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# Tectono-Magmatic Precursors for Porphyry Cu-(Mo-Au) Deposit Formation

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## Abstract

Porphyry Cu-(Mo-Au) deposits are relatively rare but reproducible products of subduction-related magmatism. No unique processes appear to be required for their formation, although additive combinations of common tectono-magmatic processes, or optimization of these processes, can affect the grade and size as well as the location of the resulting deposits. These various contributing processes are reviewed, from partial melting in the mantle wedge overlying the subducting plate, through processes of magma interaction with the lithosphere, to mechanisms for magma emplacement and volatile exsolution in the upper crust. Specific ore-forming processes, such as magmatic-hydrothermal fluid evolution, are not discussed.

Hot, hydrous, relatively oxidized, sulfur-rich mafic magmas (predominantly basalts) generated in the metasomatized mantle wedge above a subducting oceanic slab rise buoyantly to the base of the overlying crust where they stall due to density contrasts. Because these magmas are oxidized, sulfur is dominantly present as sulfate, and chalcophile elements such as Cu and Au are incompatible (i.e., they are retained in the melt). As these magmas begin to crystallize they release heat which causes partial melting of crustal rocks. Mixing between crustal- and mantle-derived magmas yields evolved (andesitic to dacitic), volatile-rich, metalliferous, hybrid magmas, which are of sufficiently low density to rise through the crust. Magma ascent is driven primarily by buoyancy forces and is dominantly a fracture-controlled phenomenon. As such, crustal stress and strain patterns play an important role in guiding the ascent of magma from the lower crust. In particular, translithospheric, orogen-parallel, strike-slip structures serve as a primary control on magma emplacement in many volcanic arcs worldwide. A feedback mechanism operates, whereby preexisting faults facilitate magma ascent, the heat from which further weakens the crust and focuses strain. Certain structural geometries, such as fault jogs, step-overs, and fault intersections, offer low-stress extensional volumes during transpressional strain. Such sites represent vertical conduits of relatively high permeability, up which magmas will preferentially ascend. Large upper crustal plutonic complexes may therefore be localized within these structural settings. Having delivered a sufficient volume of evolved, fertile arc magma to a focused position in the upper crust, magmatic fractionation, recharge, and volatile exsolution lead to the development of ore-forming magmatic-hydrothermal systems. To a first approximation, the size of the resulting deposit will be limited by the magma volume delivered to the upper crustal magma chamber. System-specific details such as magmatic-hydrothermal evolution, the nature of the country rocks, and subsequent erosional and weathering history will ultimately control the value of the deposit, but these factors fall outside the scope of this paper.

## Introduction

PORPHYRY Cu-(Mo-Au) DEPOSITS (hereafter referred to as porphyry Cu deposits) are the world's primary source of Cu and Mo and an important source of Au. The occasionally giant size of these deposits (several deposits over 1 billion metric tons @ >0.5% Cu; e.g., Bingham Canyon, Utah; Collahuasi, Chuquicamata, La Escondida, El Teniente, Chile) makes them valuable exploration targets. These deposits are formed in association with subduction-related magmas and are found sporadically in magmatic arcs worldwide. Their formation involves the exsolution of metalliferous and sulfur-rich hydrothermal fluids from calc-alkaline arc magma and deposition of ore minerals in response to fluid phase separation, cooling, wall-rock reaction, and mixing with external fluids. Exsolution of magmatic volatiles is a common phenomenon in cooling intrusive rocks but its extent in large porphyry deposits, as evidenced by the scale of hydrothermal alteration and mineralization, implies an optimization of processes in both space and time. These various processes are not in themselves rare or unique, but the sequence of their combination and the magnitude of their effects are crucial in determining whether conditions suitable for ore formation will be achieved (e.g., Henley and Berger, 2000). In support of this argument, it is noted that the broad, global uniformity of porphyry Cu deposits, in terms of their associated magmatism, alteration, and mineralization styles, has been demonstrated in many studies, and the seminal work of Lowell and Guilbert (1970) still stands as the type-description of this class of deposit (although, of course, variants exist). This uniformity would seem to preclude the essential involvement in ore formation of any process that is not common in arc tectonics and magmagenesis, and I herein adopt the conclusion of Dilles (1987) and Cline and Bodnar (1991) that calc-alkaline magmas in general have the potential to form porphyry Cu deposits.

Most exposed deposits in known porphyry Cu districts have been discovered, and present-day exploration is focused on searching for covered deposits, using indirect geophysical and geochemical methods and geological information derived from distal exposures. Such strategies require knowledge of geological history on a regional scale and an understanding of porphyry Cu genesis within the broader context of tectono-magmatic arc processes, in addition to a deposit-scale appreciation of ore-forming processes. The latter subject has been the focus of intense study by economic geologists for many decades and is not discussed here (see reviews by Beane and Titley, 1981; Titley and Beane, 1981; Hedenquist and Richards, 1998; Henley and Berger, 2000; Richards, 2004),

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but the study of large-scale tectono-magmatic processes has generally been left to other disciplines in the geological sciences. In an attempt to bring some of the insights from this research to bear on the question of porphyry Cu metallogeny, I review current hypotheses for magma generation and transport in arcs and discuss the role of tectonism in controlling the timing and localization of porphyry Cu-forming magma emplacement and volatile exsolution. It is concluded that the prospectivity of magmatic suites in arcs can be evaluated from studies of regional tectono-magmatic history, and that large porphyry copper deposits around the world share a common relative timing at the end of major tectono-magmatic cycles.

## Arc Tectonics

Arc magmatism is inextricably linked to tectonic processes at convergent plate margins; as such, all arc magmas are syntectonic or synkinematic (Vigneresse, 1995b). Considerations of magma genesis, transport, and emplacement cannot, therefore, ignore the role of tectonic stress and strain. Convergent margin dynamics imply a differential stress field, which is often assumed to be compressional in the direction of convergence between the impinging plates. This view is overly simplistic at several levels, however. Although compressional stress characterizes many convergent margins (Zoback, 1992), others are clearly under tensional stress (e.g., Uyeda and Kanamori, 1979; Mercier, 1981; Hamilton, 1988, 1995). In particular, the existence of tensional stress in back-arc environments and during arc rifting is well known, and Hamilton (1995, p. 7–8) has gone so far as to say that "the common regime above subducting slabs is extensional and not, as in popular fantasy, compressional." The source of this tension lies partly in the fact that most slabs, particularly older and colder ones, have negative buoyancy relative to the asthenosphere and are actually sinking away from the trench axis (slab rollback). The upper plate is therefore drawn toward the trench by trench suction or slab pull (Bott et al., 1989; Apperson, 1991; Royden, 1993; Shemenda, 1993). Tension in the back-arc region of the upper plate may reflect transmission of this stress from the trench but may also relate to asthenospheric upwelling, or to crustal thickening, uplift, and weakening, leading to local gravitational collapse (Apperson, 1991; Ziegler, 1992).

In addition to normal compressional and tensional stress, shear stress is ubiquitous in destructive margins because the convergence direction is rarely orthogonal. Compressive or tensional shear stress may be transmitted into the upper plate by frictional coupling, where strain is commonly partitioned into contractional and/or extensional and shear components (Jarrard, 1986; Apperson, 1991; Teyssier et al., 1995; McNulty et al., 1998), resulting in the coexistence of crustal shortening and/or extension and strike-slip faulting. Thus, contractional folds or thrusts may form in association with strike-slip faults in a transpressional orogen, and extensional domains may exist within shear or contractional structures, such as pull-aparts at fault step-overs or in fold hinges.

From the above it is clear that there is no unique set of stress conditions in the overriding plate in collisional arcs; they may vary from compressional to tensional as well as shear and may vary in three dimensions (vertically through the lithosphere, laterally along the arc, and transversely from fore- to back-arc). Additionally, and most importantly, stress conditions vary over time and sometimes on very short time scales (<1 m.y.; Bott et al., 1989; James and Sacks, 1999). Transmission of compressional or shear stress into the overriding plate requires frictional coupling with the downgoing slab, which in turn depends on a number of parameters such as convergence rate, relative convergence vector, slab dip, and slab buoyancy. It is well known from sea-floor spreading records and hot-spot traces that convergence rates and directions change frequently on a time scale of millions of years or less (e.g., Pilger, 1984; Pardo-Casas and Molnar, 1987), and slab dip and buoyancy (at least relative to the arc) change over periods of a few million years (Soler and Bonhomme, 1990). In addition, intermittent and diachronous stresses may be caused by subduction of anomalous features on the sea floor, such as seamounts and ridges (e.g., Ramos and Kay, 1992; Bangs and Cande, 1997; Ramos et al., 2002).

With such rich scope for variability of stress fields and resultant strain in the upper plate, the existence of long-lived tectonic features such as cordilleran arcs might seem surprising. However, when studying the geologic history of such arcs, it rapidly becomes clear that they do not represent steady-state conditions but are in a constant state of flux on a time scale similar to the ones noted for stress change (e.g., Coira et al., 1982; Jordan and Gardeweg, 1989). Within this context, the epochal nature of porphyry copper genesis in many arc systems can begin to be rationalized, and a possible explanation in terms of transient arc tectono-magmatic processes is offered below.

## Arc Magmagenesis

Theories of arc magmagenesis have evolved substantially since early models, which proposed that andesitic magmas were formed by direct melting of the subducted slab (although adakitic lavas may represent rare examples of such melts formed under special conditions of shallow subduction of young, buoyant slabs; Defant and Drummond, 1990; Sajona et al., 1993; Peacock et al., 1994; Martin, 1999; Yogodzinski et al., 2001). It now seems likely that dehydration at the blueschist-eclogite transition at a depth of ~100 km, rather than melting, is the key process affecting most subducting slabs (Fig. 1; Ringwood, 1977; Wyllie, 1978; Tatsumi, 1989; Davies and Stevenson, 1992; Peacock, 1993). Solute-rich aqueous fluids released from the slab metasomatize the overlying wedge of mid-ocean ridge basalt-like (MORB-like) asthenospheric mantle, enriching it in volatiles, sulfur, silica, and fluid-mobile large ion lithophile elements (LILE), such as Rb, K, Cs, Ba, and Sr (Tatsumi et al., 1986; Davidson, 1996; de Hoog et al., 2001). Certain high field strength elements (HFSE), such as Ti, Nb, and Ta, are not mobilized by this process, however, and may be retained in the downgoing slab in minerals such as rutile (Brenan et al., 1994; Foley et al., 2000). The effect is to enrich the asthenospheric wedge in volatiles and LILE but not Ti, Nb, and Ta. Alternative models to explain this relative depletion in Ti, Nb, and Ta suggest that rutile or titanite are retained as a restite phase during melting of the slab or the mantle wedge (e.g., Ryerson and Watson, 1987; Foley and Wheller, 1990; Prouteau et al., 1999).

Hydration and metasomatism of the peridotitic subarc mantle wedge generate new mineral phases such as amphibole

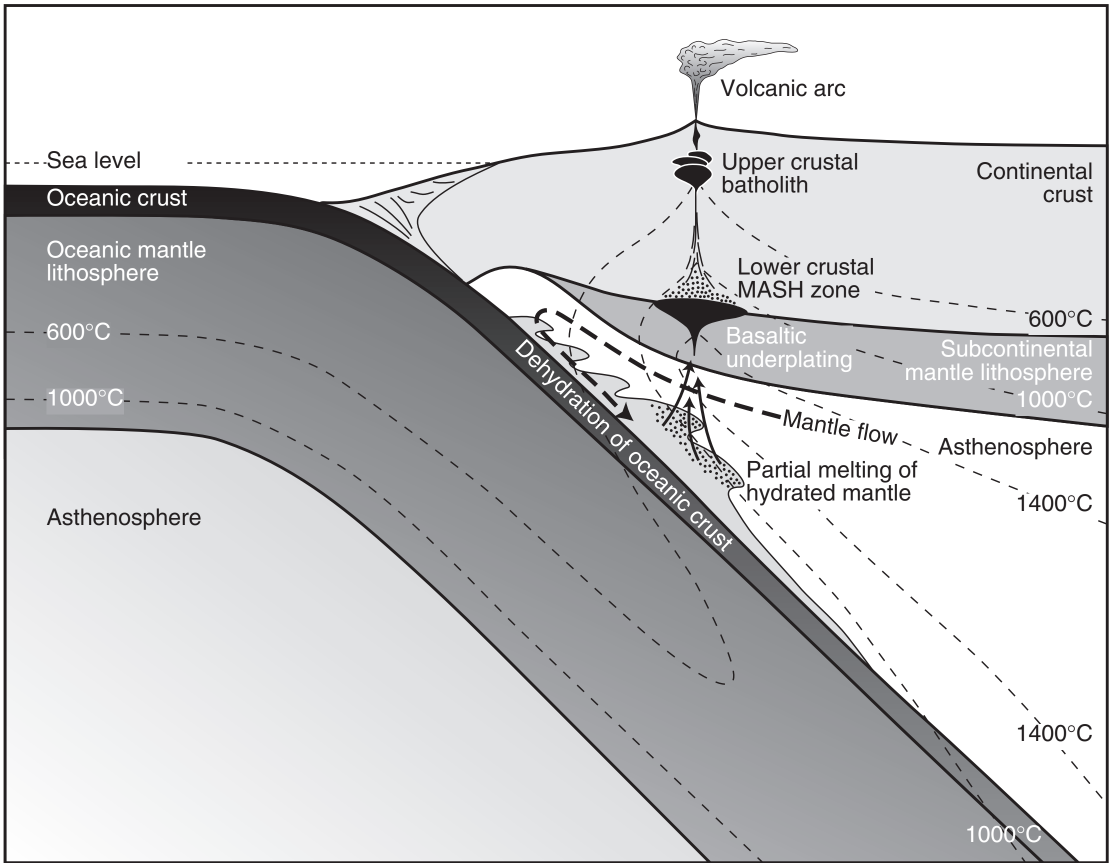


FIG. 1. Cross section through a subduction zone and continental arc (modified from Winter, 2001). Dehydration of the subducting oceanic crust leads to hydration of the overlying mantle. Partial melting occurs when this hydrated material is convected into hotter regions of the asthenospheric mantle wedge. Hydrous basaltic melts intrude the overlying lithosphere and pool at the base of the crust (a density barrier), where they fractionate and interact with crustal materials (MASH process: see text for details). More evolved, less dense magmas rise to upper crustal levels.

and mica and lower the mantle solidus temperature to the point at which melting begins (Tatsumi et al., 1986; Peacock, 1993; Arculus, 1994). The products of such melting are basaltic but are distinguished from MORB by their higher $H_2O$ and LILE and anomalously low Ti, Nb, and Ta contents, reflecting the metasomatized source composition (Fig. 2; Ringwood, 1977; Perfit et al., 1980; Pearce, 1983; Plank and Langmuir, 1988; Arculus, 1994; Stolper and Newman, 1994; Pearce and Peate, 1995). However, eruption of primitive[1] arc basalts is rare except in immature island arcs, and andesites and dacites predominate in continental arcs (Hildreth and Moorbath, 1988; Carmichael, 2002).

Arc andesites have been modeled as direct differentiates from primary mantle melts (e.g., Ringwood, 1977; Grove and Kinzler, 1986). However, there is now overwhelming evidence for multiple and multistage processes in andesite petrogenesis, for example, involving crustal melting and assimilation by primary basaltic magmas, magma storage at the base of the crust, and magma homogenization (as envisaged in the MASH model of Hildreth and Moorbath, 1988; see also DePaolo and Wasserburg, 1977; Hawkesworth, 1982; and Brown et al., 1984, for early discussions). Despite the apparent complexity of this multicomponent MASH process, the global uniformity and distribution of arc andesites (e.g., Gill, 1981) suggest that it is governed by repeatable and predictable mechanisms.

Fundamental to the MASH hypothesis is that mafic magmas ascending from the mantle wedge are more dense than most crustal rocks (Herzberg et al., 1983; Fig. 3) and will therefore pool near the base of the crust, forming an underplated layer (Hildreth, 1981; Fyfe, 1992; Fig. 1). The crust acts as a density filter, and the crust-mantle boundary represents the level of neutral buoyancy for mafic magmas (Walker, 1989). Note that the concept of level of neutral buoyancy is distinct from the concept of hydraulic head (Walker, 1989; Lister and Kerr, 1991). In theory, a magmatic system with hydraulic connectivity from the asthenosphere to the surface would always erupt because the bulk density of the lithosphere is greater even than that of mafic magma (Fig. 3). However, the ductility of the lithosphere serves to break this

[1] Definition of the terms "primitive" and "primary" magma follows that of the Glossary of Geology, 4th edition (Jackson, 1997): A primary magma is one that has not been chemically modified since its extraction from the source region (in this case the mantle), whereas a primitive magma is one that has not significantly evolved from the primary magma composition. In particular, primitive magmas have high magnesium numbers and high concentrations of compatible elements such as Ni and Cr, indicating little fractionation of olivine and spinel.

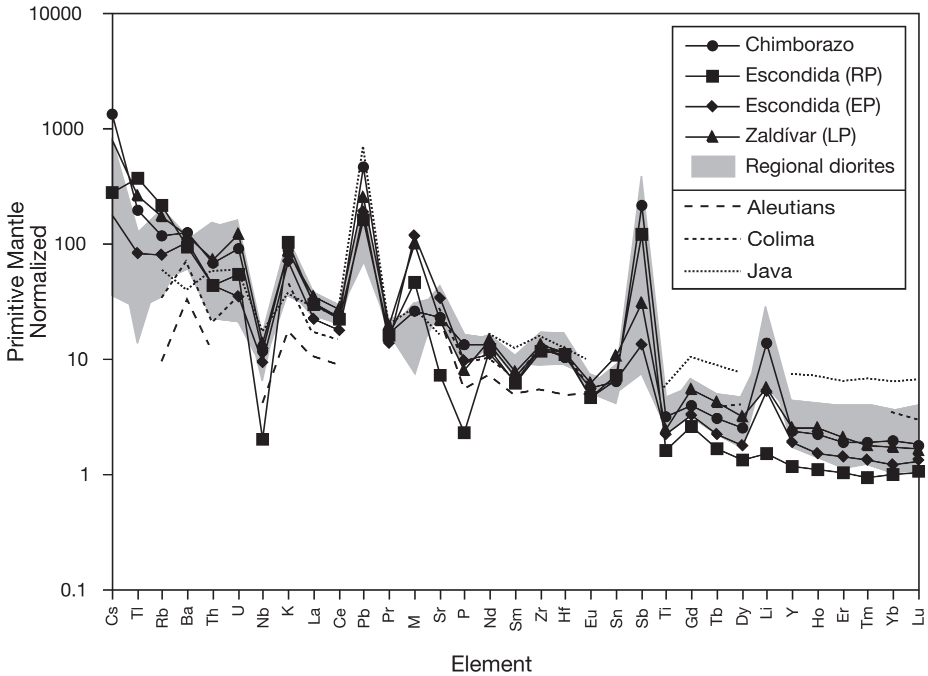


FIG. 2. Primitive mantle normalized trace element compositions of selected primitive arc volcanic rocks, compared to compositions of late Eocene-early Oligocene porphyry Cu intrusions from Escondida, Zaldívar, and Chimborazo, northern Chile, and coeval regional diorite intrusions. Negative anomalies for all rocks in Nb and Ti (Ta not plotted) and positive anomalies in Pb, Sb, and Li, as well as a general enrichment in incompatible elements, are characteristic features of arc magmas (Sun and McDonough, 1989; Pearce and Peate, 1995). Note that the geochemical patterns for the porphyries and diorites are closely similar, suggesting that they are cogenetic. In detail, the more evolved (dacitic to rhyolitic) porphyries display higher concentrations of incompatible elements (left side of diagram) and lower concentrations of compatible elements (right side of diagram) compared to the diorites, consistent with fractionation from dioritic parent magmas. Data from DeBari and Sleep (1991), Luhr (1992), Richards et al. (2001), and Reubi et al. (2002); primitive mantle normalization values from Sun and McDonough (1989). EP = Escondida porphyry, LP = Llamo porphyry, RP = Rhyolitic porphyry.

connectivity by conduit collapse, such that magmas rise largely in response to local buoyancy forces. An analogy may be sought in a siphon made from a punctured or soft rubber hose: the hose will leak or deform due to the fluid pressure and it will be impossible to draw liquid up to the level of its hydraulic head in the reservoir. In contrast, a rigid hose will sustain the excess fluid pressure and support the hydraulic head.

The hydrous nature of primary arc magmas (1.2–2.5 wt % $H_2O$; Sobolev and Chaussidon, 1996) results in the suppression of plagioclase precipitation and crystallization of olivine, pyroxene, spinel, and hornblende over an extended fractionation range (up to 50%; Müntener et al., 2001) as the magma cools. Precipitation of these dense mafic minerals leads to the development of thick ultramafic cumulate layers at the base of the crust, which may define the seismic Moho in evolved continental arcs (Hildreth, 1981; Herzberg et al., 1983; Huppert and Sparks, 1988; Bergantz and Dawes, 1994; Müntener et al., 2001).

The accumulating volume of mafic magma also represents a significant addition of heat to the base of the crust, which is released as the magma begins to crystallize and fractionate to more evolved and volatile-rich compositions (Fig. 4; Green, 1982; Herzberg et al., 1983; Huppert and Sparks, 1988). This build-up of heat, combined with invasion by increasingly evolved and hydrous magmas, will cause partial melting and assimilation of lower crustal rocks. The formation of a layer of lower density hybrid magma will further limit the ability of dense mafic magma to penetrate the crust, and this zone will become a region of extensive interaction and exchange between mantle- and crust-derived materials (Bergantz and Dawes, 1994). The product of this process will be a melt of intermediate (basaltic andesitic to dacitic) composition, to which Hildreth and Moorbath (1988) suggest that the deep crust may have made a contribution of up to tens of percent. Most importantly from the point of view of metallogenic potential, these evolved hybrid melts will be enriched in volatiles, sulfur, and other incompatible chemical components. The relatively high oxidation state of arc magmas (up to two log $f_{O_2}$ units above the fayalite-magnetite-quartz buffer; Brandon and Draper, 1996) ensures that the bulk of the sulfur is dissolved in sulfate form (Carroll and Rutherford, 1985), with the result that sulfide-compatible (chalcophile) elements such as Cu and Au will also behave as incompatible elements and will be retained in the evolving magmas (Hamlyn et al., 1985; Bornhorst and Rose, 1986; Richards et al., 1991; Spooner, 1993; Richards, 1995). Such evolved and hydrous melts have densities comparable to granodioritic and granitic rocks and will therefore have sufficient buoyancy to rise into the upper crust (Fig. 3; Herzberg et al., 1983; Walker, 1989).

Hildreth and Moorbath (1988) described the MASH zone as a complex of intrusions, dikes, and sills and suggested that

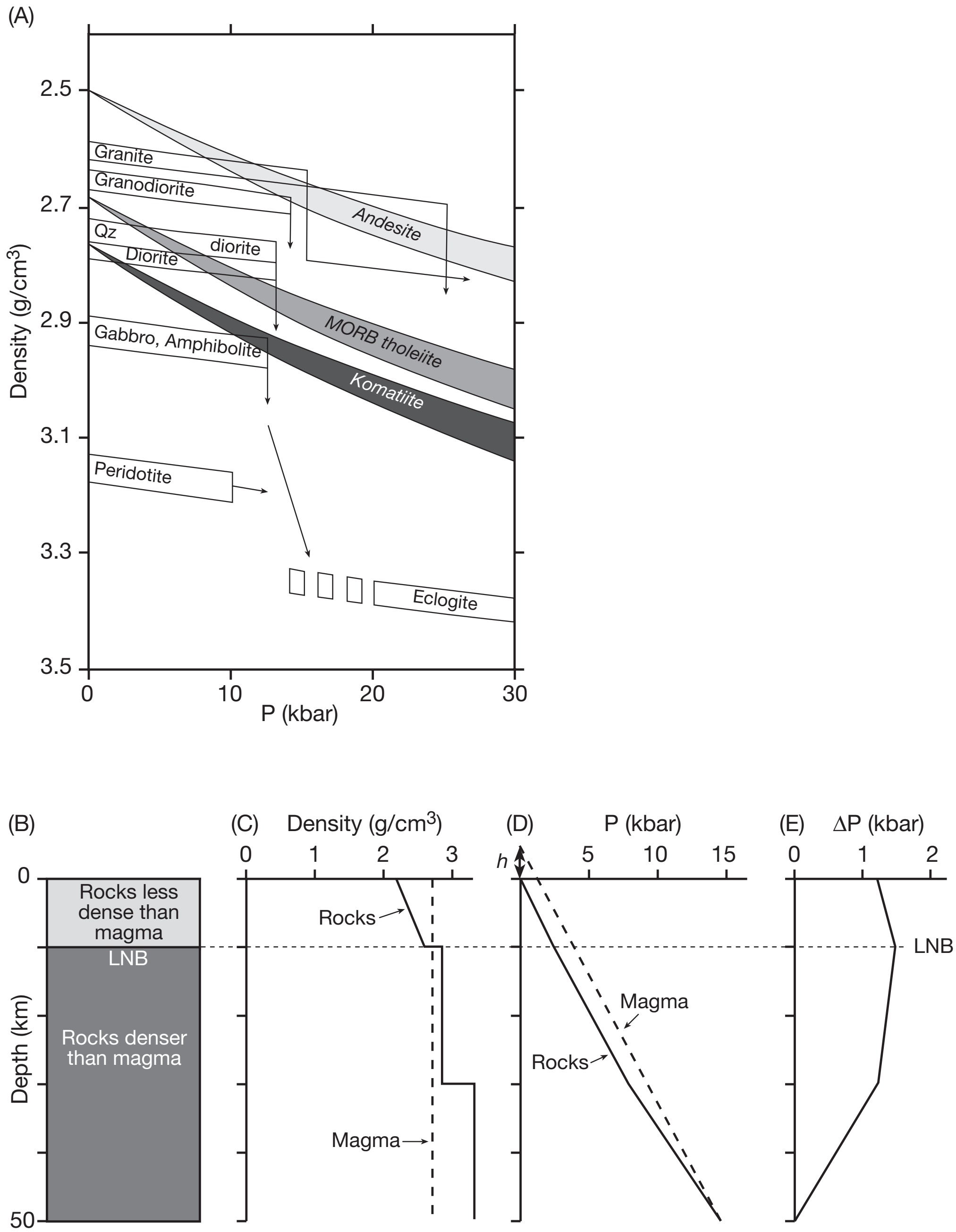


FIG. 3. A. Variation of densities of magmas and rocks with pressure (after Herzberg et al., 1983). Basaltic magmas are denser than typical continental crustal rocks and may become trapped at the base of the crust. In contrast, andesitic magmas are less dense than most crustal rocks and become increasingly so as they depressurize on their ascent toward the surface. B.–E. Depiction of the effect of contrasting density in a crustal column on the level of neutral buoyancy (LNB) of a magma (after Walker, 1989). Magma in a chamber at 50-km depth experiences a lithostatic pressure due to the weight of the column of overlying rock (C, D). Because the net density of the overlying rock is greater than that of the magma (C), if hydraulic connection to the surface is made (e.g., along a dike), the magma will have a hydrostatic head ($h$) and may erupt (D). However, if hydraulic connectivity is blocked (e.g., by conduit collapse), then local buoyancy forces will restrict the ascent of magma to the LNB.

"the distinction between mixing of magmas and ductile mixing of partially molten rocks may blur" (Hildreth and Moorbath, 1988, p. 483). They further proposed that the base-level geochemical and isotopic signatures of local arc magmatic suites are defined by the mix achieved in their source MASH zones. Subtle geochemical variations between volcanic centers within a given arc may therefore reflect the vagaries of the MASH process, rather than fundamental inhomogeneities in the asthenospheric mantle source (e.g., Kay et al., 1991, 1999; Wörner et al., 1992, 1994; Feeley and Hacker,

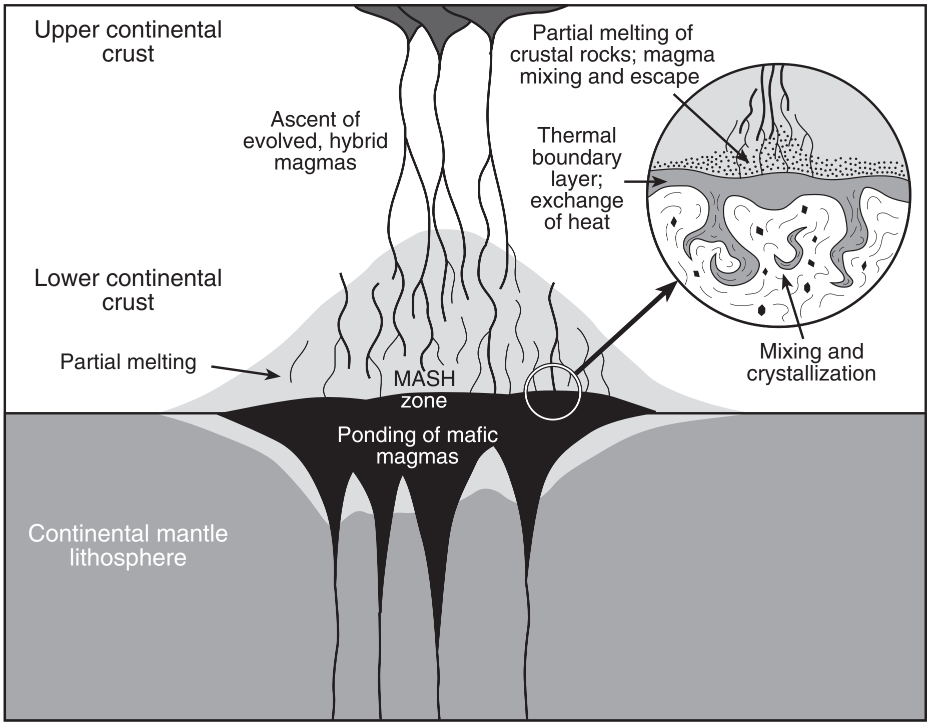


FIG. 4. Visualization of the MASH zone at the base of the crust, where hot, dense, mafic magmas pool and interact with crustal rocks to generate less dense, hybrid, andesitic-dacitic magmas (modified from Hildreth, 1981; inset from Huppert and Sparks, 1988).

1995; Richards and Villeneuve, 2002). Support for this geochemical model is found in recent multitechnique geophysical surveys of the Central Andes, which clearly indicate the existence of partial melts (up to 20% by vol) throughout the crustal column beneath the active volcanic arc (Schmitz et al., 1997; Masson et al., 2000; Schilling and Partzsch, 2001). In particular, Schilling and Partzsch (2001) argued on the basis of thermal considerations that these magmas are felsic and therefore contain a high proportion of crustal anatectic melt.

It is emphasized that arcs are evolving tectono-magmatic systems both in space and time, and that the development of MASH zones is likely to be an ephemeral process. At the onset of subduction or after a fundamental shift in the locus of arc magmatism (e.g., due to a change in dip of the subducting slab), primitive magmas may penetrate to shallow levels through relatively cold and brittle lithosphere, driven by hydrostatic pressure (a similar process may be involved in the eruption of mafic magmas in back-arc environments). However, continued magmatic input leads to crustal softening and trapping of dense melts at depth, leading to further crustal heating and melting. The duration and intensity of the magmatic underplating event will control how far this process goes and may ultimately lead to orogenic crustal thickening events and epochs of explosive felsic volcanism. It is also likely to have an important effect on metallogeny, as discussed below.

## Arc Magma Transport

The mechanics of felsic magma transport through the crust have been the subject of heated debate for over two centuries, since James Hutton first proposed that granites were formed from molten rocks ascending from depth (Hutton, 1788, 1794; Pitcher, 1997). The argument continues today and is polarized largely between those who favor diapiric ascent of magmas (e.g., Singer et al., 1989; Miller and Paterson, 1999), and those who favor ascent along fractures (i.e., dikes; e.g., Clemens and Mawer, 1992; Rubin, 1993; Petford, 1996). The debate is complicated by the fact that the shapes of solidified plutons exposed for inspection record arrival processes and tell us nothing about transport processes en route (Clemens and Mawer, 1992). In fact, they may not even accurately represent the shapes of magma chambers, which are transient features within crystallizing plutons and which expand and contract relative to previously solidified material.

Dike ascent and diapirism are probably both valid mechanisms for magma ascent under different circumstances. Key parameters that will control ascent behavior include magma viscosity and host-rock ductility: in ductile rocks, viscous magmas may rise buoyantly as diapirs, whereas in brittle rocks, even quite viscous magmas will advance by crack propagation (Shaw, 1980; Emerman and Marrett, 1990; Lister and Kerr, 1991; Bergantz and Dawes, 1994; Petford et al., 1994; Vigneresse, 1995a; Weinberg, 1996). This division implies a fundamental change in transport behavior at the brittle-ductile transition in the midcrust, although even this inference is an oversimplification. As is well known to ore deposits geologists who have worked on mesothermal vein deposits, the nature of the brittle-ductile transition is also dependent on strain rate, rheology, and fluid pressure, and there is clear evidence for fracturing of rocks even at mantle depths (Glazner and Ussler, 1988). Thus, the lithosphere below the so-called brittle-ductile transition is not truly ductile but is better described as visco-elastic (e.g., Rubin, 1993). Consequently, such rocks can be expected to fracture under conditions of high strain rate or high fluid (magma) pressure.

Accepting that highly viscous granitic magmas may ascend by diapirism under certain crustal conditions, the ascent of less viscous hydrous magmas such as those generated in deep crustal MASH zones seems most likely to be controlled by fracture propagation. Geophysical modeling of the deep crust beneath the Central Andean arc (Schilling and Partzsch, 2001) and direct observations of exhumed migmatite terranes (Collins and Sawyer, 1996; Brown and Solar, 1999) both point to melt connectivity in fracture networks, and Collins and Sawyer (1996) and Brown and Solar (1999) further suggest that migmatite leucosomes connect to and feed larger conduits or dikes that may ultimately supply shallower level plutons. The role of strain in aiding melt segregation and ascent is discussed separately below.

An extensive literature exists relating to the mechanics of dike propagation (e.g., Shaw, 1980; Lister and Kerr, 1991; Rubin, 1995a) and opinions vary as to its efficacy for large-scale magma transport, from those who consider that felsic magmas will freeze up close to their source (e.g., Rubin, 1995b) to those who suggest that felsic dikes will be self-propagating (e.g., Clemens and Mawer, 1992). Magma viscosity is a key variable in the balance between buoyancy-driven crack propagation and the tendency to solidify due to cooling, because viscosity controls the rate of magma flow: a faster flowing body of magma will convect heat more rapidly and will therefore tend to stay molten. Flow volume is important too, because a narrow dike will lose heat to its wall rocks more quickly per unit volume of melt than a thick dike. Thus, Clemens and Mawer (1992) argued that felsic magmas will ascend efficiently in dikes of ≥3-m width, and Petford et al. (1994) gave estimates of 2 to 10 m for the critical dike width of cordilleran granitoid magmas.

Initial buoyancy-driven expulsion of magma from the source region is aided by the significant volume increase of melting, which may range from 2 to 18 vol percent in rocks containing hydrous minerals, especially muscovite (Clemens and Mawer, 1992; Vigneresse et al., 1996; Rushmer, 2001). As suggested by evidence from migmatite terranes, melt first segregates into a network of small fractures which, with continued melting, coalesce to feed larger magma bodies. The melt fraction required for efficient segregation will depend again on magma viscosity (and also on strain; see below) and has been estimated to range between 20 and 30 vol percent (Wickham, 1987; Vigneresse and Tikoff, 1999), comparable to the 20 vol percent of melt inferred to be present in the crust beneath the Central Andean arc (Schilling and Partzsch, 2001). Once sufficiently high melt fractions and volumes are achieved, dikes will begin to propagate upward. Lister and Kerr (1991) modeled the fluid mechanics of dike propagation and found that magma buoyancy dominates tectonic or hydrostatic forces in driving dike growth. In the absence of freezing, this upward force is balanced by viscous drag in the magma to control growth rate and minimum dike width, whereas maximum dike width is controlled by magma supply rate. Lister and Kerr (1991) also argued that the fracture strength of most rocks is small compared to the available driving forces, such that magmas will create their own fractures in the absence of preexisting planes of weakness.

Magma ascent in a dike will continue until either the magma freezes or its driving force is exhausted or balanced. Cooling will inevitably occur as the magma ascends into colder country rocks but, as noted above, this effect can be offset by high volume flow which continuously convects fresh hot magma into the propagating dike tip. Similarly, repeated emplacement of dikes along a conduit over a short time interval (with respect to cooling rate) will warm the crustal column, thereby aiding the ascent of subsequent magma pulses (cf. Singer et al., 1989). Ascent of magma toward the surface above a sustained source might therefore occur progressively over time as the thermal anomaly is extended upward by repeated dike injection.

If the magma does not freeze, it will continue to ascend until its driving force is lost or balanced. For a buoyancy-driven magma, this will typically occur at its level of neutral buoyancy, as discussed above for basaltic melts. In contrast to basalts, however, the lower density of felsic or hydrous intermediate melts means that this level will be in the upper crust or even at the surface if low-density supracrustal rocks are absent or if the magma vesiculates (Elder, 1978/1979; Walker, 1989).

Magma emplacement and eruption phenomena are discussed further in a later section, but first the effects of tectonic stress and strain on magma ascent are reviewed.

## Tectonic Controls on Magma Ascent

To assess the potential effects of tectonism on arc magmagenesis and transport we return to the MASH zone at the base of the crust, where partial melts must first be separated and assembled into a sufficient volume to initiate ascent. Vigneresse and Tikoff (1999) studied the segregation of magma from partial melting zones and found that shear strain reduces the melt escape threshold and focuses melt accumulation into shear bands (Fig. 5A); see also Sawyer, 1994). These results imply that oblique tectonic stress will enhance segregation of magma from its source region and concentrate it in sites of shear strain. An alternative perspective is that the presence of melt focuses shear strain (D'Lemos et al., 1992; Davidson et al., 1994; Tommasi et al., 1994; Corti et al., 2002).

As reviewed above, theoretical, geophysical, and field evidence suggest that magmas ascend through the crust as dikes. Magma pressure (arising mainly from buoyancy forces but also initially due to volume expansion on melting) will reduce the effective stress on the host rock, but it will not remove any stress differential (Cox et al., 2001). In other words, magma pressure ($P_{magma}$) is subtracted from all of the principal stress components ($\sigma_n$) such that effective normal stress $\sigma_n' = \sigma_n - P_{magma}$; thus, the differential stress $\sigma_1 - \sigma_3 = \sigma_1' - \sigma_3'$ (where $\sigma_1$ and $\sigma_3$ are the maximum and minimum principal stresses, respectively; Fig. 6). Where differential stress is low, increased magma pressure may induce extensional fracturing and dike formation (Davidson et al., 1994), whereas under higher differential stress, shear failure will occur (cf. conditions for hydraulic fracturing; Cox et al., 2001). A dike is merely a

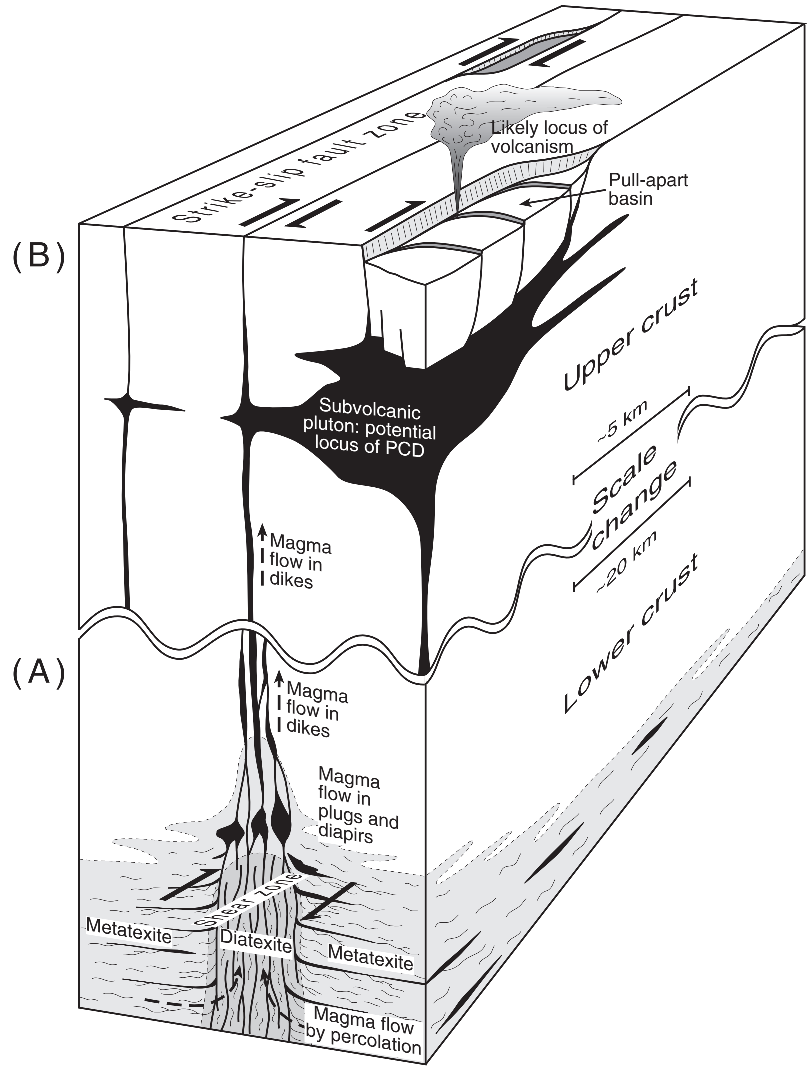


FIG. 5. Schematic cross section of a translithospheric shear zone, inspired by Brown (1994) and Vigneresse and Tikoff (1999). A. Migmatitic (i.e., MASH) zone in the lower crust. The bulk of the region contains partial melt at volumes below the critical melt fraction (metatexite), such that melt migrates by percolation to regions of lower pressure. Under general horizontal compression, magma will tend to accumulate in horizontal lenses or sills (depicted in black). Localized shear strain generates extensional shear bands into which magma is drawn and up which it will begin to rise as buoyant plugs or diapirs (diatexite zone), coalescing upward to form more continuous dikes. B. In the upper crust (note scale change), the shear zone is represented by a set of brittle strike-slip faults, along which jogs or step-overs may give rise to extensional volumes (pull-apart basins at surface). Magma ascent is focused along these structures as dikes and may pool at a shallow level of neutral buoyancy within an extensional zone. Porphyry copper deposits (PCD) may form at this point, and volcanism may occur at the surface.

magma-filled crack in rock and thus obeys the same rules as extensional veins and faults for fracture orientation and opening direction (e.g., Hobbs et al., 1976; Sibson, 2001). In general, fractures open perpendicular to $\sigma_3$ and propagate in the $\sigma_1$–$\sigma_2$ plane, and thus the orientation of dikes can be predicted from local and regional stress fields; conversely, the orientation of solidified dikes can be used to interpret paleostress fields (e.g., Nakamura, 1977; Mériaux and Lister, 2002). In strike-slip and extensional environments, the least principle stress, $\sigma_3$, is in the horizontal plane, and dilational structures (including dikes) will be vertical. Given that buoyancy forces driving magma ascent are also oriented vertically, such structures provide paths of least resistance and highest permeability for magma flow. Under compressional stress with $\sigma_3$ oriented vertically, however, extensional structures will lie in a horizontal plane, favoring the formation of sills (Parsons et al., 1992). Although magma buoyancy may override this constraint (Lister and Kerr, 1991; Paterson and Fowler, 1993), a compressional stress regime is clearly not as favorable for vertical magma flow as a tensional or shear regime and may delay magma ascent until higher degrees of partial melting are achieved (Simakin and Talbot, 2001). In strike-slip environments, extensional domains have a more complex relationship to the major structures that accommodate shear strain. Dilation of shear structures is possible under conditions of low differential stress and high fluid

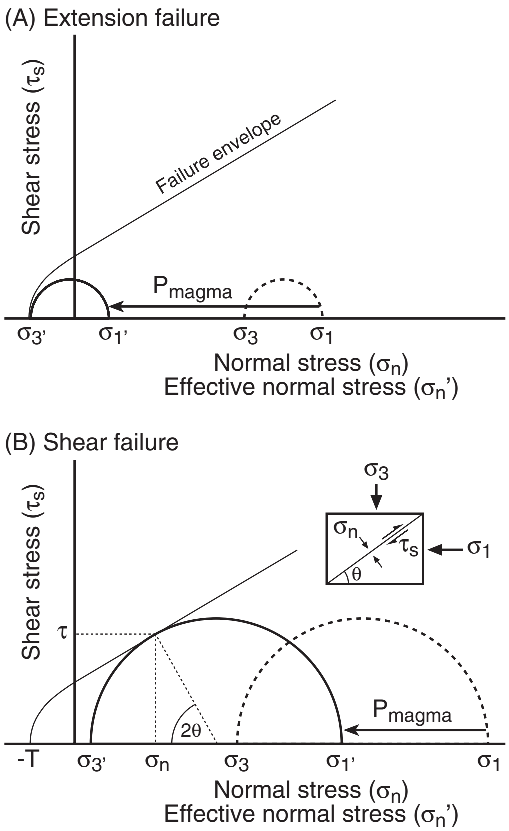


FIG. 6. Mohr circle diagram, showing the effect of fluid (magma) pressure on reducing effective normal stress $\sigma_n' = \sigma_n - P_{magma}$ (modified from Cox et al., 2001). A. Under conditions of low differential stress (small $[\sigma_1 - \sigma_3]$) and high magma pressure, failure will occur by extension to form a magma-filled fracture, or dike. B. Under conditions of higher differential stress (large $[\sigma_1 - \sigma_3]$) and high magma pressure, failure will occur by shear forming a closed fracture along which magma will not necessarily be able to flow. This analysis suggests that conditions of relatively low differential stress are required for efficient magma flow in dikes. T = tensile strength of rock; $\tau$ = shear strength of rock.

(magma) pressure, but extension is more commonly achieved where fault bends, intersections, or step-overs promote fracturing and dilation perpendicular to $\sigma_3$, at approximately 45° to the trend of the shear structure (Fig. 5B).

In arc settings, translithospheric extension or transtension is probably quite rare except during rifting, and at these times magmatism is characterized by mafic volcanism (not prospective for porphyry copper deposits; e.g., Uyeda and Nishiwaki, 1980; Luhr, 1997). Contractional deformation, although likely promoting MASH processes at the base of the crust through crustal thickening, does not facilitate magma ascent into the shallow crust because increased horizontal stresses will oppose the propagation of dikes and promote sill formation (Bussell, 1976; Parsons et al., 1992; Pitcher, 1997; Ida, 1999; Richards, 2000; Richards et al., 2001; Tosdal and Richards, 2001). In contrast, optimum conditions for focused magma ascent are achieved during periods of shear stress, when transpressional deformation provides structurally localized foci for magma ascent and emplacement along extensional conduits at fault intersections and jogs (Fig. 5; Brown, 1994). The literature contains numerous examples of the localization of plutons, volcanoes, and related ore deposits in transpressional settings within arcs (e.g., Bussell, 1976; Aydin et al., 1990; Glazner, 1991; D'Lemos et al., 1992; Tikoff and Teyssier, 1992; Bellier and Sébrier, 1994; Tommasi et al., 1994; Tobisch and Cruden, 1995; Román-Berdiel et al., 1997; Acocella et al., 1999; Brown and Solar, 1999; Benn et al., 2000; García-Palomo et al., 2000; Adiyaman et al., 2001; Gleizes et al., 2001; Hildenbrand et al., 2001; Richards et al., 2001; Chernicoff et al., 2002). Additionally, numerous studies have shown how regional stress fields and resultant crustal strain influence the orientation and structure of volcanic-plutonic systems (Robson and Barr, 1964; Pollard and Muller, 1976; Nakamura, 1977; Weaver et al., 1987; Gudmundsson, 1988; Takada, 1994; Alaniz-Alvarez et al., 1998; Román-Berdiel, 1999; Mériaux and Lister, 2002).

The question of whether existing translithospheric shears promote and focus magma ascent or whether magmatism focuses shear strain and thereby propagates crustal shears is a classic "chicken-and-egg" debate. However, from the point of view of mineral exploration strategies that seek structural vectors to ore deposits, the debate may be side-stepped because both processes produce the same empirical relationship between regional-scale faults and plutons. Paterson and Schmidt (1999) and Schmidt and Paterson (2000) recently argued on the basis of statistics that no such relationship exists. However, in a discussion of these papers, Richards (2001) pointed out that pluton emplacement not only obliterates evidence of precursor structures (initially they become dikes), but also that plutonism is focused in localized regions of extension within or peripheral to broad transpressional fault zones and not necessarily along the strike-slip faults themselves (in much the same way as gold lodes in shear zone-hosted mesothermal deposits occur in second- or third-order structures and not in the first-order shears).

In summary, although conditions of bulk compressional or extensional stress in the lithosphere do not prohibit magma ascent to the upper crust, focused ascent of fertile MASH zone magmas, assumed to be a prerequisite for subsequent porphyry Cu deposit formation, is best achieved under conditions of mild shear stress. Transpressional strain produces vertical, extensional volumes (pull-aparts) at localized discontinuities on strike-slip fault systems, which can channel the ascent and pooling of magma in the upper crust.

## Arc Magma Emplacement

The preceding sections have considered processes affecting mantle and deep-crustal magmagenesis and upward transport of those magmas through the crust. In the following sections, factors that control magma emplacement and eruption are discussed.

The level of neutral buoyancy concept provides one explanation for the upper crustal emplacement of felsic to intermediate composition plutons of dimensions ranging from small stocks to batholiths, because the level of emplacement is independent of magma volume and dependent only on the relative local density of magma and crust (Ryan, 1987). If magma is continually fed to this level, and if the country rocks are not impermeable and rigid such that they can support a hydrostatic head, magmatic overpressure will deform or fracture the conduit walls to form sills or laterally extending bladed dikes (Walker, 1989; Lister and Kerr, 1991).

Even where felsic magmas have positive buoyancy relative to the upper crust, they may nevertheless stall beneath the surface for other reasons, including reduction of magma pressure or supply, intersection of a rheologically strong horizon that acts as a physical barrier to magma ascent, or the increasing viscosity of the magma as it cools to near-solidus temperatures (Clemens and Mawer, 1992).

A key factor in sustaining upper crustal magmatism is the magma supply rate. If the supply dwindles, then the flux of heat required to maintain the flow of magma in dikes will diminish, and the magma will freeze up (Clemens and Mawer, 1992). However, if the supply rate is maintained, an upper crustal intrusive complex of batholithic dimensions can be constructed in a remarkably short space of time. For example, Petford (1996) estimated that for realistic magma ascent rates in dikes of ~$10^{-2}$ m/s, large felsic plutons can be filled on a time scale of <$10^4$ yr. Similarly, Paterson and Tobisch (1992, p. 291) allowed "no more than a few million years" for the same process. Such figures are supported by field and geochronological studies of batholithic terranes, which commonly indicate rapid batholith assembly. For example, the La Posta suite granodiorites of the Peninsular Ranges batholith in southern and Baja California, which crop out over an area of 15,000 km$^2$, were apparently emplaced over a time interval of ≤7 m.y. between 99 and 92 Ma (Kimbrough et al., 2001).

Various mechanisms have been proposed for the upper crustal emplacement of plutons that endeavor to address the problem of space for intrusion. Many arc plutons appear to be tabular in shape, with space being created either by floor depression (lopoliths) or roof lifting (laccoliths; Cruden, 1998; Vigneresse et al., 1999; de Saint-Blanquat et al., 2001; Acocella and Rossetti, 2002). Magmas are envisaged to spread laterally at their level of neutral buoyancy or beneath a horizontal rheological barrier by vertically displacing the country rocks. In contrast, horizontal displacements of the host rocks, particularly in the upper brittle crust, are limited by their capacity for elastic strain or the rate of tectonic extensional strain (Paterson and Tobisch, 1992; Johnson et al., 2001; Acocella and Rossetti, 2002). Consequently, plutons in arc terranes (including porphyry Cu-forming plutons) are commonly intruded at the base of a coeval volcanic pile or at the basement-supracrustal contact, because the overlying volcanic and sedimentary rocks are typically weak and of lower density than andesitic-dacitic magmas.

## Arc Volcanism

Sillitoe (1973) first suggested that porphyry Cu deposits might be overlain by composite volcanoes at the time of formation (Fig. 7). Although this is very likely the case, there is no actual requirement in currently accepted models of porphyry metallogenesis for volcanism to play a critical role in magmatic-hydrothermal ore formation (although volcano sector collapse has been proposed as a trigger for volatile exsolution through sudden depressurization; Sillitoe, 1994). Indeed,

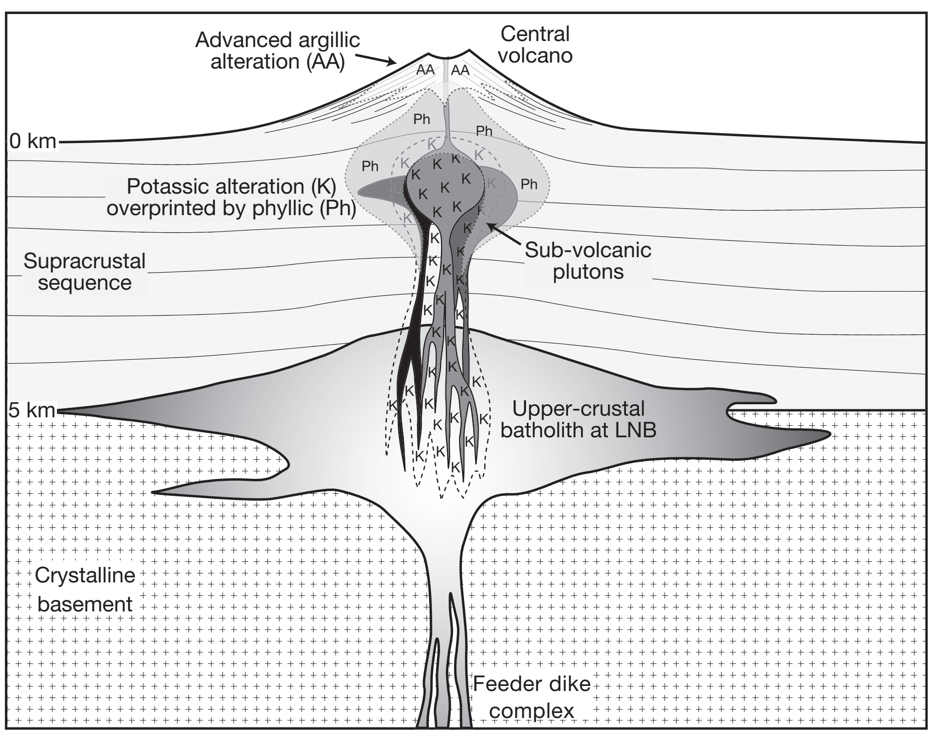


FIG. 7. Schematic cross section through a porphyry Cu-forming volcano-plutonic system. An upper crustal batholith complex of andesitic composition is fed by dikes rising from a lower crustal MASH zone. After further fractionation at this level, evolved, volatile-rich dacitic magmas are emplaced at shallow levels and may vent to the surface to build a volcanic edifice. Volatiles exsolved from the large volume of crystallizing batholithic magma are channeled upward along the subvolcanic structural axis of the system and generate magmatic-hydrothermal potassic alteration (K), potentially with Cu mineralization. As the magmatic-hydrothermal system wanes, phyllic alteration (Ph) overprints the peripheral potassic alteration and advanced-argillic alteration (AA; fumarolic alteration at surface) affects the volcanic edifice. Propylitic alteration (not shown) caused by circulating heated ground waters affects the country rocks in a wide zone around the system.

Pasteris (1996) has suggested that loss of key volatiles such as sulfur during major volcanic eruptions, as in the case of Mount Pinatubo in 1991, might short-circuit the porphyry-forming process and yield negative porphyry systems. Nevertheless, some amount of volcanism is an almost inescapable consequence of intruding large volumes of magma into the upper crust, and eruptions are likely to accompany, if not be integrally related to or caused by, volatile exsolution processes. A brief consideration of the causes of volcanism is therefore warranted, particularly in the light of the previous discussion of buoyant magma ascent, which might suggest that few magmas should erupt at the surface except under special conditions of hydrostatic overpressure.

In arc environments, the most common direct cause of volcanism is magmatic vesiculation, which can quickly reduce the density of the magma to below that of unconsolidated supracrustal materials, while at the same time greatly increasing its volume (and therefore magma chamber pressure; Walker, 1989; Jaupart and Allègre, 1991; Carrigan et al., 1992). Volatile exsolution during explosive volcanism can increase the magma volume by up to 99 percent, resulting in complete fragmentation of the magma and its ejection as a high-velocity Plinian eruption column (Sparks and Wilson, 1976; Gardner et al., 1996). Another mechanism for eruption is the displacement of magma from a shallow-level chamber by recharge with fresh magma from depth (Ryan, 1987; Eichelberger, 1995). If the chamber and conduit walls are sufficiently strong such that they can support hydrostatic pressure, lava will be extruded at the surface. Additionally, these two processes may be combined, with recharge by hot, volatile-rich magma triggering explosive degassing in the resident magma (e.g., Eichelberger, 1995; Hattori and Keith, 2001). Other external processes that facilitate or promote volcanism over plutonism include tectonic and magmatic stresses that may fracture the overlying crust or dilate existing structures (Ryan, 1987; Clemens and Mawer, 1992; Gudmundsson, 1998) and changes in hydrologic or lithologic load due to meteorological or physical process such as sector collapse (Voight et al., 1981; Sillitoe, 1994).

Magmatic volatile exsolution is an essential step in the formation of porphyry Cu deposits and is an inevitable result of the shallow-level crystallization and cooling of hydrous arc magmas (Whitney, 1975; Burnham, 1979; Eichelberger, 1995). However, it is important that the process does not catastrophically vent the volatiles essential for hydrothermal ore formation. Explosive vesiculation and eruption most commonly occur in viscous felsic magmas because gas bubbles cannot separate quickly enough from the melt; thus, ignimbrite-forming eruptions are typically generated from high-silica dacitic and rhyolitic magmas. In contrast, lower viscosity intermediate-composition magmas such as andesites and lower silica dacites are able to degas more readily, with passive dispersion of volatiles through the volcanic edifice to vent as fumaroles or to condense into ground water. Hedenquist et al. (1998) suggested that hypogene advanced argillic alteration found in the upper parts of some porphyry systems represents this shallow-level degassing (Fig. 7), with the corollary that fumarolic alteration provides a surface indication of deeper seated magmatic-hydrothermal activity and potential porphyry-type ore formation.

## Tectono-Magmatic Cycles and Porphyry Cu Deposit Formation

A spatial and temporal relationship between tectono-magmatic cycles in arcs and porphyry Cu formation has long been recognized (e.g., Sillitoe, 1972). In addition, a close spatial relationship to major arc-parallel transcurrent faults is evident in many porphyry provinces, the best known example being the West Fissure zone of northern Chile, which hosts several of the world's largest deposits (Collahuasi, Chuquicamata, La Escondida, El Salvador-Potrerillos; Fig. 8). Clark et al. (1976) and Sillitoe (1981, 1988) showed that porphyry deposits in this region occur within several linear belts of coeval Cenozoic magmatism, corresponding to the positions of the magmatic arc in the Paleocene-early Eocene (Central Valley belt), late Eocene-early Oligocene (West Fissure zone), and early-middle Miocene (El Indio and Maricunga belts). Similar broad spatial-temporal relationships are noted in Perú and Colombia (Sillitoe, 1972, 1988; McKee and Noble, 1989), Mexico (Clark et al., 1982; Damon et al., 1983; Barton et al., 1995), the North American Cordillera (Titley and Beane, 1981; Barton, 1996), the southwest Pacific (Titley, 1981; Titley and Beane, 1981), eastern Australia (Horton, 1978), the Tethyan belt of Turkey-Iran-Pakistan (Waterman and Hamilton, 1975; Glennie, 2000), and Siberia and Mongolia (Berzina et al., 1999). These relationships are summarized in Table 1.

At a superficial level, this tectono-magmatic association might be taken merely to indicate that porphyry Cu deposits are linked to arc magmatism and therefore that one should explore in any partially eroded arc. But magmatic arcs extend over very large areas, and porphyry deposits represent very small point features within those arcs. Is it possible to predict more accurately the timing and magmatic association of deposits within this overall tectono-magmatic framework and perhaps even to predict the location of potentially ore-forming magmatic systems within a given belt?

Maksaev and Zentilli (1988), McKee and Noble (1989), Hammerschmidt et al. (1992), Cornejo et al. (1997), and Richards et al. (2001) have demonstrated that major porphyry Cu deposits in the Peruvian-Chilean belt are formed late within a given magmatic cycle, the porphyry intrusions typically representing the last intrusive event in a given area (Table 1). For example, Richards et al. (2001) have shown that the Escondida porphyry deposit is one of three mineralized dacitic centers within a large, coeval (~38 Ma), cogenetic, shallow-level, dioritic plutonic complex (Fig. 2) that was emplaced at the end of a protracted period of Eocene andesitic volcanism. In many cases there is also a clear spatial relationship between the mineralized centers and major transcurrent fault zones and in particular to intersections of transverse lineaments with these structures (Fig. 8; Salfity, 1985; Lindsay et al., 1995; Richards, 2000; Richards et al., 2001). It is suggested that this relationship can be understood within the context of the preceding discussion of arc tectono-magmatic processes, as follows.

During a period of stable subduction, in which the slab dips beneath the arc at a constant angle and subducts with a constant velocity, slab dehydration and magma generation in the mantle wedge occur in a relatively narrow band at a depth of around 100 km and at a fixed distance from the trench. Under

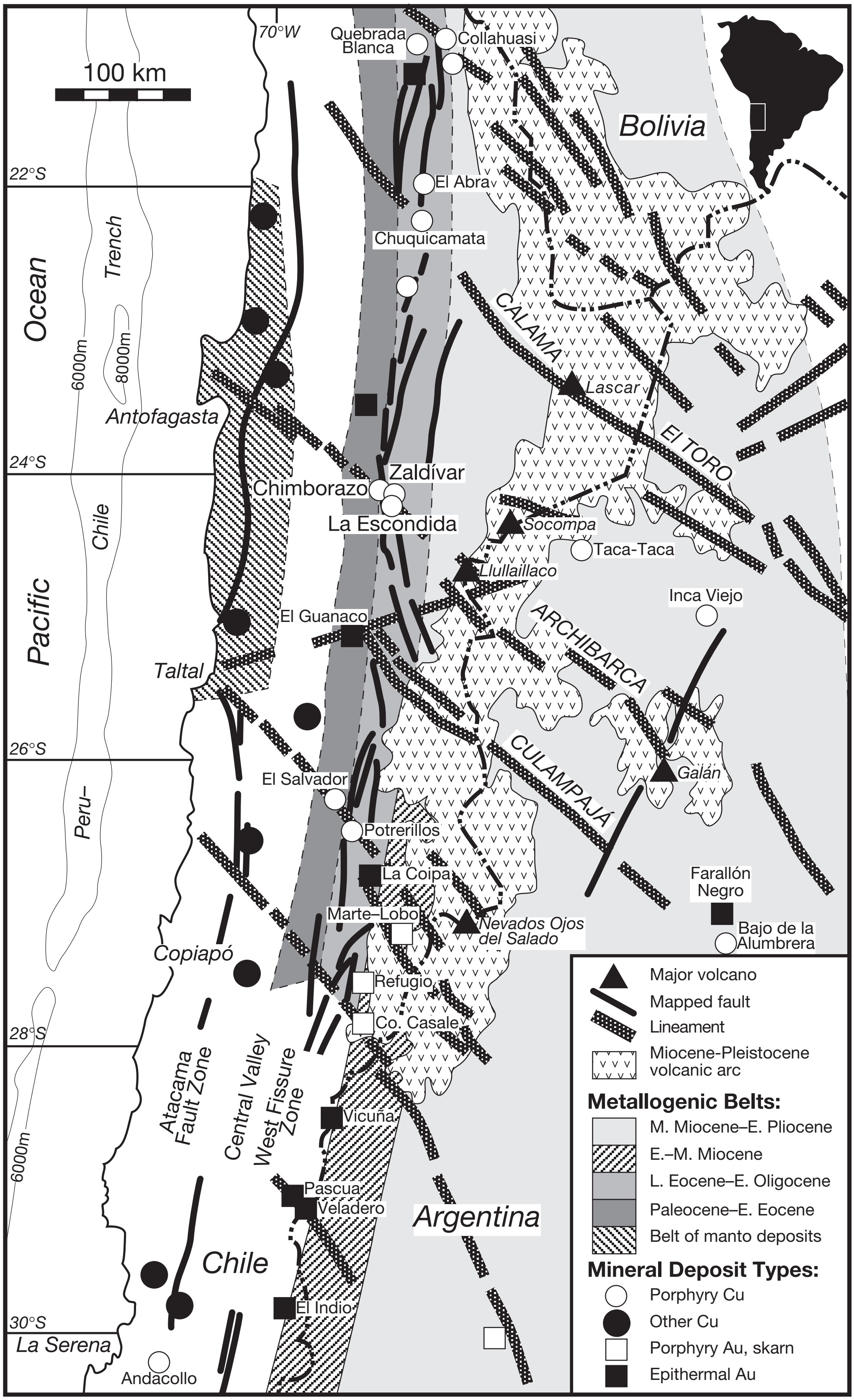


FIG. 8. Geologic sketch map of northern Chile, showing the locations of major Cu and Au deposits in relation to arc-parallel belts of coeval magmatism (after Sillitoe, 1992) and regional-scale faults and lineaments (modified from Salfity, 1985; Salfity and Gorustovich, 1998). Figure modified from Richards et al. (2001).

TABLE 1. Relationships between Tectonic, Magmatic, and Porphyry Cu Deposit (PCD) Cycles in Selected Volcanic Arcs

| Region | Tectonic cycles | Volcanic cycles | Major porphyry cycles | Relationship of PCD to tectonic cycles | Spatial relationship to regional structures | References |
|---|---|---|---|---|---|---|
| Northern Chile (Eocene-Oligocene porphyry Cu belt) | 44–38 Ma Incaic orogeny (diachronous); shortening related to period of rapid convergence | 48–38.5 Ma Precordillera magmatic arc | 42–32 Ma | PCD formed during late- or post-orogenic intrusive phase of Precordillera magmatic arc, prior to Oligo-Miocene shallowing of subduction angle and eastward shift of volcanism | Strong relationship to orogen-parallel Domeyko fault system (West Fissure zone), and secondary relationship to intersections with transverse lineaments | Maksaev and Zentilli, 1988; Hammerschmidt et al., 1992; Scheuber and Reutter, 1992; Tomlinson and Blanco, 1997; Richards et al., 2001 |
| Central Chile (Miocene Maricunga belt) | 20–17 Ma contraction and crustal thickening, followed by uplift and extension | 26–21 Ma volcanism with epithermal Au-Ag; 16–12 Ma Maricunga-Cadillal volcanic group | 13–12 Ma (porphyry Au deposits: Marte, Lobo) | Gold-rich PCD formed during stress relaxation at end of compressional tectono-magmatic cycle, followed by shallowing of subduction angle and eastward shift of volcanism | Orogen-parallel magmatic belt and N-S to NNE-SSW thrust faults; intersection with NW-SE fault sets controlled loci of mineralization | Vila and Sillitoe, 1991; Mpodozis et al., 1995; Kay et al., 1994, 1999; Muntean and Einaudi, 2000 |
| Central Chile (Miocene El Teniente belt) | 19–16 Ma contraction and crustal thickening | 15–7 Ma Teniente volcanic complex; volcanic phase ends 7 Ma | 7–5 Ma | PCD and breccia-related pipes formed at the end of the tectono-magmatic cycle; accompanied by uplift and eastward shift of volcanism over shallowing slab | Orogen-parallel alignment of intrusions, breccias, and N-S faults along 80 km axis suggests basement structural control | Camus, 1975; Skewes and Stern, 1995; Kurtz et al., 1997; Kay et al., 1999 |
| Perú (Eocene) | 84–79 Peruvian orogeny; 59–55 Incaic I orogeny | 75–59 Toquepala volcanics | 57–52 Ma | Large PCD form at the end of major tectono-magmatic cycle, followed by shallowing of subduction angle | Toquepala PCD located at intersection of major NW-SE Incapuquio fault system and NNE-trending faults of the Toquepala lineament | Clark et al., 1990; Sandeman et al., 1995; Zweng and Clark, 1995; Benevides-Cáceres, 1999 |
| Perú (Neogene) | 19–17 Ma Quechua I orogeny; smaller transpressional pulses at 10–9 Ma, 7–5 Ma, 2.5 Ma[1] | Major volcanic episodes from 22–17 Ma, 16–8 Ma, 7–0 Ma | 20–18 Ma, 15.5–7 (peaks at 15.5–13 Ma, 10–7 Ma) | Small PCD form at the end of relatively short-lived tectono-magmatic cycles; followed by shift in axis of volcanism | PCD restricted to orogen-parallel and transverse (NE-SW) magmatic belts, defined by intersecting fault sets | McKee and Noble, 1989; Clark et al., 1990; Noble and McKee, 1999; Benevides-Cáceres, 1999; Petersen, 1999 |
| Central Iran | Miocene-Pliocene (diachronous) closure of Neo-Tethys between Eurasia/Arabia | Oligo-Miocene Urumieh-Dokhtar volcanic arc | Middle Miocene (~12 Ma) | PCD emplaced at end of magmatic cycle prior to continental collision | Orogen-parallel structures bounding Urumieh-Dokhtar belt reactivated as Pliocene-Quaternary dextral strike-slip faults | Alavi, 1994; Bushara, 1996; Glennie, 2000 |
| Southwestern North America (Mesozoic) | Sevier orogeny (Late Jurassic-Paleocene); flattening of Farallon plate subduction | 105–89 Ma eastward migration of magmatism from Peninsula Range batholith | 110–95 Ma (Ruth) 85–70 Ma (Buckingham, Bagdad) | PCD emplaced at end of magmatic cycles | Emplacement into extensional structures within regional compressional stress field | McCandless and Ruiz, 1993; Albino, 1995 |
| Southwestern North America (Cenozoic) | 80–40 Ma Laramide orogeny: rapid normal convergence, crustal thickening | 80–50 Ma eastward expansion of volcanism | 60–55 Ma (Copper Creek, Ray, Morenci) 40–38 Ma (Bingham, Battle Mt.) | PCD emplaced at end of magmatic cycles | Possible relationship to E-ENE- and NW-trending basement structural fabric, reactivated during Laramide orogeny; regional lineament control at Bingham | McCandless and Ruiz, 1993; Albino, 1995; Titley, 1981, 1995 |

[1] There is disagreement between Benevides-Cáceres (1999) and Noble and McKee (1999) over the timing of the Quechua orogenic pulses; the dates of Noble and McKee (1999) are reported here to be consistent with the dates of porphyry systems; (Benevides-Cáceres, 1999: 17 Ma Quechua I orogeny; smaller pulses at 8–7 Ma, 5–4 Ma, 2–1.6 Ma.)

broadly compressional stress, primitive arc magmas pool at the base of the crust of the overriding plate and begin to develop a relatively narrow but linearly extensive MASH zone along the length of the proto-arc. Sufficient partial melting and homogenization to form lower density, intermediate-composition magmas probably occurs within a few million years of MASH zone initiation, and these magmas will then begin to rise to shallower levels of the crust. Advection of heat into the crust along this axis causes thermal weakening, and shear strain may either be focused along favorably oriented preexisting structures or may initiate new structures. Magma ascent is channeled along these zones of structural weakness, providing a feedback effect and resulting in the restriction of magmatism to narrow, arc-parallel tectonic belts.

The late timing within the magmatic cycle and the relatively evolved (dacitic) composition of many porphyry systems (Anthony and Titley, 1988; Richards et al., 2001) suggest that extended evolution of the MASH zone is required before sufficient volumes of magma of adequate fertility (i.e., containing high levels of oxidized sulfur, metal, and water) to generate giant porphyry deposits can be produced. Moreover, these fertile magmas need to be intruded in bulk into the upper crust without excessive eruption, such that further shallow-level evolution and volatile exsolution can take place.

Voluminous shallow-crustal emplacement of fertile magmas at a late stage in the magmatic cycle is optimized under conditions of mild transpressional stress. Some degree of horizontal compressional stress is required to prevent short-circuiting of the translithospheric magmatic system and the eruption of mafic magmas along rift faults (as in extensional arcs or back-arcs), but excessive compression as observed during orogenic episodes is likely to favor entrapment of magmas in sill complexes at the base of the crust. Moderate transpressional stress promotes the formation of localized, vertically connected extensional zones (pull-aparts) at fault intersections, fault bends, or step-overs, which offer high-permeability pathways for the focused ascent of magma from lower crustal MASH zones. Voluminous ponding of these magmas in upper crustal chambers under conditions of low deviatoric stress permits further magmatic differentiation and bulk volatile saturation and exsolution, with the concomitant opportunity to form large porphyry-type hydrothermal ore deposits.

## Concluding Remarks

The genesis of porphyry Cu deposits cannot be viewed in isolation from the tectono-magmatic origins of their source magmas. In this paper I have attempted to review the current state of understanding of the various processes that comprise tectono-magmatism in (principally continental) subduction-related arcs, and I have shown that the emplacement of potential porphyry Cu-forming magmas can be understood within the context of these processes. The broad uniformity of porphyry Cu deposits worldwide suggests that their mechanism of formation must be quite straightforward and reproducible, requiring no unique processes or magma types (e.g., Cline and Bodnar, 1991). Variations in grade and size are likely to be a function of the convergence of various contributing processes and their cumulative effects. The default product is likely to be a barren or weakly mineralized magmatic system, but where processes converge optimally, large ore deposits may form.

To some extent, this convergence can be understood and predicted. For example, it seems logical that large magmatic-hydrothermal deposits will be formed where large supplies of magma are available (whether currently exposed or not), which in turn implies a large, long-lived, focused, tectono-magmatic event. Such events can be sought in the geological record of a given prospective terrane, and the locations of focused magma ascent can be predicted in terms of crustal structural architecture. Magma may be emplaced and porphyry deposits may form anywhere along the arc, but large deposits will most likely be formed where magma ascent is concentrated by this structural framework and particularly where structural intersections provide the opportunity for the formation of vertically extensive (transcrustal) dilational conduits. The prevailing crustal stress regime at the time of magma supply will dictate whether such structures are indeed dilational and will therefore serve as magma conduits. Indeed, magma ascent and formation of large porphyry deposits may be restricted to periods when such structural zones are extensional. Thus, there is a logical link between tectonic and magmatic cycles in arcs and porphyry formation: during major tectonic cycles, horizontal compression throughout the lithosphere hinders the upward ascent of magma and favors pooling in deep-crustal sill complexes where magmas evolve and interact with lower crustal materials. The termination of compressional orogenic cycles is related to changes in plate motion (convergence direction, speed) or angle of subduction, and porphyry Cu intrusions are commonly the last major magmatic event in the arc prior to shifting of the locus of magmatism to a new position above the relocated Benioff zone. As horizontal stress relaxes at the end of the orogenic period and shear strain is partitioned into strike-slip fault movement, buoyant, evolved magmas rise along structurally controlled extensional pathways to the upper crust, where they may again pool to form batholiths at their level of neutral buoyancy. Further magmatic fractionation, emplacement of shallow-level apophyses, and volatile exsolution may generate late-stage porphyry Cu deposits.

These considerations explain why certain age-related magmatic belts are prospective for large porphyry deposits, and others, which lack key parameters such as longevity or structural focus, are not. For example, the well-developed late Eocene-early Oligocene plutonic arc in northern Chile hosts several of the world's largest porphyry Cu deposits, but the less well-developed Paleocene-early Eocene arc, which is dominated by volcanic rocks, hosts fewer and generally smaller deposits. Within prospective belts, an understanding of regional structural patterns and their dynamic histories may be helpful in predicting specific loci of maximum magmatic flux and therefore maximum potential for ore formation. However, as Chernicoff et al. (2002) have pointed out, translithospheric structures are commonly represented by zones of discrete faults, often 30 to 50 km wide, and individual fault intersections at the surface cannot be expected to reflect the detailed structure of the base of the crust where magma ascent begins. Thus, the explorationist still has plenty of fieldwork to do in order to pinpoint a 1-km$^2$ porphyry orebody within the zone of influence of a translithospheric structural intersection.

In this review, I have not attempted to consider late-stage magmatic and hydrothermal processes that will exercise the final controls on whether or not an economic porphyry deposit is formed (see Richards, 2004). Such processes are likely to be deposit specific and are less predictable from the point of view of regional exploration. They include the depth of magma emplacement, its specific volatile content and oxidation state, the eruption history, and the history of magma recharge. Furthermore, postmineralization processes such as uplift and weathering may completely obliterate a deposit or may turn it into a giant like La Escondida (Alpers and Brimhall, 1989).

## Acknowledgments

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