reaching saturation leads to a sudden rise of the magma solidus temperature, which quenches the matrix and gives rise to the porphyritic texture. This magmatic-derived hydrothermal fluid is, in turn, implicated in the transport and deposition of metals to form porphyry mineral deposits.

The hydrous nature of the melt derives from magmagenesis in an arc environment where water is carried down into the upper mantle by subduction of altered oceanic crust (Fig. 3). Slab dehydration during descent to ~100 km depth metasomatizes the overlying mantle wedge, thereby generating conditions suitable for extraction of hydrous, basaltic magmas by decompression melting (Hamilton, 1981; Arculus, 1994). As these magmas ascend toward the surface, a combination of processes, including depressurization and crystallization of anhydrous mineral phases, leads to a decrease in volatile solubility and an increase in the volatile content of the residual silicate melt, eventually leading to saturation (Burnham, 1967, 1979; Whitney, 1975). At this point, a separate volatile phase is exsolved, which becomes concentrated at the top of the magma chamber (Figs. 4 and 5A). In calc-alkaline to alkaline arc magmas, this water-rich volatile phase will also contain other water-soluble volatile components such as Cl and S species. Depending on the depth (pressure) of emplacement, initial volatile exsolution yields either (1) a high-pressure, homogeneous, supercritical, aqueous fluid with salinity in the range 2 to 10 wt percent NaCl equiv, or (2) at shallower depths (above approximately 5 km) two immiscible fluids, consisting of a low-salinity vapor phase and a high-salinity brine (Burnham, 1979; Cline and Bodnar, 1991; Shinohara, 1994; Gustafson and Quiroga, 1995; Hedenquist and Richards, 1998; Hedenquist et al., 1998). These fluid phases can transport base and precious metals as chloride and possibly bisulfide complexes (Candela and Holland, 1984, 1986; Williams et al., 1995; Gammons and Williams-Jones, 1997; Loucks and Mavrogenes, 1999).

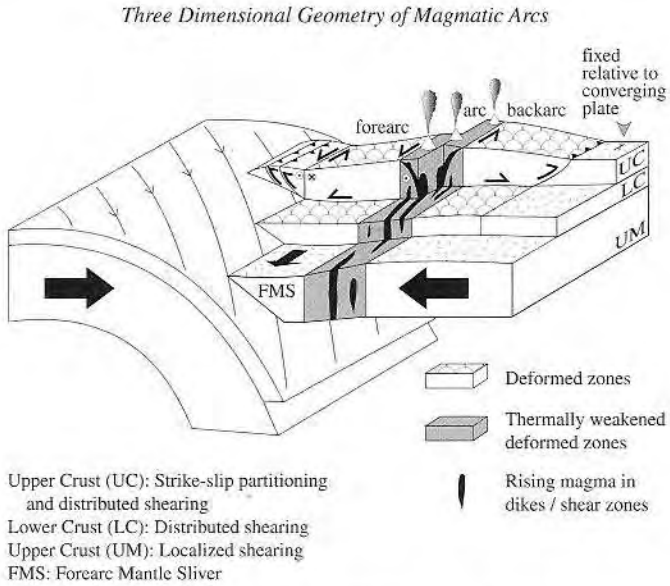

FIG. 3. Geometry of a magmatic arc formed at an obliquely convergent plate margin. Hydrous magmas are generated in the mantle wedge and intrude the lithosphere along broad deformation zones undergoing varying amounts of strike-slip motion depending upon the degree of obliquity along the convergent margin. Modified from de Saint Blanquat et al. (1998).

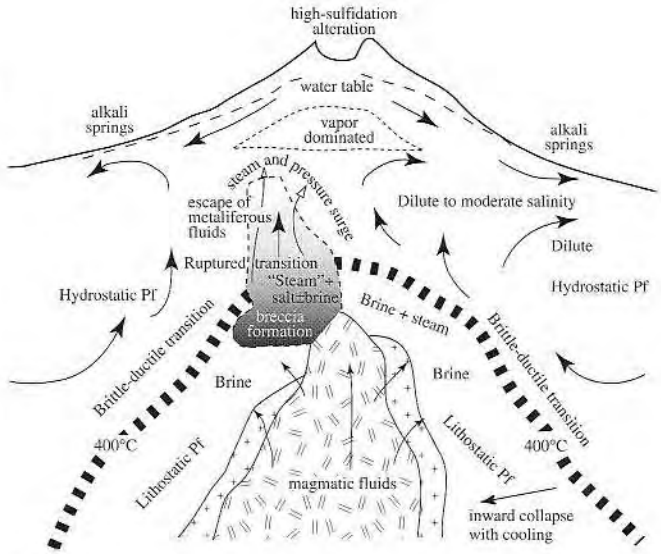

FIG. 4. Fluid circulation patterns in a porphyry copper deposit forming in the cupola of a stock at depths in the range of 1 to 3 km. Exsolution and trapping of magmatic fluids in a zone of lithostatic fluid pressure (P_l) beneath an isotherm of ~370° to 400°C, which marks the brittle–ductile transition. Within this zone, brine and "steam" interact with the country rocks forming the potassic alteration stage. Above the transition zone, circulating fluids under hydrostatic pressure alter the country rocks to propylitic mineral assemblages. In an unperturbed and cooling porphyry Cu deposit, the brittle–ductile transition gradually collapses onto the early potassic alteration zone, leading to phyllic overprinting and the influx of meteoric fluids into the system. If the brittle–ductile transition is suddenly perturbed or breached because of tectonism or later magmatism, the fluid lid ruptures, leading to sudden transition from lithostatic to hydrostatic fluid pressures (see left side of stock on diagram). The fluid-saturated carapace undergoes sudden phase separation leading to "steam" plus brine plus crystals, with the potential expulsion of metal-bearing fluid upward into the epithermal environment. Within the porphyry Cu environment, these explosive pressure transitions trigger the formation of hydrothermal breccias. Adapted from Fournier, 1999.

Exsolution of a volatile phase from magma necessarily involves a large positive change in volume (Phillips, 1973; Burnham, 1979; Burnham and Ohmoto, 1980). Depending on the stress regime and the structure and permeability of the confining host rocks, this volume change may be accommodated by pervasive brecciation of the volatile-saturated carapace of the intrusion, or direct escape toward the surface along fault zones or breccia pipes (Fig. 4). Both mechanisms involve creation of fracture permeability and porosity into which ore minerals may subsequently be precipitated, thereby leading to formation of porphyry deposits.

Hydrothermal fluid exsolved at magmatic temperatures is initially in equilibrium with the coexisting magma and its mineral phases. High-temperature alteration assemblages (Fig. 2B), therefore, consist of minerals such as quartz, K feldspar, biotite, anhydrite, and magnetite (the "potassic" alteration assemblage; Lowell and Guilbert, 1970; Gustafson and Hunt, 1975). Base metal sulfide minerals, typically combinations of chalcopyrite, bornite, and molybdenite with pyrite (Fig. 2B), are deposited with these alteration minerals as disseminations and in thin, commonly wavy or ductilely deformed, wispy veinlets that lack parallel vein walls. These early veins were termed "A veins" by Gustafson and Hunt (1975), based on their study of the El Salvador porphyry deposit in Chile. These veins are usually cut by intramineral porphyry stocks where such stocks are present. Marginally later "B veins" at El Salvador are quartz-rich with more abundant molybdenite and chalcopyrite, and straighter vein walls; these veins cut all but post-mineral stocks. The straighter vein walls and mineral textures within these veins indicate formation at lower temperatures than the A veins.

As the high-temperature fluid migrates away from the parent magma, it rises owing to buoyancy and begins to cool. Two processes may happen, as follows:

1. If the fluid was exsolved originally as a homogeneous supercritical phase, then upon cooling it will likely intersect its solvus and separate into liquid and vapor phases. This phase separation event involves a further increase in volume leading to additional fracture permeability and brecciation, most commonly in the fluid-soaked carapace above or at the top of the porphyry stock. Phase separation may also occur during sudden pressure drops such as massive hydrofracturing, sudden removal of lithostatic load, or rupturing of the brittle-ductile transition zone separating the deeper lithostatic from the overlying hydrostatic environment (Fig. 4; Fournier, 1999).

2. Sulfur originally dissolved in the high-temperature fluid as SO_2 will react with water and disproportionate to form H_2S and sulfuric acid (Burnham, 1979).

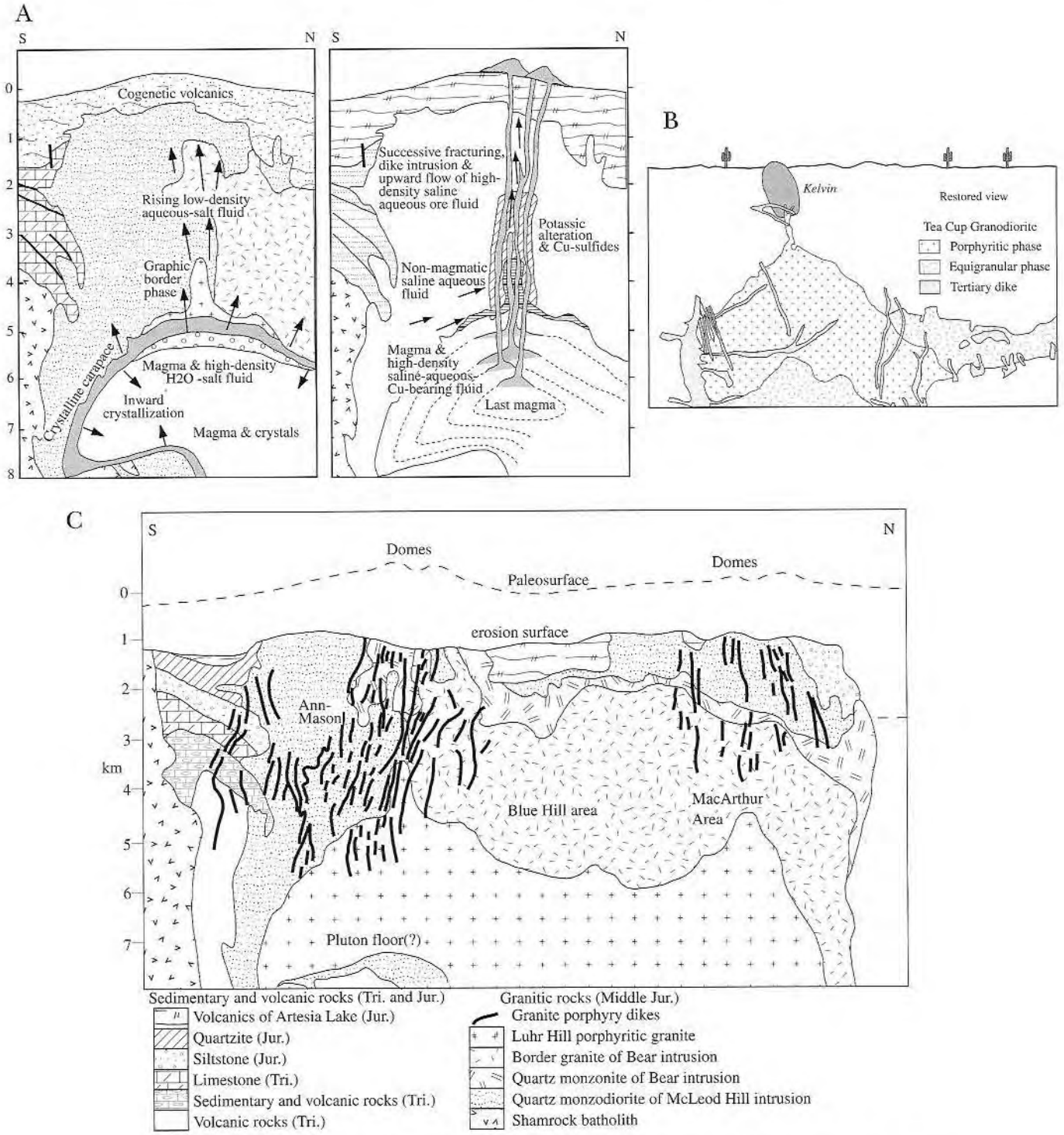

FIG. 5. Relationship between porphyry Cu deposits, porphyry stocks, and dikes to underlying parental magma chambers, which commonly lie at depths of 6 km or greater. A. Evolutionary model for the exsolution of high-density saline aqueous fluid from tabular plutons and the upward escape of this fluid into the roof zone where a porphyry Cu deposit forms (Dilles, 1987). The saline aqueous fluid trapped along the roof of a cooling batholith is channeled upward along dikes or stocks that emanate from the inwardly crystallizing magma. B. The Kelvin porphyry Cu prospect in Arizona is associated with a small porphyry stock that emanates from the apical part of the Tea Cup Granodiorite (modified from Cornwall and Krieger, 1975; Creasey, 1984). C. The Ann-Mason porphyry Cu deposit in the Yerington batholith is associated with dikes emanating from successively deeper parts of the Luhr Hill porphyritic granite (Dilles, 1987). Jur. = Jurassic; Tri. = Triassic.

These processes generate a new set of veins and breccias characterized by gangue and alteration minerals stable at lower temperature and pH. In particular, an increase in activity of sulfate in the fluid owing to SO_2 disproportionation will result in voluminous precipitation of anhydrite and further release of more acid in the form of HCl (Burnham, 1979). In feldspar-rich host rocks, the characteristic alteration assemblage from these acidic fluids is fine-grained muscovite (sericite) with quartz ("phyllic" alteration assemblage, Fig. 2B; Lowell and Guilbert, 1970). In addition, sulfide minerals are precipitated in response to the increase in H_2S activity. Associated D veins (Gustafson and Hunt, 1975) are coarse-grained and sharply defined with strong alteration haloes, and consist of sulfide minerals (abundant pyrite, with lesser Cu Fe-sulfide minerals, sulfosalt minerals, sphalerite, and galena), anhydrite, and minor quartz. Typically, the highest grades of hypogene Cu ore are present close to the boundary between the potassic and phyllic alteration zones, again reflecting the increased activity of H_2S in the fluid as a result of SO_2 disproportionation. The phyllic alteration zone is commonly observed as a shell surrounding the potassic core of the porphyry system (Lowell and Guilbert, 1970), and may have a diameter of several hundred meters to a few kilometers.

Separation of a vapor phase may lead to the formation of a low-density vapor plume rich in acidic volatiles such as SO_2, HCl, HF, and CO_2 (Fig. 4). Cooling and condensation of these vapors at shallow depths causes intense acid leaching ("advanced argillic" alteration; Fig. 4), characterized by total destruction of primary minerals to leave an insoluble residue of quartz, clays (kaolinite or pyrophyllite), aluminum hydroxides (diaspore), and alunite. Such alteration zones, being very porous and permeable, may serve as hosts to high-sulfidation styles of epithermal mineralization (Hedenquist et al., 1993; Hedenquist, 1995).

Emplacement of a hot body of magma into the shallow crust typically involves interaction with local ground waters (Fig. 4). The degree of that interaction depends on crustal permeability, but in many cases hydrothermal convection cells driven by heat from the intrusion can be shown to have operated over radii of many kilometers (Taylor, 1974; Titley et al., 1986). The fluids involved are typically dilute ground waters, and the alteration arising from their flow is characterized by hydration, producing minerals such as chlorite and epidote, commonly with carbonate minerals ("propylitic" alteration; Fig. 2B).

Many variations upon the general porphyry model have been recognized throughout the world, particularly in terms of the sequence of sulfide minerals deposited in veins. For example, at Mineral Hill, Arizona, molybdenite is the oldest sulfide mineral recognized and much of the copper sulfide mineralization is paragenetically younger, generally associated with D-type veins (Wilkinson et al., 1982). Nevertheless, the general sequence of alteration and sulfide mineral deposition described above characterizes the majority of porphyry Cu deposits, regardless of their size.

From a structural standpoint, three processes contribute to fracturing during porphyry Cu formation, and each process enhances permeability in the form of fracture or breccia networks through which hydrothermal fluids may flow and precipitate minerals (Figs. 5–7): (1) intrusion of pre-, syn-, and post-mineral porphyry stocks or dikes, which may be intruded to near-surface depths (1–3 km; Koide and Bhattachariji, 1975); (2) exsolution of hydrothermal fluid from a cooling parental magma chamber, with upward focus into the solidified or partially solidified stocks and warm wall and roof rocks (Fig. 5A); and (3) phase separation of the hydrothermal fluid. Each event involves volume expansion and fracturing wherever the fluid pressure exceeds the ambient minimum effective stress ($\sigma_3' = \sigma_3 - P_f$, where P_f is the fluid pressure plus the tensile strength of the rock; Jaeger and Cook, 1979; Sibson, 2000). Multiple or recurrent pulses of hydrothermal activity, which are common in porphyry systems, lead to repetition of fracturing and superimposition of veins and alteration, leading to increased complexity.

Vein-dominated versus breccia-dominated porphyry Cu deposits

Porphyry Cu deposits developed within noncalcareous country rocks have a considerable range of morphology and show variations between vein-dominated and breccia-dominated systems with considerable overlap between the two. Vein-dominated porphyry Cu deposits represent a relatively simple environment: veining events are consistent with emplacement of a porphyry stock, exsolution of a magmatic-derived hydrothermal fluid with local influx of external fluids, and cooling of an evolving magmatic-driven hydrothermal system. These deposits are the most common variety of porphyry Cu deposits, with numerous well-documented examples (see volumes edited by Titley and Hicks, 1966; Sutherland Brown, 1976; Titley, 1982; Pierce and Bolm, 1995; Schroeter, 1995).

The other end-member is breccia-dominated (Sillitoe, 1985). Breccias in the porphyry Cu environments have a range of morphology and genesis, being formed from igneous, magmatic-hydrothermal, and phreatomagmatic processes. The most common types of breccia are pre- to syn-mineral bodies that are irregular to pipe-like in shape, and are characterized by angular clasts, sharp or gradational contacts against enclosing rocks, and an intimate association with porphyry stocks or dikes. The breccia matrix consists of igneous rocks or hydrothermal silicate and sulfide minerals (Richard and Courtright, 1958; Camus, 1975; Zweng and Clark, 1995). Post-mineral breccias commonly but not always have a milled rock-flour matrix and rounded clasts; they are referred to as pebble breccias. These breccias essentially excise parts of a prior hydrothermal system (Richard and Courtright, 1958), although locally, such as at Toquepala, the pebble breccias are intramineral and contain chalcopyrite in the matrix. Zweng and Clark (1995) proposed that breccia-dominated porphyry Cu deposits form at shallow levels based upon fluid inclusion data from Toquepala. They suggested that vein-dominated deposits would more commonly form at slightly greater depths. Implied in their model is a primary influence of the lithostatic load of the overlying column of rock on breccia formation (see discussion below).

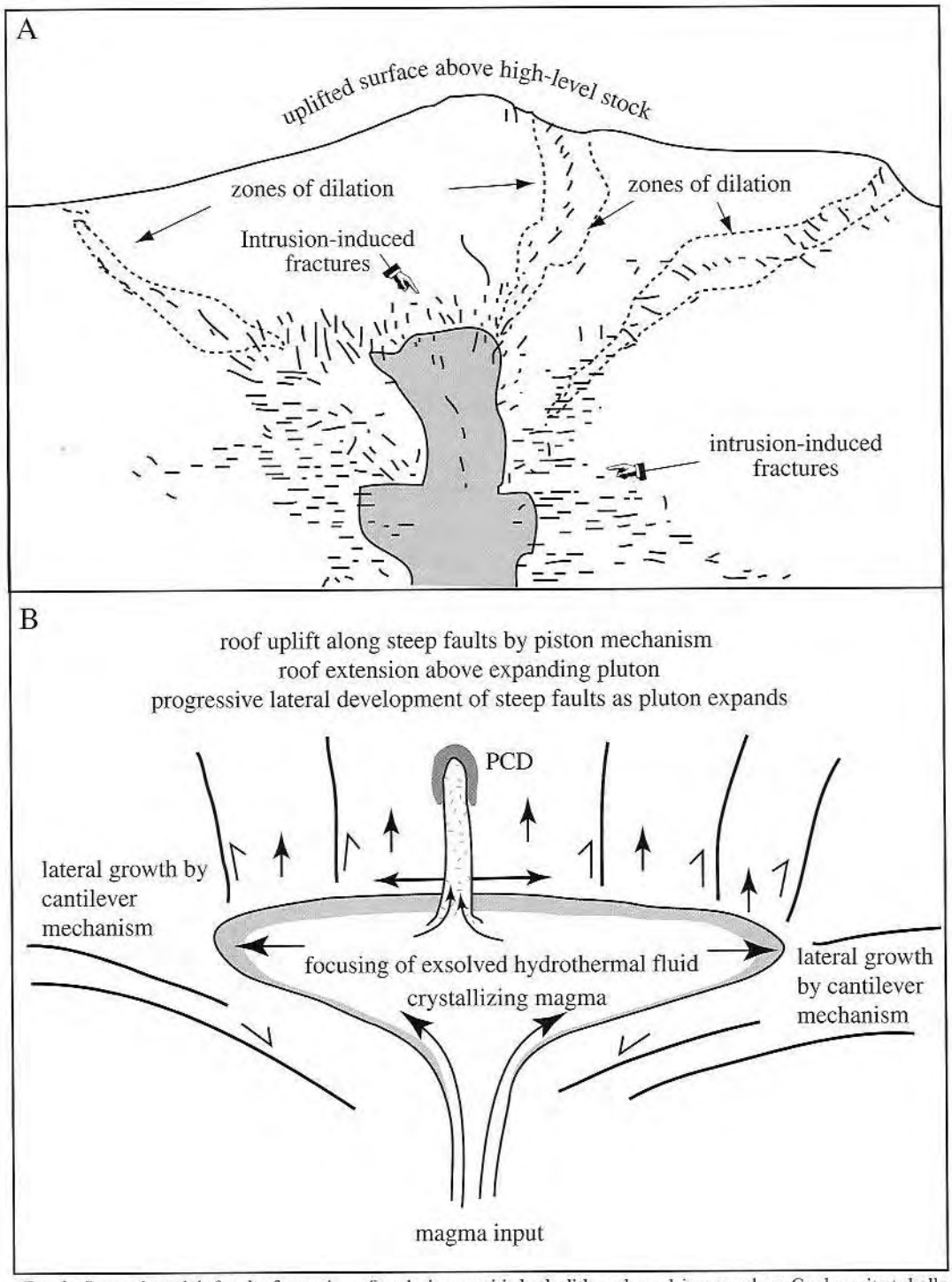

FIG. 6. General models for the formation of a tabular granitic batholith and overlying porphyry Cu deposit at shallow levels in the crust. A. Computer simulation of growing tabular batholith fed from below (Australian Geodynamics Research Group, 1997). Subhorizontal fracture and fault systems propagate outward from the expanding tips of the tabular pluton. Rising above the growing batholith are stocks extending to shallow levels in the crust. The surface of the Earth is displaced upward by the shallow stock, whose emplacement develops concentric fracture and fault systems as well as intense, steep fracture networks in the overlying volcanic carapace. B. A tabular pluton expands laterally and grows dominantly by a cantilever mechanism. Lifting of the pluton roof along steep dipping normal faults accommodates the vertical expansion of the stock. Upward expansion of the pluton may be assisted by stoping of roof rocks and potentially by ballooning. Roof uplift is likely accompanied by extension, which leads to weak zones into which apophyses or stocks may intrude. These apophyses focus the ascent of exsolved magmatic-derived hydrothermal fluids, and subsequent ore formation (see Fig. 5 for examples). PCD = porphyry Cu ± Mo ± Au deposit.

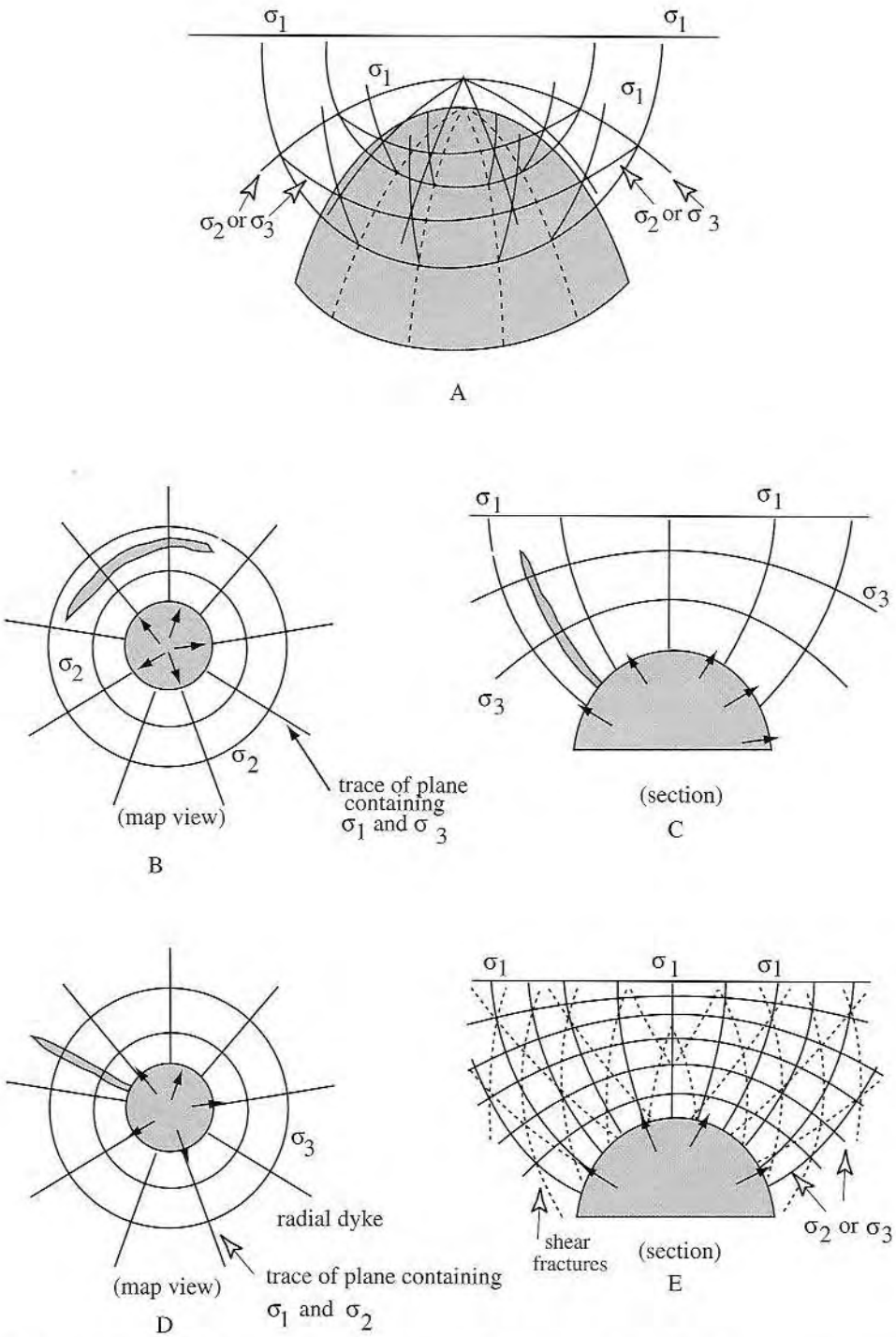

FIG. 7. Simplified predicted stress trajectories above a rising, shallow-level stock where the stress state approaches uniaxial extension above a cylindrical stock. In isotropic rocks, formation of fractures, veins, and dikes would be guided by the stress trajectories. In anisotropic rocks, older structural fabrics, such as faults, bedding, and folds, will strongly influence the stress distribution, and favorably oriented planes of weakness will be preferentially opened. In all cases, the maximum principal stress (σ_1) is oriented vertically. Dikes and veins develop normal to the minimum principal stress (σ_3) and in the plane containing σ_1 and the intermediate principal stresses (σ_2). Other planes of weakness can form conjugate systems of shear fractures, distributed symmetrically distributed about σ_1. A. Stress trajectories generated above a domical stock. B, C. Map and cross-sectional views of orientation of stress for generation of cone sheets or concentric fractures. D. Map view of stress trajectories for generation of radial dikes and fractures. E. Cross-sectional view of the orientation of shear fractures generated by a stress system above a domical stock. Modified from Park (1983).

Convergent Plate Margin Magmatism

Porphyry Cu deposits are the end result of a complex sequence of events initiated by magmagenesis at convergent plate margins, and it is, therefore, useful to review the process of magma ascent and emplacement in the lithosphere. Convergent margin magmatism is linked to subduction of an oceanic plate beneath an overriding continental or oceanic plate, upon which the arc is constructed (Fig. 3). In this environment, plate motion vectors are commonly oblique at varying angles to the plate margin, and stress transmitted to the overriding plate, and the resulting deformation, reflect the degree of obliquity. Strain is partitioned into arc-normal (convergent or divergent) and arc-parallel (strike-slip) components. A wide range of structural styles can, thus, be expected in the upper plate depending on the degree of convergence obliquity, convergence rate, and the angle of subduction (Fitch, 1972; Dewey, 1980; Jarrard, 1986). Of additional importance is whether trench retreat or advance occurs at the plate margin (Hamilton, 1988; Royden, 1993). In the former situation (trench retreat), the overall tectonic setting will be extensional, or transtensional where a significant degree of net convergence obliquity is present. In the latter case (trench advance), contraction or transpression dominates. Regardless of convergence geometry, some relationship between strike-slip tectonics and magmatism, including porphyry magmatism, is present in convergent margin arcs, although the relative importance of strike-slip motion will vary (e.g., Glazner, 1991; de Saint Blanquat et al., 1998). The ubiquitous association of plutonic and volcanic complexes and major fault zones in arcs implies that rising magma concentrates in zones of deformation. A corollary to this observation is either that deformation is crucial to the emplacement of magma, or that magma assists the deformation processes which, in turn, helps its ascent to higher crustal levels (McCaffrey, 1992; de Saint Blanquat et al., 1998). Paterson and Schmidt (1999; see also Schmidt and Paterson, 2000) have argued that many plutonic complexes are not necessarily emplaced along faults but come to rest in the less deformed rocks between major faults. However, their analysis applies to contractional orogenic belts characterized by thrust faults, rather than transpressional or transtensional strike-slip zones in volcanic arcs (Richards, 2000a).

Exactly how magma moves through the crust is poorly understood. Nevertheless, there is general agreement that the following factors are important (Hogan and Guilbert, 1995; Hutton, 1997): (1) magmas, being fluid, will always move down a hydrodynamic gradient toward regions of low pressure; (2) in the absence of significant magmatic overpressure, the density contrast between a granitic melt and the surrounding rocks will drive the melt upwards to an "equilibrium" level of neutral buoyancy in the crust; (3) increasing vapor pressure during late stages of crystallization generally increases magma pressures; (4) the lithostatic load of the overlying crust may act as a pressure seal hindering the rise of magma; and (5) temperature-dependent viscous resistance to magma flow inhibits ascent. Magmatic overpressure may drive magmas beyond their level of neutral buoyancy, commonly resulting in violent eruptions. Overpressured magmas may intrude by forcing open dikes or pipes, or by penetrating old faults or fractures (Bussell et al., 1976; Delaney et al., 1986; Ida, 1999). The prevailing crustal stress field controls the orientations of dikes, favoring intrusion along older faults oriented at high angles to the least principal stress (σ_3). Alternatively, tectonic stress and resultant strain may episodically create reduced pressure or tensional zones in the crust, such as pull-apart zones at fault intersections or jogs, which may focus magma ascent (Hutton, 1988; Tikoff and Teyssier, 1992; Grocott et al., 1994; Richards et al., 2001). Thus, the varied styles and mechanisms of magma emplacement in the crust involve an important dynamic interplay between tectonic and magmatic processes. The final three-dimensional form of the stock is controlled by strain rate, magma supply, country rock versus magma rheologic differences, and tectonic environment (Castro and Fernandez, 1998; Corriveau et al., 1998; Hogan et al., 1998; Paterson and Miller, 1998a, b; Yoshinobu et al., 1998; Wilson and Grocott, 1999).

Growth of a magma chamber

Magmatism at the Earth's surface or in near-surface environments is a manifestation of the relative effects of tectonic stress and magmatic overpressure. Whereas tectonic or vertical stress acting on magma decreases with decreasing depth, the effect of magmatic overpressure should remain essentially unchanged because of hydraulic connectivity to depth. Hence, final emplacement of magma at shallow crustal levels will, in large part but not entirely, reflect magma dynamics or magma overpressure (Takada, 1994; Hutton, 1997; de Saint Blanquat et al., 1998; Ida, 1999). Coherent magma chambers recharge and grow through the aggregation of small pulses of magma, perhaps through the intrusion of multiple dikes or lobate bodies (Petford et al., 1993, 1994). Buoyant diapiric rise of magma or lateral translation by pushing aside of wall rocks helps create room for an expanding pluton (Tikoff et al., 1999). de Saint Blanquat et al. (1998) proposed that magma ascent into the crust stops where the magma becomes capable of deforming or displacing its enclosing wall rocks, thereby creating space for a growing or ballooning magma chamber. Evidence for varying amounts of wall-rock strain around plutons at all depths of emplacement lends support to this model.

Granitoids emplaced at high levels in the crust tend to have tabular shapes with flat tops and bottoms and steep sides, and lateral dimensions that are much larger than their vertical dimensions (Fig. 5B, C; Hamilton and Myers, 1967; Dilles, 1987; McCaffrey and Petford, 1997). Such magma chambers grow above narrow vertical feeder zones or dikes and spread laterally as sill-like intrusions that inflate by floor displacement or by roof-lifting (Fig. 6; Johnson and Pollard, 1973; Pollard and Johnson, 1973; Corry, 1988). Floor depression leading to lopolithic complexes seems to be the most common form of granitoid emplacement except at high-crustal levels (<3 km depth),

where roof lifting is important and laccoliths may form (Corry, 1988). Mechanisms of lateral growth of a magma chamber are poorly understood, but they involve relative downward transport of the country rocks as the magma expands laterally. Mechanisms include cauldron subsidence (Myers, 1975), stoping (Paterson et al., 1996), folding (Hamilton and Myers, 1967), and downward transport on fringing brittle or ductile shear zones along the margins of the pluton (Paterson et al., 1996; Cruden, 1998). Lateral growth of a tabular body is aided by faults and fracture networks that extend laterally from the edges of the plutons (Fig. 6A; Australian Geodynamics Research Group, 1997). Such subhorizontal fracture systems could be ductile shear zones or brittle fault zones depending on the pressure-temperature conditions and, hence, depth. Emplacement of additional dikes, sills, or veins along the leading fracture tip would facilitate the lateral growth of the pluton.

Cruden (1998) evaluated various mechanisms for the thickening of tabular granitoid bodies in the upper crust. He proposed that tabular plutons result from piston or cantilever mechanisms, or a combination of the two (Fig. 6B). Both mechanisms result in plutons that have flat roofs. Steep faults developed in the roof or floor of the growing magma chamber would also assist in vertical chamber growth (Fig. 6B). If chamber growth is accompanied by an upward expansion somewhat akin to ballooning, or by a piston mechanism on inward dipping faults, extension of the roof rocks would be expected. Lateral growth in such circumstances would result in the outward development of a sequence of steep faults that presumably nucleated at the tip of the expanding chamber. Alternatively, the steep faults may nucleate along older planes of weakness such as intrusive contacts or older faults.

Plutons emplaced by a piston mechanism will have flat floors and steep sides that correspond to brittle or ductile faults or to feeder dikes. The bell-jar stocks in the Peruvian Coastal batholith are excellent examples of this mechanism (Myers, 1975). Emplacement by a cantilever mechanism results in displacement of the floor of the chamber downward, along steeply or shallowly dipping shear zones, leading to an inward sloping floor to the magma chamber. Combinations of the two mechanisms are likely, and should be expected for chambers emplaced in the brittle upper crust at depths near or below ~3 km. Such depths are believed to be typical for the parental magma chambers that source the shallower-level porphyry-related stocks and dikes (Fig. 5B, C). Steep pluton margins are commonly controlled by regional fracture sets, active faults, or older fault systems (Bussell, 1976; Dehls et al., 1998). Growth and replenishment of the chamber is via feeder dikes (Petford, 1996), which also act to recharge the magma system with fresh magma. Each influx of new magma could be of different composition, and may carry additional volatiles (Folch and Martí, 1998). This process leads to evolutionary changes in chemistry of the magma body (Matthews et al., 1995; Keith et al., 1998) that perhaps contribute to porphyry Cu formation (Clark, 1993).

Volcanoes and Shallow-Level Intrusive Complexes: The Environment of Porphyry Cu Deposits

In the volcanic and near-surface subvolcanic environment (<3 km depth) where porphyry Cu deposits form, the free boundary provided by the Earth's surface is a critical constraint on magma emplacement, growth, and eruption (Nakamura, 1977; Takada, 1994). The outward and upward pressure exerted by a rising magma chamber lifts and extends the roof and adjacent walls of the stock, leading to characteristic patterns of brittle deformation in the overlying carapace, inflation of wall rocks, and local emanation of radial dikes or cone sheets from the central stock (Figs. 6 and 7; Koide and Bhattachariji, 1975; Park, 1983). Fracturing is most intense across the roof of the stock. Most of the roof uplift is accomplished along faults that root in the edges of the stock, and outwardly progressing roof fault systems are expected above a series of composite subvolcanic stocks. The faults and fracture networks provide channelways for intrusion of later stocks and dikes, as well as hydrothermal fluids (Fig. 5A).

Within this near-surface environment, several factors govern growth and final form of a magma chamber and the overlying volcanic edifice if present (Nakamura, 1977), and, by extension, the characteristics of associated magmatic-hydrothermal mineral deposits. Volcanoes and subvolcanic stocks and dikes are commonly emplaced along fault zones. These fault zones may have formed contemporaneously with magmatism, or may have been older structures reactived during magmatism (Bussell, 1976; Shaw, 1980). Whether or not these fault zones were active during magmatism depends upon the regional tectonic stress field, strain developed as a consequence of this stress, and magmatic pressure. The tectonic stress regime controls which fault orientations might be dilatant, as well as whether or not original formation of the structures was contemporaneous with magmatism. Favorably oriented faults provide dilational channels permitting emplacement of magma and fluids (Sibson, 1985, 2000). Takada (1994) proposed that the final form of volcanoes and their shallow-level intrusive complexes reflects the degree of magma supply coupled with the differential horizontal stress or strain rate (Fig. 8). Areas of small differential horizontal stress and high magma supply rate are conducive to the formation of large plutonic complexes and overlying polygenetic stratovolcanoes. In contrast, regions with a similar magma supply but large differential horizontal stress (and correspondingly higher strain rate) are characterized by numerous largely monogenetic volcanoes and small, more dispersed, subvolcanic plutonic complexes.

Subvolcanic plutons range from circular to elliptical in map view. The orientation of the stock's ellipticity and any associated dikes will be at high angles to the orientation of the finite extension direction and minimum principal stress, whether that is a regional tectonic-induced or local pluton-induced stress. The exact angle depends greatly on the degree of structural anisotropy in the immediate area of the complex. Where the plutonic complex is emplaced in

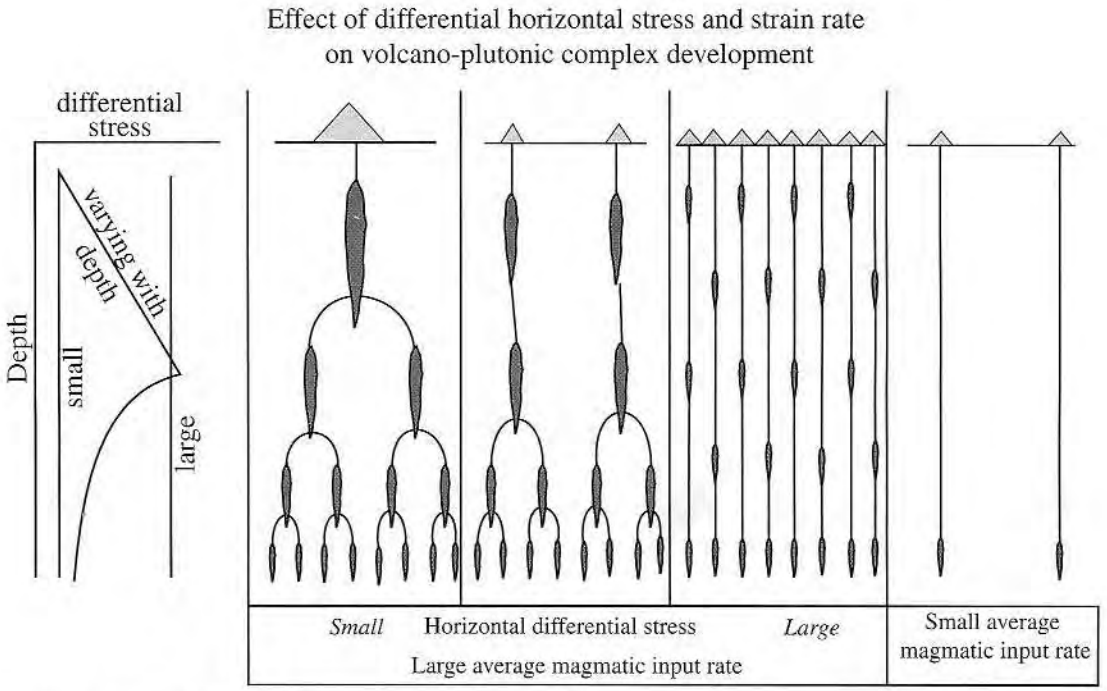

FIG. 8. Effect of differential horizontal stress and magma supply on distribution of volcanoes and subvolcanic intrusions; modified from Takada (1994). A large, near-surface, degassing magma chamber is required for the formation of a porphyry Cu deposit. Formation of such magma chambers in the upper crust is seemingly favored by large to probably intermediate magma input rates, coupled with small to intermediate differential horizontal stress and strain rates.

isotropic or little-faulted rocks, the long axis of the complex may approach orthogonality to σ_3 depending upon the magnitude and style of deformation. However, where the complex is emplaced into highly faulted crust, its long axis could be at an oblique angle to σ_3, because favorably oriented faults and other structural fabrics will be exploited preferentially. The ellipticity of the stock also gives some indication of the differential horizontal stress and strain rate during magmatism (Nakamura, 1977; Takada, 1994). For example, where radial dikes are well developed, a uniform horizontal stress field during magmatism is indicated (Fig. 9). This situation seems to dominate the near-vent areas of many volcanoes (Nakamura, 1977), and implies that associated subvolcanic stocks will be circular to slightly elliptical in map view. In contrast, where significant differential horizontal stress or high strain rate is established, radial dike swarms characterize only the core of the magmatic system where magmatic pressure exceeds the tectonic stress (Fig. 9). Away from the volcanic core, dikes, peripheral intrusions, satellite volcanoes, and axial grabens develop along fault systems that are aligned at high angles to the minimum principal stress. An important part of this transition from the volcanic core to peripheral environments is the effect of temperature. Cool rocks away from the magmatic center will fracture under the influence of a regional stress field, whereas near the warm core of the magmatic system, interaction between regional and magmatic pressures will control the final form of the volcano, subvolcanic stocks, and, at still greater depths, the parental plutonic body.

Of additional importance to the final form of shallow plutonic complexes is the interplay between the cooling, solidified edges of the chamber and the dynamic, partially molten, and potentially recharged core. The importance of this interplay extends to the country rocks where the mechanical contrast between a cooling pluton and its host rocks governs the spatial development of faults, fractures, and shear zones, which may in turn control the locations of later intrusions and hydrothermal systems (Gow and Ord, 1999). Where a strong rheological contrast is present, such as where plutons intrude sedimentary rocks, faults may form along margins of the central stock, thereby permitting access of late magma and fluid to the country rocks. Where there is a weaker contrast, however, such as where plutons intrude volcanic rocks, deformation is most likely concentrated within the stock, thereby enhancing episodic magmatism and hydrothermal circulation. In addition, the numerical models of Gow and Ord (1999) indicate that conjugate zones of volumetric dilation around the pluton refract at the pluton margins, and do not reach its center. The greatest dilatancy, therefore, occurs at the intersection of these conjugate zones and above the pluton. Gow and Ord (1999) further suggested that if horizontal compressive stress is sufficient, fluid or magma will be drawn up out of the stock and into the dilatant zone in the carapace. External fluids may also be drawn into this zone leading to fluid mixing, a prediction consistent with evidence for influx of external fluids during potassic alteration in some deposits (e.g., Dilles et al., 1995; Bouse et al., 1999).

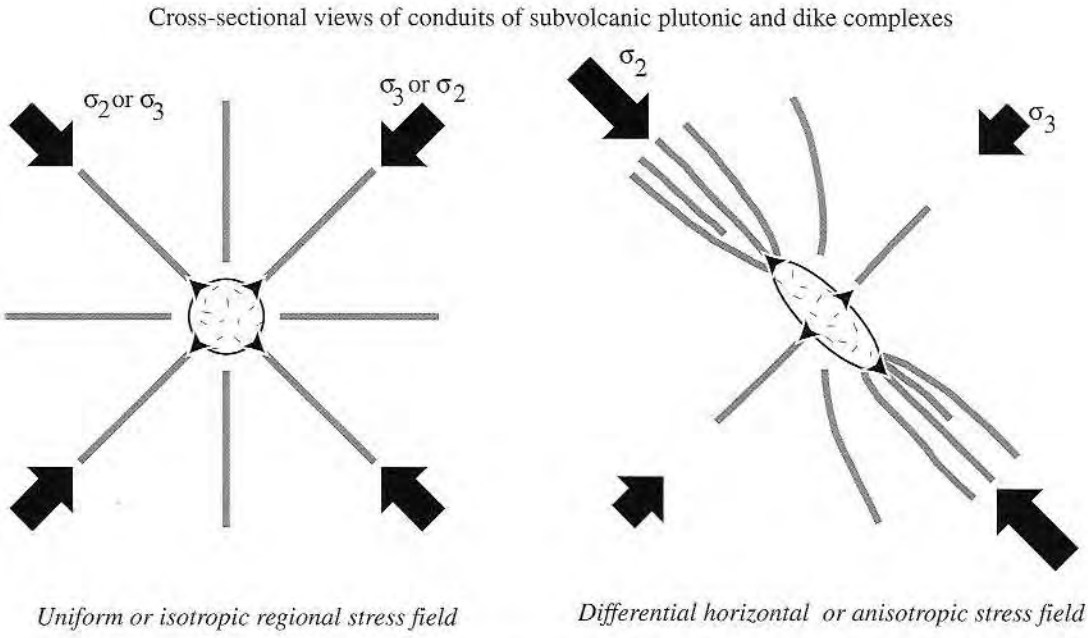

FIG. 9. Radial dike distribution, polygenetic volcano morphology, and underlying map view of plutonic complexes in different idealized stress regimes; modified from Nakamura (1977, and references therein). A. Uniformly distributed radial dikes, subcircular volcanic edifice, and underlying subvolcanic stock(s) form in an isotropic stress field where the horizontal stresses are subequal (e.g., $\sigma_2 \approx \sigma_3$). B. Asymmetric dike distribution and curviplanar dikes developed under an anisotropic stress field where a differential horizontal stress is extant (e.g., $\sigma_2 > \sigma_3$). The volcano and underlying stocks would have strongly elliptical outlines in map view with the long axis at high angles to σ_3.

Where older structural anisotropy is present, faults may be reactivated and may guide magma from the underlying source chamber into the shallow crust or volcanic environment. Reactivation of old structures is a common theme in many porphyry Cu-rich regions such as Arizona (Heidrick and Titley, 1982) and the Central Andes (Salfity, 1985; Richards, 2000b; Richards et al., 2001). On a regional scale, reactivation of old fault and fracture systems is enhanced if they are oriented favorably with respect to the ambient stress field. Furthermore, intersections of such structures may provide focal points for magma emplacement and potential porphyry Cu generation if the conditions are appropriate (Fig. 10; Heidrick and Titley, 1982; Richards, 2000b; Richards et al., 2001). On a local scale, however, structures may be reactivated during magmatism and hydrothermal circulation, even where they are not ideally oriented with respect to the regional stress field, implying that magmatic or hydrothermal fluid pressures locally exceed tectonic stresses.

Examples of misoriented structural fabrics that have been reactivated during intrusion and porphyry Cu formation occur in Arizona (Heidrick and Titley, 1982), the Maricunga belt of Chile (e.g., compare convergence geometry outlined by Pardo-Casas and Molnar, 1987, with vein orientations described by Vila et al., 1991), and at El Salvador, Chile. The alignment of Eocene subvolcanic stocks along northeast-striking Paleocene faults around the El Salvador porphyry provides a particularly well-documented example of this phenomenon (Fig. 11; Gustafson and Hunt, 1975; Cornejo et al., 1997). During the Eocene, this region experienced transpressive deformation, with the maximum compressive stress oriented northwest-southeast, at high angles to northeast-striking Paleocene faults that formed as a result of earlier volcanism and tectonism (Tomlinson et al., 1993; Cornejo et al., 1997). Under the Eocene stress field, northeast-striking faults would not have been under tension and should not, therefore, have localized magma intrusion. Nevertheless, Eocene stocks are aligned parallel to Paleocene faults and fracture systems, beginning with the Cerro Pelado and Old Camp complexes to the northeast, and progressing southwestward to Cerro Indio Muerto (Fig. 11A). A weak porphyry Cu system developed in the older and slightly more silicic Cerro Pelado and Old Camp complexes (Gustafson and Hunt, 1975), but the major porphyry Cu system formed in association with granodioritic stocks underlying Cerro Indio Muerto. The latter stocks are elliptical in plan view with long axes striking northwest (Fig. 11B; Gustafson and Hunt, 1975), at high angles to the northeasterly alignment of stocks at the district scale. Late dacitic (latite of Gustafson and Hunt, 1975) dikes also strike northwest, thereby implying northeasterly extension, consistent with emplacement within the inferred Eocene regional stress field (Cornejo et al., 1997). Thus, although misoriented Paleocene faults broadly localized plutonism on a district scale, the form of these intrusions reflects the prevailing Eocene stress field. Either the horizontal differential stress was not great, or magmatic pressures were sufficiently high such that the tectonic stresses acting to close the north-

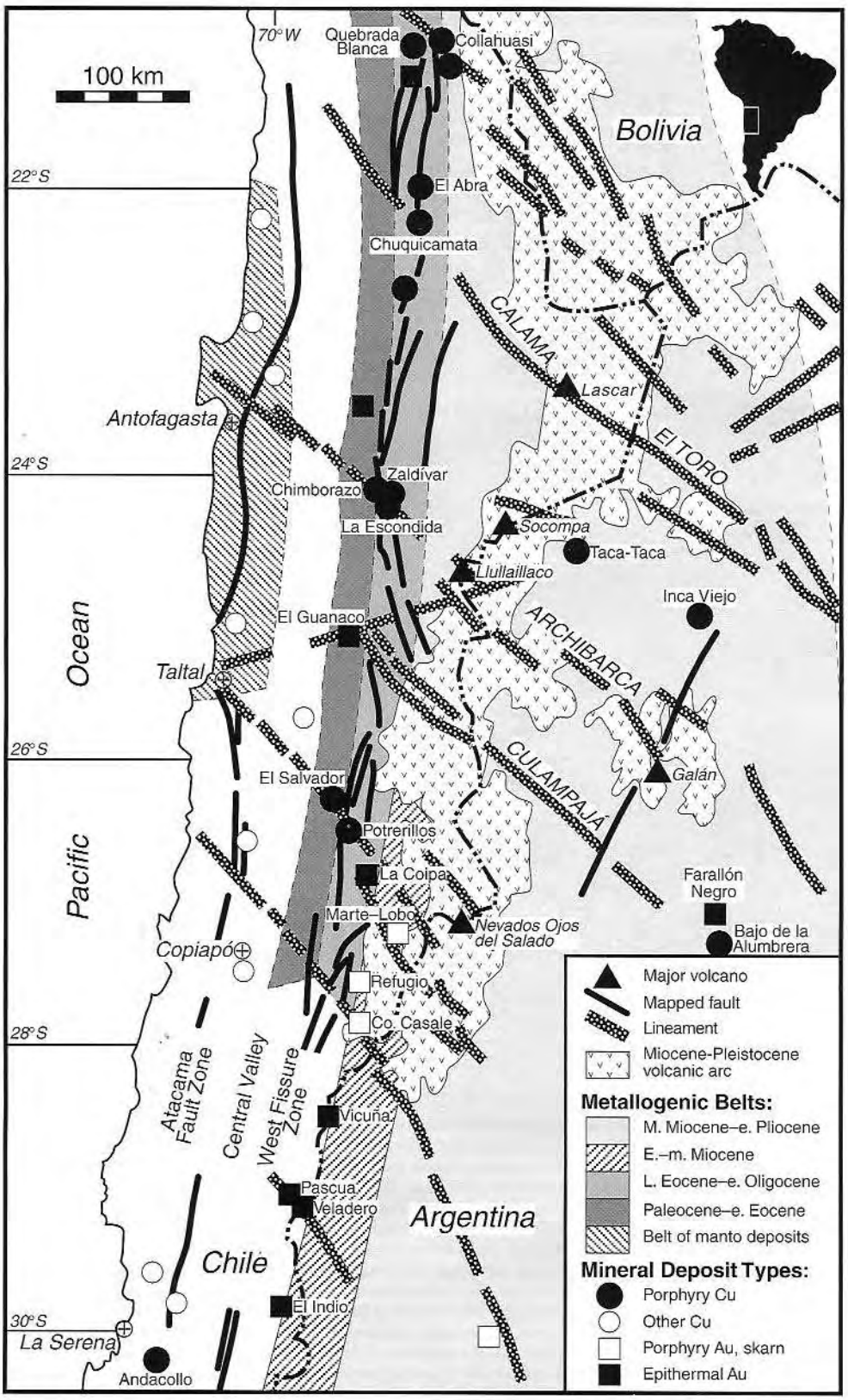

FIG. 10. Geological sketch map of the Central Andes showing the locations of major Cu and Au deposits (from Sillitoe, 1992) and their spatial relationships to the West Fissure Zone and its intersections with cross-orogen lineaments (from Salfity, 1985). Modified from Richards (2000b).

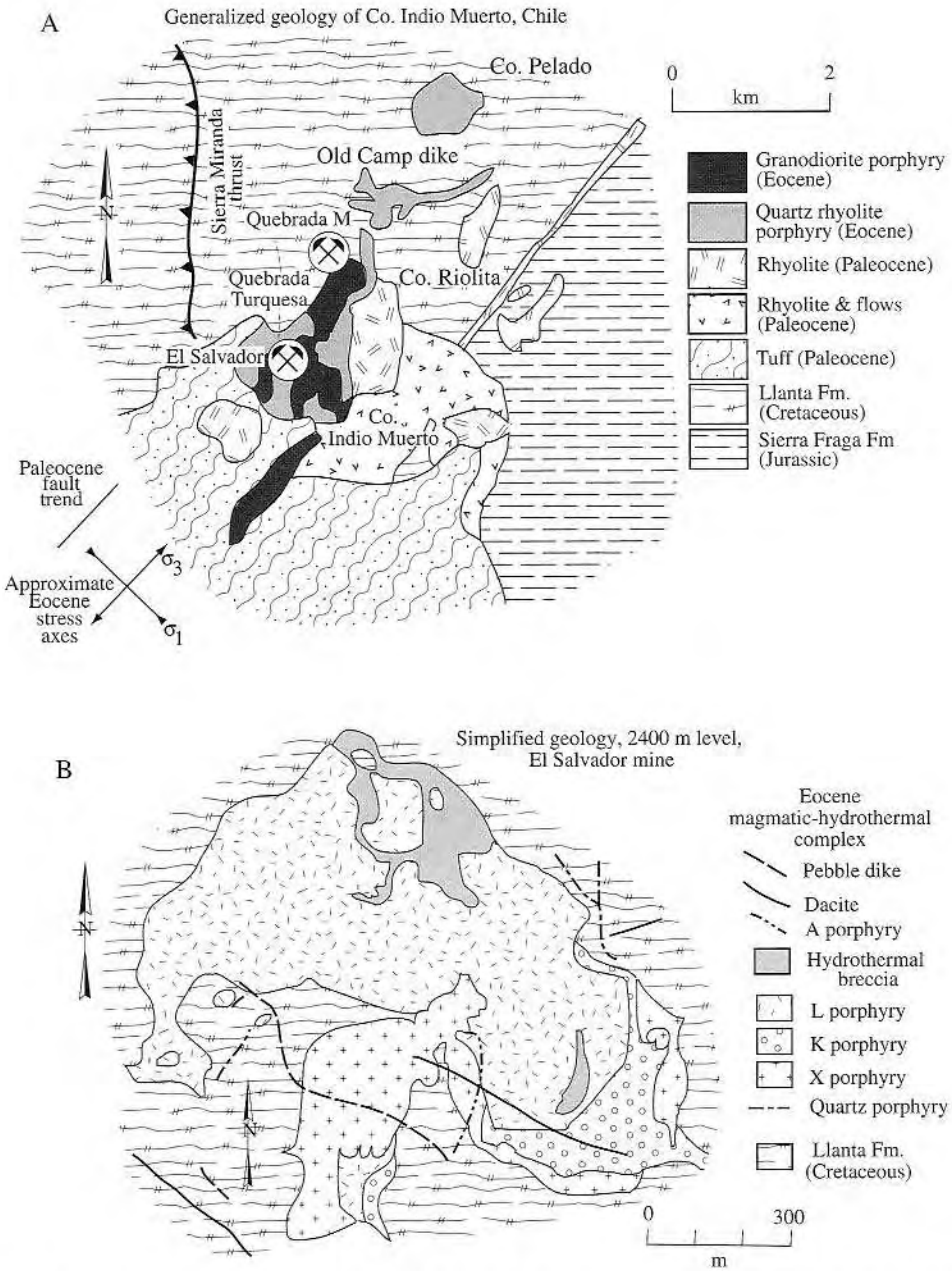

FIG. 11. Geology of Cerro Indio Muerto and El Salvador porphyry copper deposit, northern Chile. A. Cerro Indio Muerto consists of a Paleocene silicic dome complex and tuffs that unconformably overlie deformed Mesozoic volcanic, volcaniclastic, and sedimentary rocks. The complex sits on the edge of the Paleocene El Salvador trap-door caldera. Caldera margin faults and subparallel faults strike northeast. Paleocene rhyolite dikes, such as the linear one immediately east of Cerro Riolita, intruded along northeast-trending faults and fracture systems. The inferred orientation of the Eocene stress field would have favored dilation along northwest-trending structures, but the strong northeasterly alignment of the Eocene intrusions suggests that their emplacement was nevertheless controlled by the older Paleocene structural fabric. Quartz rhyolite or quartz porphyry intrusions are about 43 Ma old, whereas the granodiorite porphyries are 41 to 42 Ma. Modified from Gustafson and Hunt (1975) and Cornejo et al. (1997). B. Simplified geological map of the 2400-m level of the El Salvador mine showing the distribution of porphyry intrusions, listed in order of decreasing age from bottom to top in the explanation. Note at this scale the northwesterly long axis of the plutonic complex, compared with the regional northeast-alignment of plutons; the majority of the complex consists of the late-mineral L porphyry. The oldest stocks, the X and K porphyries, seem to be largely semi-circular intrusions in the lower right-hand portion of the map area. Post mineral dacite dikes and genetically related pebble dikes have prominent northwesterly strikes. The orientation of the dikes and long axis of the stocks is consistent with the Eocene regional stress field shown in A. Modified from unpublished mapping by Anaconda geologists and provided by Guillermo Muellar, Compania del Cobre–Chile (CODELCO) at El Salvador.

east-trending structures were overcome by the force of intrusion. Because no evidence is present for any explosive volcanic activity associated with the Eocene complexes, it is more likely that the differential horizontal stress was low.

Structural Sequence in Porphyry Systems

Early veins in porphyry systems (A-veins of Gustafson and Hunt, 1975) form at near-magmatic temperatures and under lithostatic pressures within a ductile environment, which is surrounded and overlain by a brittle region dominated by hydrostatic pressures (Fig. 4; Gustafson and Hunt, 1975; Muntean, 1998; Fournier, 1999). Depending upon rheologic contrast, the transition between the brittle and ductile zones may lie within temporally associated porphyries or in the country rocks. As the system evolves and cools, later veins at the same depth form under hydrostatic pressures and under brittle conditions. Within the ductile region, fractures may form in response to sudden changes in internal fluid pressure (e.g., owing to chamber recharge by fresh, volatile-rich magma), rapid strain rates that exceed the capacity of the rock to deform plastically, or rupturing of the pressure seal leading to a sudden transition from lithostatic to hydrostatic pressure conditions. In the latter case, the volatile phase is likely to undergo phase separation and a large positive volume change, which may lead to extensive hydrofracturing (crackle brecciation). As temperature decreases, veins become more regular in orientation because they form by brittle failure under the influence of external tectonic stresses (Figs. 7 and 12). Fracture formation and vein filling continues episodically so long as sufficient hydrothermal fluid is channeled into the evolving carapace from the tabular pluton at depth.

Variations in vein and mineralized fracture orientation

Comprehensive vein and mineralized-fracture analyses in porphyry Cu deposits are few. Those that have been undertaken typically reveal a variation in orientation and style with inferred depth of formation of the porphyry Cu deposit that is consistent with a magmatic and structural framework of shallow subvolcanic stocks (Figs. 7 and 12; Heidrick and Titley, 1982). At shallow depths, concentric or quasi-concentric and radial fracture patterns characterize some deposits (Langerfeld, 1964; Baumer and Fraser, 1975; Corn, 1975; Dunn, 1982; Heidrick and Titley, 1982). For example, at the San Juan deposit, Arizona, concentric and radial fractures and veins are centered around one lobe of a stock, but dikes that form an orthogonal pattern are offset from the center of the fracture and vein array (Fig. 12A; Heidrick and Titley, 1982). In the shallow levels of El Salvador, two overlapping radial arrays of post-mineral pebble dikes, veins, hydrothermal breccias, and fractures reflect emplacement of two apophyses or stocks at depth (Fig. 12B; Langerfeld, 1964). It is not clear at what point during the life of the El Salvador porphyry system these fractures formed, although they are now filled by late-D veins and post-mineral pebble dikes (Langerfeld, 1964; Gustafson and Hunt, 1975). Beneath the concentric pattern in the shallower levels of the system, Gustafson and Hunt (1975) documented a crude radial and concentric or elliptical arrangement of late-D veins (Fig. 12C). The patterns are broadly centered on two L-porphyry stocks, which are late intra-mineral intrusions that largely postdate sulfide mineral deposition. The northwestern L-porphyry stock directly underlies the center of one of the radial vein and dike arrays, whereas a composite stock of granodiorite porphyries cored by another L porphyry underlies the radial vein and dike array to the southeast. The association with the L porphyry implies a genetic linkage, and that the radial fractures formed late in the deposit's evolution. However, because the southeast array lies above a composite stock of X, K, and L porphyry, formation of that radial array could conceivably have been initiated earlier. A northwesterly elongation to the overall pattern suggests that late-D veins at depth reflect not only magmatic stress associated with emplacement of the composite porphyry stocks, but also the northeast–southwest-oriented minimum principal stress associated with regional transpressional deformation.

At deep levels in porphyry systems it has been proposed that a singular trend dominates the veins and fractures, as exemplified by the simple pattern at Sierrita, Arizona (Fig. 12D; Heidrick and Titley, 1982; Titley et al., 1986; Titley, 1993). Here, veins and fractures are parallel to quartz monzonite porphyry dikes emplaced into the host granitic rocks as part of the magmatic evolution of the system. The densest concentration of fractures is centered on the apophyses of quartz monzonite porphyry, which also localized the Sierrita and Esperanza porphyry Cu deposits. Support for this model is found at El Salvador, where late pebble dikes exploit older fracture networks that become progressively more linear and aligned to a northwesterly strike at depth (compare Figs. 11B and 12B, C). Strong linear arrays of veins are common in many deposits (Britten and Marr, 1995; Pierce and Bolm, 1995; Schroeter, 1995), which, according to the model of Heidrick and Titley (1982), should therefore represent the deep levels of porphyry Cu systems.

As with many models of porphyry systems, there are exceptions to these rules. Linear vein arrays at the shallow Marte porphyry Au deposit are clearly inconsistent with the model of Heidrick and Titley (1982), because that deposit can be reasonably assumed to have formed at depths <1 km (Vila et al., 1991). Exceptions are also found in several other deposits, where vein and fracture networks vary with depth, paragenetic stage, and location in the deposit (e.g., Wilkinson et al., 1982; Clark, 1990; Lindsay et al., 1995). At Chuquicamata, for example, interpretation of the vein geometry is complicated by the apparent superimposition of two discrete porphyry systems separated by as much as 3 m.y. (Reynolds et al., 1998). Nonetheless, copper-bearing veins associated with potassic alteration generally trend northeast and indicate formation under a regional stress field, in this case dextral strike-slip along the Domeyko fault system. Earlier quartz-molybdenite veins, in contrast, form a crude elliptical fracture pattern around the porphyry complex. Quartz-sericite-sulfide and enargite veins are superimposed on the potassic alteration at Chuquicamata, and their emplacement was controlled by old fractures and veins as well as the regional tectonic stress field

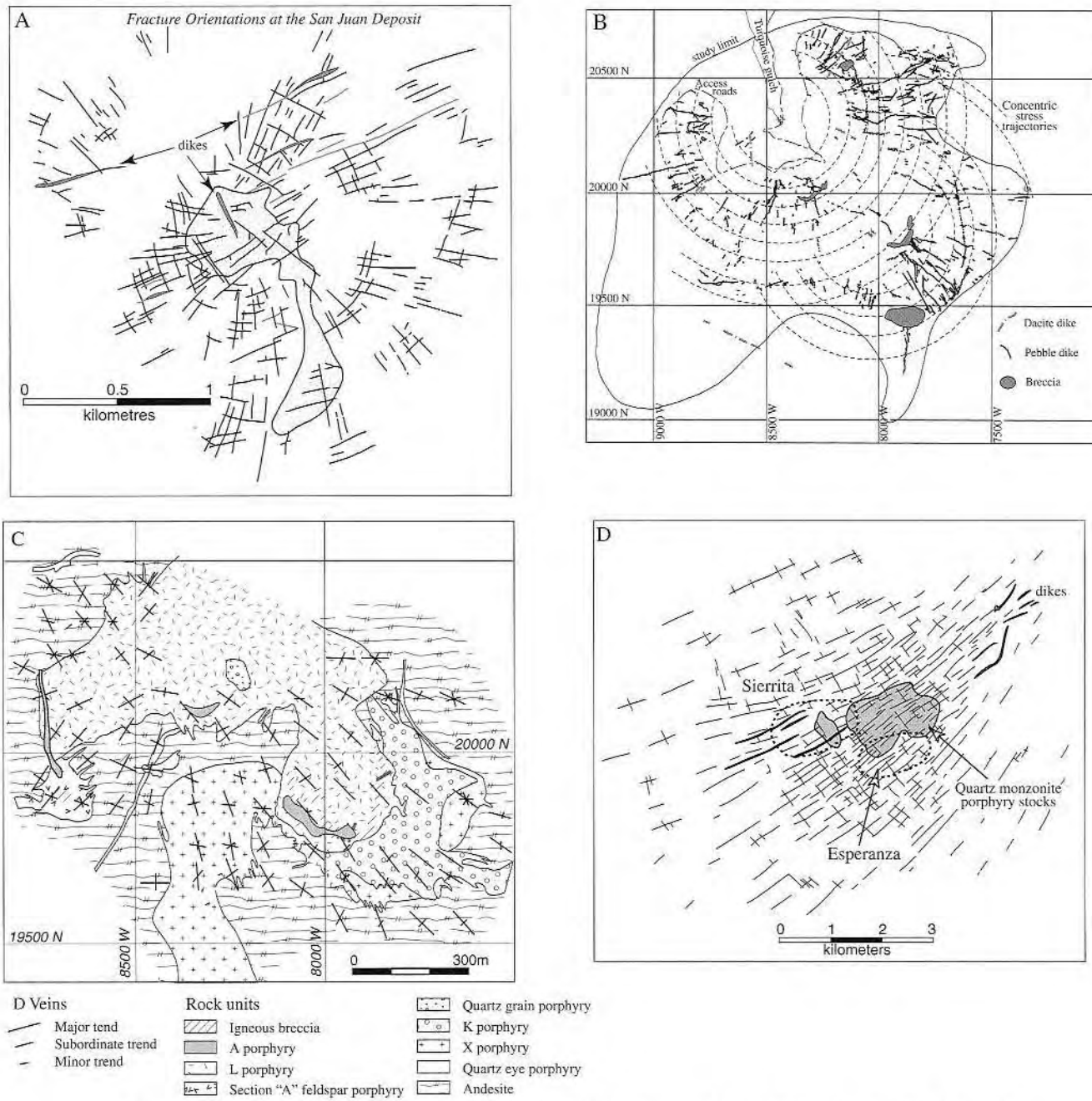

FIG. 12. Mineralized fracture and vein patterns developed in porphyry copper deposits. A. Concentric and radial mineralized fractures and veins developed at high levels in the San Juan mine area, Safford Mining District, Arizona. Note superposition of radial and concentric veins and mineralized fractures. In contrast, dikes do not conform to this pattern but form orthogonal sets. Modified from Heidrick and Titley (1982). B. Surface distribution of dacite dikes, pebble breccias, and hydrothermal breccias on Cerro Indio Muerto as mapped by Langerfeld (1964). Pebble breccias are related to the post-mineral dacite dikes and D veins, and are paragenetically late. The pebble dikes decrease in abundance with depth and generally lie in the upper parts of the deposit above a zone that separates sericite-dominated alteration from an overlying mixed sericite-kaolin alteration assemblage. With depth and where present, pebble dikes become more linear with dominantly northwesterly strikes (see C and Fig. 11B). The hydrothermal breccias are cemented by hydrothermal minerals and, thus, represent intermineral bodies. Superposed on the dikes are interpreted concentric stress trajectories (dashed lines) that suggest the presence of two domical stocks at depth. When compared with the geological map of the subsurface (C), the center of the concentric rings correspond generally to the center of the composite stock of X, K, and L porphyries to the southeast, and the main mass of L porphyry to the northwest. C. Distribution of country rock units, Eocene granodiorite porphyries, and cross-cutting D veins on the 2600- and 2660-m levels of the El Salvador porphyry deposit. Note the roughly radial and elliptical distribution of veins. Modified from Gustafson and Hunt (1975). D. Generalized fracture distribution and orientation around the Sierrita and Esperanza porphyry deposits, Arizona. Modified from Titley et al. (1986) and Titley (1993).

(Lindsay et al., 1995); evidently, the influence of magmatic pressure over tectonic stress was transitory. The enargite veins may represent shallower levels of a younger porphyry Cu deposit emplaced during uplift and unroofing of the older system (Sillitoe, 1994). It is interesting that at Chuquicamata, the dominant faults and mineralized fractures within the composite porphyry Cu deposit form a crude conjugate fracture network, a pattern predicted by numerical modeling of porphyry stocks (see above), despite the local structural control exerted by the Falla Oeste (Domeyko fault system). Conjugate vein patterns also characterize many other deposits, such as Panguna (Papua New Guinea; Clark, 1990) and Highland Valley (British Columbia; Osatenko and Jones, 1976; Casselman et al., 1995).

Implications for stress during mineralization

Despite the inherent beauty of Heidrick and Titley's (1982) vertical zonation model, the variety of vein orientations noted above suggests that it requires some modification. Concentric and radial fractures are expected to be concentrated above and exterior to the intruding stock to which they are related; where multiple intrusions are present, fractures should form overlapping arrays (Fig. 12B, C). In contrast, after solidification, the stock and its host rocks will undergo brittle fracture associated with tectonic and hydrothermal activity. The orientations of resulting fractures will reflect a new set of stress conditions controlled either by magmatic or tectonic forces. Magmatic stress results from renewed intrusion, whereas tectonic stress is externally imposed. Fractures developed under a tectonic stress field will reflect that field, and should be more ordered and linear. They will also develop later in the system's evolution, and will be concentrated within the stocks themselves. The elongate arrays of type-D veins within Eocene granodiorite stocks at El Salvador (Fig. 12C) and the concentric and radial arrays in the overlying rocks can be attributed to superposition of fracture systems derived from early magmatic- and later tectonic-dominated stress fields (Fig. 12B, C).

Multiple vein sets are normal in porphyry Cu deposits. The veins are usually cogenetic, and open episodically and repeatedly at different paragenetic stages during formation of the deposits. In addition, there is usually either a consistent vein orientation persisting throughout the deposit laterally and vertically, or the veins have a consistent angular relationship between them, or both. Where documented, veins are commonly orthogonal or conjugate, and are rarely completely randomly oriented with respect to one another. A common angular pattern is not what would be expected in the case where the veins reflect simple fracturing above a stock under conditions approaching uniaxial extension, or where they result from volume expansion caused by exsolution of hydrothermal fluids. It is the consistent deposit-scale angular relationship among veins which suggests that intrusion or hydrothermal-related models for vein formation are simplistic. Recognizing a common angular relationship is also critical to understanding the state of stress that dominates during porphyry Cu formation, which in turn has important implications for the tectonic setting of these deposits.

Taking the simplest case as illustrated in Figure 12, an orthogonal set of veins characterizes many porphyry systems (see also Stanley et al., 1995) regardless of depth of formation. In some deposits such as El Salvador (Fig. 12C) and locally in Sierrita (Titley et al., 1986), other veins bisect the orthogonal veins. During brittle failure of rocks, extensional fractures are the easiest to form, and are oriented normal to the minimum effective principal stress, σ_3' (Jaeger and Cook, 1979; Sibson, 2000, 2001). Such fractures should dominate fluid-saturated environments such as porphyry Cu deposits, and they will be filled by hydrothermal minerals whenever open. However, formation of extensional shear or compressional shear veins requires a higher differential stress (Sibson, 2000, 2001). Such fractures and veins lie at high angles to σ_3' and should bisect the intersection of orthogonal fractures; they will also be less common than extensional fractures unless a planar fabric is already present within the rock, or, if the numerical models of Gow and Ord (1999) are correct, little or no rheologic contrast exists between the stocks and host rocks. Hence, from simple rock mechanical consideration, formation of veins in multiple orientations during the life of the porphyry Cu deposit requires specific stress conditions. In view of the ease of forming extensional fractures, the simplest interpretation of the vein pattern in a porphyry Cu deposit, particularly of orthogonal sets, is that σ_3' changed orientation many times during the life of the system. In essence, the existence of orthogonal veins requires that σ_3' and the other horizontal stress, either the maximum effective principal stress (σ_1') or the intermediate effective principal stress (σ_2') depending upon the stress field, swapped episodically during the life of the system. This situation can be visualized by comparing the inferred stress trajectories above a stock in anisotropic rocks, such as is shown in Figure 7C and E.

Porphyry Cu deposits form over a short period of geological time based upon thermal models of cooling stocks (Cathles, 1977) and geochronological data (e.g., Chesley and Ruiz, 1997; Cornejo et al., 1997; Marsh et al., 1997). Because of this relatively short duration of hydrothermal activity, it seems unlikely that a significant and episodic rotation of the regional stress field would occur during deposit formation. Therefore, in order to explain the repeated rotation of the effective minimum stress direction called for above, a low-differential stress field, modified by fluctuating fluid pressures, may be the norm during porphyry Cu formation.

Origin of Breccia Pipes in Porphyry Cu Deposits

As noted previously, hydrothermal breccia pipes are of varying importance in porphyry Cu deposits. They range from insignificant in vein-dominated systems, to forming the major portion of the ore body in breccia-dominated systems (Sillitoe, 1985). Most breccia pipes are pre- to synmineral, but some pebble breccias are late and are usually barren (Richard and Courtright, 1958; Sillitoe, 1985). All hydrothermal breccias in the porphyry Cu environment, regardless of their genesis, reflect sudden expulsion of fluid at pressures that exceed the lithostatic load, thereby caus-

ing hydraulic fracturing (Burnham, 1985; Sillitoe, 1985). This fluid release is generally attributed to the process of second boiling in the porphyry magma, and rupturing of a pressure seal at the brittle–ductile transition (Fig. 4; Phillips, 1973; Burnham, 1985; Fournier, 1999). The resultant catastrophic fragmentation of the host rocks propagates upwards from a narrow source toward areas of reduced effective mean stress. Most porphyry Cu-related breccias are rooted in porphyry intrusions or dikes but apparently do not vent to the paleosurface (Sillitoe, 1985), resulting in lensoid, ovoid, or irregular pipe-like geometry. Hydrothermal minerals fill the resulting voids leading locally to bonanza concentrations of ore.

Zweng and Clark (1995) proposed that hydrothermal breccia pipes are more common in porphyry Cu deposits formed at shallow levels because of the decreased lithostatic load imposed by the mass of country rocks above the stock. However, not all shallow porphyry Cu deposits are dominated by breccias as shown by vein-dominated Cu-Au and Au-Cu systems emplaced within 1 km of the surface in the Maricunga belt, Chile. Hence, simple magmatic and fluid exsolution processes may not have been sufficient in all deposits to permit breccia formation. Because breccias are present to varying degrees in all porphyry Cu deposits, the obvious question is what controls their formation and abundance? Porphyry Cu deposits represent dynamic systems where magmatism and fluid circulation are affected by tectonic stresses, so there is no shortage of potential local causes that could trigger breccia formation. Amongst these local triggers are fluid-pressure build-up (Burnham, 1985), stock or dike intrusion that disrupts the evolving porphyry Cu system (Langerfeld, 1964), sector collapse of the overlying volcanic edifice (Sillitoe, 1994), and local seismic activity perhaps associated with intrusion and volcanism. Skewes and Stern (1994) also proposed that tectonic-driven changes in magmatism triggered breccia formation in the late-Miocene porphyry Cu deposits of central Chile. Still another potential trigger for breccia formation is remote (>500 km) earthquake activity, which has been shown to generate enhanced seismic activity in active geothermal and magmatic systems (Hill et al., 1993). Evidently, propagation of seismic energy into a magmatic or hydrothermal system is sufficient to induce gas separation leading to volume expansion (Linde et al., 1994; Sturtevant et al. 1996). The magmatic pressure increase could give rise to volcanic eruptions (Sahaglan and Proussevitch, 1992), whereas in the hydrothermal environment of a porphyry Cu deposit, increased fluid pressures could be sufficient to trigger rupturing of the pressure seal above the porphyry system, leading to generation of breccia pipes.

Tectonic Setting of Porphyry Cu Deposits

A simple model of porphyry Cu deposit genesis involves formation from hydrothermal fluids exsolved from calc-alkaline to alkaline magmas in a subvolcanic arc environment. However, many shallow-level porphyritic plutonic complexes are devoid of or have only weak porphyry-style mineralization. The barren Paleocene ring complexes of the San Jeronimo superunit in the Peruvian Coastal batholith are examples from an arc otherwise rich in porphyry Cu deposits of broadly similar age (Bussell, 1985; Pitcher, 1985; Zweng and Clark, 1995). On a more global scale, porphyry Cu deposits are sparse in the western Pacific (Uyeda and Nishiwaki, 1980), with notable exceptions in Papua New Guinea and the Philippines (MacDonald and Arnold, 1994; Sillitoe, 1997; Hedenquist et al., 1998). In addition, porphyry Cu formation is characteristically episodic and localized, being concentrated during limited time ranges within narrow, margin-parallel belts (Sillitoe, 1972, 1988; Titley and Beane, 1981; Clark et al., 1982; Damon et al., 1983). These considerations suggest that additional factors must be involved in determining whether a porphyry Cu deposit will form or not, including details of magma chemistry (e.g., water content, oxidation state) and tectonic setting (Uyeda and Nishiwaki, 1980).

A review of structural settings of porphyry Cu deposits indicates that there are no unique environments into which these deposits are emplaced. They are found within strike-slip fault zones with only limited displacement contemporaneous with mineralization (Maksaev and Zentilli, 1988; Clark et al., 1990; Sapie and Cloos, 1994; Lindsay et al., 1995), within regions affected by slightly older to concurrent transpressional or contractile strain and uplift (Titley and Heidrick, 1978; Heidrick and Titley, 1982; Olson, 1989; Tomlinson, 1994; Sillitoe, 1997), in areas undergoing limited extensional deformation (Presnall, 1997), and in areas lacking major active fault systems (Sillitoe, 1997). Many but not all deposits are localized along older fault systems that provided crustal permeability (Heidrick and Titley, 1982; Richards et al., 2001). Significant porphyry Cu deposits are, however, seemingly absent from regions undergoing large-scale extensional strain, a negative association that has long been recognized, and which seems to preclude those arcs or times within arcs where such tectonics dominate. A common theme throughout these convergent arcs is the limited deformation that is demonstrably contemporaneous with porphyry Cu development.

Times of porphyry Cu generation have been causally correlated with times of shallow-dipping subduction (Sillitoe, 1972; Nielsen, 1976; Uyeda and Nishiwaki, 1980; Titley and Beane, 1981), increased convergence rates (Clark et al., 1990), subduction of aseismic ridges (Skewes and Stern, 1995), subduction zone reversals (Solomon, 1990), or following uplift (Sillitoe, 1997). Many but not all porphyry Cu deposits in Arizona (McCandless and Ruiz, 1993), southern Perú (Clark et al., 1990), Chile (Maksaev and Zentilli, 1988; Skewes and Stern, 1995; Richards et al., 2001), and Papua New Guinea (Titley and Heidrick, 1978; Sillitoe, 1997) formed near the end of major periods of magmatism, contractile or transpressional strain, and uplift. Sillitoe (1997) suggested that rapid uplift in an arc is an essential precursor to the formation of giant porphyry Cu deposits, which appear near to or just after the end of that event.

The above observations, combined with the near-circular plan of many porphyries and the common deposit-scale ori-

entations of associated veins and dikes, suggest that porphyry Cu deposits are developed during periods of low-differential or near-isotropic horizontal stress in the arc. The achievement of such stress states is likely driven by fundamental changes in subduction geometry and convergence rates. For example, changes in the coupling of stress between the subducting and overriding plate may lead to shifts in the location of zones of upper-plate deformation with respect to the magmatic arc, either inboard toward the back-arc or outboard toward the convergent margin. This shift in the locus of deformation relative to magmatism may permit the arc to approach a neutral stress state. The giant Miocene and Pliocene porphyry Cu deposits in central Chile (Los Pelambres, Rio Blanco-Los Bronces, El Teniente) seem to fit this scenario because they formed in the hanging wall of a fold-and-thrust belt that lies to the east in the Argentinean foreland (Jordan et al., 1993; Ramos et al., 1996). In an alternative scenario, relocation of the volcanic arc may be accompanied by relaxation of compressional or transpressional stress in an old, established, plutonic-volcanic arc, resulting in late-stage development of porphyry magmatism. The late-Eocene–early-Oligocene porphyry Cu deposits of northern Chile are proposed to have formed during such a period of stress relaxation following a prolonged episode of Eocene transpression and effusive volcanic activity; relaxation coincided with flattening of the subduction angle and subsequent inland shift of magmatism (Maksaev and Zentilli, 1988; Clark, 1993; Richards et al., 2001).

These characteristics may be speculatively tied to models of magmagenesis and crustal emplacement as reviewed above (Fig. 13). First, the simple textbook form of a destructive margin—that of orthogonal compression, deep-crustal or mantle melting, and widespread effusive magmatism—is not conducive to porphyry Cu formation (Fig. 13A). Neither is a state of arc rifting (extension), where crustal melting is short-circuited and mantle-derived basaltic melts ascend directly to the surface (Fig. 13B) conducive to porphyry Cu formation. Instead, the various tectonic environments identified during porphyry Cu generation around the world suggest that ideal conditions form during transitions between these end-member states, usually caused by changes in subduction geometry or rate. Geochemical studies show that, even in island-arc settings, porphyry Cu-related magmas are not direct products of asthenospheric melting in the mantle wedge above the Benioff zone, but result from significant amounts of fractionation and interaction of these magmas with the overlying lithosphere. Isotopic and geochemical evidence for crustal interaction in Central Andean magmas, for example, is overwhelming (Tilton et al., 1981; Barreiro, 1984; Harmon et al., 1984; Pankhurst et al., 1988; Walker et al., 1991). A period of "stewing" near the base of the lithosphere is implied, involving extensive MASH (melting, assimilation, storage, and homogenization; Hildreth and Moorbath, 1988) processes in voluminous but probably diffuse, batholithic magma chambers. Conditions appropriate to this activity are achieved during "classic" convergence periods in the arc, compressive stresses in the upper plate acting to close major orogen-parallel structures, and inhibiting ascent of all but the most overpressured magmas (Fig. 13A). Those magmas that do erupt do so violently, producing widespread effusive volcanic sequences but largely unaccompanied by shallow-level plutonism.

If compressive stress in the lithosphere is locally relaxed, however, magma dynamics subordinate tectonic stress, and melts may ascend through the crust more passively driven by buoyancy forces, rather than violently by magmatic overpressure (Fig. 13C, D). In consequence, such magmas tend to rise to levels of neutral buoyancy in the shallow crust, where they may undergo further fractionation and crustal interaction, and potentially generate porphyry-style deposits as these distillate magmas cool and saturate with volatiles. Volcanism, where it occurs, is more localized, forming discrete composite edifices directly above shallow-source magma chambers.

An analog may be found in the use of a pressure cooker. Effective stewing is achieved at elevated temperatures and pressures by sealing the pot with a lid and a safety valve (periodic violent eruptions occur only when high-pressure fluids overcome the valve weight). Sudden removal of the lid (equivalent to extensional tectonics, and not advised at home) will result in catastrophic eruption of the contents of the pot, equivalent to basaltic rift magmatism. To preserve the stewed contents just right, the pot needs to be depressurized (stress relaxation) before the lid is removed.

In this model, localization of porphyry magmatism in the crust can be understood in terms of structural permeability. During periods of arc-normal compression, major arc-parallel structures will be closed, and magma penetration will be restricted. Upon relaxation of stress, however, these structures will provide planes of elevated permeability through the crust. More specifically, fault jogs or structural intersections may generate pull-apart volumes when oblique compressive stress relaxes (Fig. 13C, D). Such volumes represent discrete vertical columns of maximum permeability through the crust, which may serve as optimum loci for magma ascent and potential porphyry Cu generation. Relatively low degrees of differential horizontal stress are implied by this model, because fault lock-up under high strain will progressively destroy permeability (Sibson, 2001).

Concluding Remarks

Porphyry Cu deposits require the precise coincidence and positive interaction of a series of individually commonplace geological processes. The role of magmas and their exsolved hydrothermal fluids have long been recognized and widely studied, and are now reasonably well understood. In contrast, the detailed influence of structure and tectonic setting on deposit localization and evolution in the porphyry environment is under-appreciated, despite several careful studies. In part, this situation stems from the immense scale of porphyry systems (commonly exceeding diameters of 10 km) and the inherent difficulties that this poses to structural analysis. It also derives from the research focus on mineralogic and geochemical processes during the

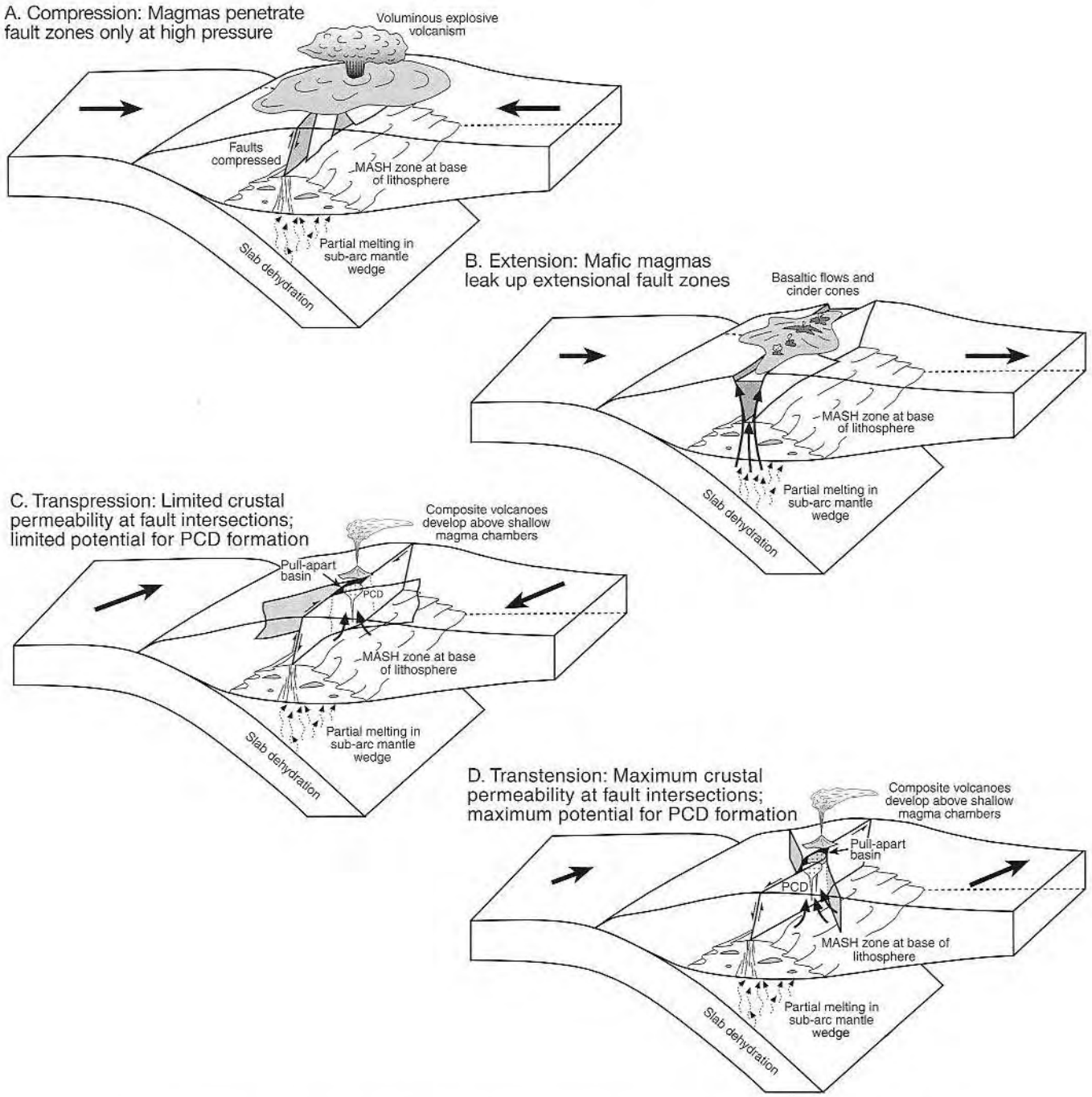

FIG. 13. Cartoon depiction of the relationship between convergent margin tectonics, upper plate structures, and magmatism. A. In orthogonally convergent settings, major arc-parallel fault zones are compressed. Magmas generated in the mantle wedge pond near the base of the lithosphere, and interact with it through processes of melting, assimilation, storage, and homogenization (MASH processes; Hildreth and Moorbath, 1988). Ascent to the surface is restricted to magmas under high overpressure, which erupt violently without significant residence in upper crustal magma chambers. B. Arcs under tension provide multiple high-permeability paths for magma ascent, to the extent that asthenosphere-derived magmas may rise directly to the surface. The distillation and crustal interaction processes that appear to be necessary for the development of porphyry Cu-prospective magmas are thereby short-circuited, and porphyry Cu deposits are not found in rifting arcs. C and D. Upon relaxation of stress, localized zones of extension may be generated during transpressional or, optimally, transtensional strain along arc-related structures. These zones provide high-permeability conduits for the passive ascent of magmas to shallow crustal levels, driven primarily by buoyancy rather than magmatic overpressure. Hypabyssal plutonism with associated localized volcanism is favored, and is an environment conducive to porphyry Cu formation. PCD = porphyry Cu deposits.

period of major porphyry Cu exploration and development in the 1960s to 1980s (Titley and Hicks, 1966; Sutherland Brown, 1976; Titley, 1982; Pierce and Bolm, 1995; Schroeter, 1995). Nevertheless, structure and tectonics clearly play a major role in porphyry Cu emplacement and genesis, and may be critical in determining where and when spatially and temporally localized porphyry Cu provinces form within convergent plate margins.

Porphyry Cu deposits and all their genetically associated deposits are a natural consequence of convergent margin magmatism. Nevertheless, the preceding review suggests that significant porphyry Cu deposits form in specific structural environments under a near-neutral stress state. These structural conditions appear to occur during periods of little deformation or in areas removed from deformation, or during stress relaxation in the magmatic arc, conditions likely to be infrequent and transitory during the life of a convergent margin. This hypothesis suggests that detailed research into arc history and dynamics may be fruitful in predicting the loci not only of porphyry Cu provinces, but also of individual porphyry centers within those provinces.

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