Porphyry copper deposit formation in arcs: What are the odds?

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ABSTRACT

Arc magmas globally are H₂O-Cl-S–rich and moderately oxidized (ΔFMQ = +1 to +2) relative to most other mantle-derived magmas (ΔFMQ ≤ 0). Their relatively high oxidation state limits the extent to which sulfide phases separate from the magma, which would otherwise tend to deplete the melt in chalcophile elements such as Cu (highly siderophile elements such as Au and especially platinum-group elements are depleted by even small amounts of sulfide segregation). As these magmas rise into the crust and begin to crystallize, they will reach volatile saturation, and a hydrous, saline, S-rich, moderately oxidized fluid is released, into which chalcophile and any remaining siderophile metals (as well as many other water-soluble elements) will strongly partition. This magmatic-hydrothermal fluid phase has the potential to form ore deposits (most commonly porphyry Cu ± Mo ± Au deposits) if its metal load is precipitated in economic concentrations, but there are many steps along the way that must be successfully negotiated before this can occur. This paper seeks to identify the main steps along the path from magma genesis to hydrothermal mineral precipitation that affect the chances of forming an ore deposit (defined as an economically mineable resource) and attempts to estimate the probability of achieving each step. The cumulative probability of forming a large porphyry Cu deposit at any given time in an arc magmatic system (i.e., a single batholith-linked volcano-plutonic complex) is estimated to be ~0.001%. Continued uplift and erosion in active convergent tectonic regimes rapidly remove these upper-crustal deposits from the geological record, such that the probability of finding them in older arc systems decreases further with age, to the point that porphyry Cu deposits are almost nonexistent in Precambrian rocks.

A key finding of this paper is that most volcano-plutonic arcs above subduction zones are prospective for porphyry ore formation, with probabilities only falling to low values at late stages of magmatic-hydrothermal fluid exsolution, focusing, and metal deposition. This is in part because of the high threshold required in terms of grade and tonnage for a deposit to be considered economic. Thus, the probability of forming a porphyry-type system in any given arc segment is relatively high, but the probability that it will be a large economic deposit is low, dictated to a large extent by mineral economics and metal prices.

INTRODUCTION

Convergent margins are sites of subduction of oceanic lithosphere, during which some constituents of the continental lithosphere and volatiles from the hydrosphere are recycled into the mantle (Fig. 1). Most of this material, including up to 11% of the volatiles (Kimura and Nakajima, 2014), eventually sinks deep into the mantle and may reappear at Earth’s surface later as components of mantle plumes (Hofmann and White, 1982; van der Hilst et al., 1997; Zhao, 2004). However, the bulk of the volatiles, along with fluid-soluble components such as large-ion lithophile elements, are released at shallow mantle depths from the subducted material, where they metasomatize the overlying asthenospheric mantle wedge and/or the base of the lithosphere (Fig. 1; Ringwood, 1977; Tatsumi, 1986; Peacock, 1993; Wallace, 2005). Hydrous metasomatism lowers the solids temperature of peridotite, leading to partial melting in the hottest central regions of the upwelling asthenospheric mantle wedge (Kushiro et al., 1968; Grove et al., 2006; England and Katz, 2010; Kelley et al., 2010; Green et al., 2014). Derivative basaltic magmas are uniquely water-rich (plus other volatiles including S and Cl) and evolve to hydrous anesitic compositions (≥4 wt% H₂O) during deep crustal fractional crystallization and assimilation processes (Hildreth and Moorbath, 1988; Grove et al., 2003; Pichavant and Macdonald, 2007; Zimmer et al., 2010; Lee et al., 2014; Chin et al., 2018). Upon ascent of this lower-density, evolved melt into the upper crust, further crystalization and depressurization result in exsolution of dissolved volatiles, which may either be vented at surface in volcanic eruptions and fumarolic emissions or may form subsurface hydrothermal systems (Fig. 1; Burnham, 1979; Hedenquist and Lowenstern, 1994; Tassi et al., 2009). Metals dissolved in the magma preferentially partition into these high-temperature, saline, sulfurous magmatic fluids, and efficient subsurface channeling and depositional processes can result in their precipitation in economic concentrations to form ore deposits (e.g., porphyry, epithermal, and skarn deposits enriched in Cu, Mo, Au, and other elements, hereafter referred to generally as “porphyry Cu deposits” unless the subject is a specific deposit type) (Burnham, 1979; Candela and Holland, 1984; Hedenquist and Lowenstern, 1994; Kouzmanov and Pokrovski, 2012). Porphyry Cu deposits are characterized by concentrations

†Deceased
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of Cu-Fe-sulfide minerals (mainly chalcopyrite, CuFeS₂), along with pyrite (FeS₂) and variable quantities of molybdenite (MoS₂), precipitated in quartz veins and as disseminations in hydrothermally altered wall rocks. Alteration associated with the primary (hypogene) mineralization is typically “potassic,” consisting of an assemblage of hydrothermal quartz, K-feldspar, biotite, magnetite, and anhydrite. Reviews of the key characteristics and variants of porphyry Cu deposits are provided by Lowell and Guilbert (1970), Gustafson and Hunt (1975), Cooke et al. (2005), Seedorff et al. (2005), John et al. (2010), and Sillitoe (2010). See Meinert et al. (2005) and Simmons et al. (2005) for reviews of skarn and epithermal deposits.

This paper reviews the formation of porphyry Cu deposits in subduction environments from a probabilistic perspective, recognizing that they are reproducible but rare events in the evolution of arc magmatic systems. The approach uses estimates of probability of various processes that affect arc magmas as they ascend from the mantle wedge source to the shallow crust. These values are not rigorous assessments, but educated guesses based on the author’s experience and assessment of the literature. These estimates attempt to reflect the cumulative probability of a large economic porphyry Cu deposit forming in association with a specific arc batholith, implying a specific location (within an area of perhaps 100 km²) and at a specific point in time (within a few million years) in the evolution of an arc. The probabilities are highest in the Cenozoic but decrease with geological age, due either to the increasing probability of loss to erosion, or to the possible non-formation of porphyry Cu deposits prior to the onset of “modern” plate tectonics.

■ OVERVIEW OF CONVERGENT MARGIN METALLOGENY

The distinctive metallogeny of convergent plate margins arises primarily due to the recycling of water and other volatiles into the mantle by subduction of altered oceanic lithosphere. The return of these volatiles toward the surface via hydrous magmas results either in explosive arc volcanism and/or the formation of shallow, subsurface hydrothermal systems that can form ore deposits. However, although all arc volcanoplutonic complexes are accompanied by some hydrothermal and fumarolic alteration, mineralization is not always present, and economic deposits (ore deposits) are rare (perhaps 1 in 1000 prospected systems actually prove to be economic for mining). Many factors contribute to the ultimate formation of an economic porphyry (or related epithermal or skarn) deposit, and omission or inefficient operation of any one of these steps can reduce the ore-forming potential (fertility) of the system (Richards, 2013). Thus the concept of magma fertility needs to be considered separately from ore-forming processes; whereas...
fertility can be evaluated and predicted to a fairly high degree of certainty, the actual formation of an ore deposit, even from a fertile system, typically cannot. Ore formation in fertile systems is thus a stochastic event, with a low but nonzero probability of occurrence.

The probability of ore formation is also a function of human valuation, because this defines what is considered to be “ore” (i.e., material from which metals can be profitably extracted). The value of a commodity is controlled by its availability and usefulness: a commodity in high demand but low availability will command a high price, with the result that smaller and/or lower-grade (more common) deposits will be economic. In contrast, a widely available commodity will have a lower price, and only the largest and richest (rare) deposits will be economic. However, there is a feedback loop, because if the largest and richest deposits become too rare, then availability declines, and prices will go up, making smaller or lower-grade deposits economic.

Porphyry deposits are characteristically enriched in Cu and/or Mo and/or Au, all of which are relatively rare and valuable metals. Consequently, even quite low-grade deposits can be economic, and porphyries are commonly referred to as large-tonnage, high-grade deposits. Typical hypogene grades (of primary mineralization, unaffected by supergene enrichment processes that occur during weathering) are in the range 0.5%–1.5% Cu, <0.01%–0.04% Mo, and 0.0–1.5 g/t (ppm) Au, with tonnages of ore commonly measured in the hundreds of millions to billions of tonnes (Sillitoe, 2010). Most such deposits are mined from large open pits, benefiting from economies of scale and efficient mineral processing infrastructure (e.g., Duggen et al., 2007; Hermann and Spandler, 2008; Miβ et al., 2011; Labanieh et al., 2012; Chen et al., 2013; Spandler and Pirard, 2013; Ribeiro et al., 2015; Schmidt, 2015; Walowski et al., 2015; Spinnell et al., 2016). However, both fluids and melts will transfer volatiles, incompatible elements, and an oxidation signature into the overlying mantle wedge (Kelley and Cottrell, 2009; Alt et al., 2012; Debret et al., 2015; Birner et al., 2017; Riel et al., 2017). Above normally dipping subduction zones (~30°), this metasomatic flux will interact with a wedge of hot asthenospheric mantle between the downgoing slab and the upper plate at depths of ~100 km, whereas during shallower subduction, this flux may directly impinge on the base of the overlying lithosphere (discussed below).

In normally dipping subduction zones, partial melting in the high-temperature core of the metasomatized asthenospheric mantle wedge will

### STEP 1: FERTILE ARC MAGMA COMPOSITIONS

Mid-ocean ridge basalt (MORB) and nascent arc magmas (boninites and tholeiitic basalts; formed prior to extensive subduction metasomatism) are both derived from partial melting of depleted upper mantle from which a high proportion of crust-forming lithophile elements has already been removed (Haraguchi and Ishii, 2007; Kodaïra et al., 2010; Escuder-Viruete et al., 2014; Whattam and Stern, 2016). Such magmas are not typically directly prospective for ore formation. However, during subduction, the wedge of mantle asthenosphere above the downgoing oceanic lithosphere becomes metasomatized by fluids and melts derived from prograde metamorphism of the slab (Ringwood, 1977; Tatsumi, 1986; Peacock, 1993). In addition to water (Tatsumi et al., 1986; Stolper and Newman, 1994; Portnyagin et al., 2007), these fluids enrich the mantle wedge in S (de Hoog et al., 2001; Wallace, 2005; Tomkins and Evans, 2015), Cl (Kent et al., 2002; Wallace, 2005; Portnyagin et al., 2007), large ion lithophile elements (K, Rb, Cs, Ca, Sr, and Ba), metals (U, Pb, and Cu), semi-metals (As and Sb), and other fluid-mobile elements such as B, Ti, and Si (Noll et al., 1996; Kogiso et al., 1997; Hattori and Guillot, 2003; Breeding et al., 2004; Manning, 1995, 2005; Hattori et al., 2005; Wysocki et al., 2006). These components then become incorporated into primary arc magmas, which are enriched relative to MORB in H₂O (~1–5 wt%, locally up to 8 wt%), Cl (500–2000 ppm), CO₂ (~3500–7600 ppm), and S (900–2600 ppm) (Sobolev and Chausssidou, 1996; de Hoog et al., 2001, 2014; Fischer and Marty, 2005; Wallace, 2005; Kimura and Ariskin, 2014; Kamenetsky et al., 2017).

There remains some debate about the relative roles of hydrous fluids versus partial melts of seafloor sediments and/or basaltic oceanic crust (e.g., Duggen et al., 2007; Hermann and Spandler, 2008; Miβ et al., 2011; Labanieh et al., 2012; Chen et al., 2013; Spandler and Pirard, 2013; Ribeiro et al., 2015; Schmidt, 2015; Walowski et al., 2015; Spinnell et al., 2016). However, both fluids and melts will transfer volatiles, incompatible elements, and an oxidation signature into the overlying mantle wedge (Kelley and Cottrell, 2009; Alt et al., 2012; Debret et al., 2015; Birner et al., 2017; Riel et al., 2017). Above normally dipping subduction zones (~30°), this metasomatic flux will interact with a wedge of hot asthenospheric mantle between the downgoing slab and the upper plate at depths of ~100 km, whereas during shallower subduction, this flux may directly impinge on the base of the overlying lithosphere (discussed below).
2. Long-lived, moderately steep subduction, leading to focused magmatic arc in upper plate

3. Detention of magma in thickened, intermediate to felsic upper plate lithosphere: MASH processing

4. Transpressional strain in upper plate lithosphere, leading to focused magma ascent, and rapid formation of large mid–upper crustal batholith

5. Formation of subvolcanic cupola zones in the batholith roof into which the ascent of metal-bearing fluids and bubbly magma is focused

6. Voluminous flow of magmatic-hydrothermal fluids through cupola zone, sustained by continuous magmatic recharge of batholith

7. Efficient precipitation of Cu-Fe, Mo sulfide minerals to form a large porphyry deposit

8. Postmineralization history: uplift and erosion to expose subvolcanic level

9. Large economic porphyry deposit (grade, tonnage, accessibility, and infrastructure sufficient for profitable mining)

10. Loss of mineralization

11. Breccia pipes, diatremes, variably mineralized

12. Weak development of mineralization

13. Multiple small volcanic and hydrothermal centers

14. Multiple cupolas

15. Efficient precipitation of Cu-Fe, Mo sulfide minerals to form a large porphyry deposit

16. Voluminous flow of magmatic-hydrothermal fluids through cupola zone, sustained by continuous magmatic recharge of batholith

17. Extension

18. Bimodal volcanism

19. Contraction

20. Deep plutonism (sills); volcanism

21. Reduced (infertile) arc magma

22. Reduced deep ocean

23. Reduced seawater sediments

24. Reduced (infertile) arc magma

25. Mafic or thin crust

26. Mafic volcanism

27. Mafic or thin crust

28. Mafic volcanism

29. Lithospheric hydration; dispersed magmatism

30. Shallow subduction

31. Limited magma flux

32. Diffuse volcanism and hydrothermal activity

33. Multiple cupolas

34. Lack of cupolas

35. Lack of recharge

36. Rapid cooling of system; limited mineralization potential

37. Lithospheric hydration; dispersed magmatism

38. Shallow subduction

39. Reduced seismicity, intermediate to felsic upper plate lithosphere: MASH processing

40. Extension

41. Bimodal volcanism

42. Contraction

43. Deep plutonism (sills); volcanism

44. Reduced (infertile) arc magma

45. Reduced deep ocean

46. Reduced seawater sediments

47. Reduced (infertile) arc magma

48. Lithospheric hydration; dispersed magmatism

49. Shallow subduction

50. Reduced magma in thickened, intermediate to felsic upper plate lithosphere: MASH processing

51. Extension

52. Bimodal volcanism

53. Contraction

54. Deep plutonism (sills); volcanism

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56. Reduced deep ocean

57. Reduced seawater sediments

58. Reduced (infertile) arc magma

59. Lithospheric hydration; dispersed magmatism

60. Shallow subduction

61. Reduced magma in thickened, intermediate to felsic upper plate lithosphere: MASH processing

62. Extension

63. Bimodal volcanism

64. Contraction

65. Deep plutonism (sills); volcanism

66. Reduced (infertile) arc magma

67. Reduced deep ocean

68. Reduced seawater sediments

69. Reduced (infertile) arc magma

70. Lithospheric hydration; dispersed magmatism

71. Shallow subduction

72. Reduced magma in thickened, intermediate to felsic upper plate lithosphere: MASH processing

73. Extension

74. Bimodal volcanism

75. Contraction

76. Deep plutonism (sills); volcanism

77. Reduced (infertile) arc magma

78. Reduced deep ocean

79. Reduced seawater sediments

80. Reduced (infertile) arc magma

81. Lithospheric hydration; dispersed magmatism

82. Shallow subduction

83. Reduced magma in thickened, intermediate to felsic upper plate lithosphere: MASH processing

84. Extension

7. Efficient precipitation of Cu-Fe, Mo sulfide minerals to form a large porphyry deposit

8. Postmineralization history: uplift and erosion to expose subvolcanic level

9. Large economic porphyry deposit (grade, tonnage, accessibility, and infrastructure sufficient for profitable mining)

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11. Breccia pipes, diatremes, variably mineralized

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37. Shallow subduction

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44. Reduced deep ocean

45. Reduced seawater sediments

46. Reduced (infertile) arc magma

47. Lithospheric hydration; dispersed magmatism

48. Shallow subduction

49. Reduced magma in thickened, intermediate to felsic upper plate lithosphere: MASH processing

50. Extension

51. Bimodal volcanism

52. Contraction

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54. Reduced (infertile) arc magma

55. Reduced deep ocean

56. Reduced seawater sediments

57. Reduced (infertile) arc magma

58. Lithospheric hydration; dispersed magmatism

59. Shallow subduction

60. Reduced magma in thickened, intermediate to felsic upper plate lithosphere: MASH processing

61. Extension

62. Bimodal volcanism

63. Contraction

64. Deep plutonism (sills); volcanism

65. Reduced (infertile) arc magma

66. Reduced deep ocean

67. Reduced seawater sediments

68. Reduced (infertile) arc magma

69. Lithospheric hydration; dispersed magmatism

70. Shallow subduction

71. Reduced magma in thickened, intermediate to felsic upper plate lithosphere: MASH processing

72. Extension

73. Bimodal volcanism

74. Contraction

75. Deep plutonism (sills); volcanism

76. Reduced (infertile) arc magma

77. Reduced deep ocean

78. Reduced seawater sediments

79. Reduced (infertile) arc magma

80. Lithospheric hydration; dispersed magmatism

81. Shallow subduction

82. Reduced magma in thickened, intermediate to felsic upper plate lithosphere: MASH processing

83. Extension

84. Bimodal volcanism

85. Contraction

86. Deep plutonism (sills); volcanism

87. Reduced (infertile) arc magma

88. Reduced deep ocean

89. Reduced seawater sediments

90. Reduced (infertile) arc magma

Figure 2. Schematic probability tree for the formation of a large economic porphyry Cu deposit as a product of arc magmatism. Each step corresponds to a section in the text, where the processes that might contribute to ultimate ore formation are discussed, and counteractive processes are considered. The roughly estimated cumulative probability of reaching each step is indicated.
generate hydrous, moderately oxidized ($\Delta$FMQ $\approx$ 1 to +2, where $\Delta$FMQ is the difference from the fayalite-magnetite-quartz oxygen buffer in log $f_O^2$ units), incompatible element-rich basaltic magmas. Such melts are quite distinct from, and evolve differently to anhydrous ($<$0.2 wt% H$_2$O; Danyushhevsky, 2001), relatively reduced ($\Delta$FMQ $\approx$ 0; Berry et al., 2018), incompatible element-poor MORB magmas.

The low water contents and reduced nature of MORB (and tholeiitic nascent arc) magmas render them unprospective for porphyry-type magmatic-hydrothermal ore formation because chalcophile and siderophile metals (Cu, Au, and platinum-group elements [PGEs]) will tend to be lost to early precipitating sulfide phases (e.g., Mitchell and Keays, 1981; Hamlyn et al., 1985; Peach et al., 1990), and large volumes of hydrothermal fluid will not be exsolved upon depressurization and crystallization.

In contrast, the slightly higher oxidation state of arc magmas means that a significant proportion of the dissolved sulfur is present as sulfate rather than sulfide, reducing the tendency of the magma to saturate early in large volumes of sulfide phases, and therefore to become depleted in metals (Carroll and Rutherford, 1985; Richards, 2003, 2011a, 2015; Jugo et al., 2005). In reality, arc magmas are probably saturated in sulfide phases at various points of their evolution (Borrok et al., 1999; Stavast et al., 2006; Jennen et al., 2010; Park et al., 2013; Richards, 2015; Williams et al., 2017), but where only small volumes of sulfide minerals are left as reside in mantle sources or fractionate during ascent, relatively abundant chalcophile elements such as Cu (50–100 ppm) are not significantly depleted, whereas sparse highly siderophile elements such as Au and PGE (a few ppb) are variably to strongly depleted (Richards, 2009, 2015). The high water content of arc magmas also means that they may exsolvemuch of this water prior to eruption during depressurization and crystallization in the upper crust. These fluids will generate subsurface hydrothermal systems that have the potential to transport and precipitate metals to form ore deposits.

Because these conditions characterize most subduction zones in the Phanerozoic, it is estimated that 90% of arc systems are fertile for porphyry ore formation (Fig. 2), consistent with the widespread occurrence of these deposits in Phanerozoic arcs worldwide. Only a few Phanerozoic arc systems are known to be largely barren, including the Paleo-Tethyan arcs of Eurasia and Japan, where it has been proposed that subduction of reduced oceanic crust (Richards and Şengör, 2017) or the presence of reduced lithologies in the upper plate (Tomkins et al., 2012; Sillitoe, 2018) may have locally degraded magma fertility by causing early sulfide saturation.

**Arc Magma Evolution**

The majority of porphyry Cu deposits form from evolved intermediate to felsic composition magmas (granodiorite to granite) in mature continental arcs, but they can also form from more mafic (dioritic) magmas in island arcs (Kesler et al., 1975; Clark, 1990; Khashgerel et al., 2006). However, the timing of porphyry formation in these oceanic settings is relatively late in the arc history (Hine and Mason, 1978; Richards, 2003; Rohrlach and Loucks, 2005; Richards and Kerrich, 2007), and few deposits are known to have formed from primitive magmas in the earliest, nascent stages of arc formation. The reason for this is likely the same as for the observation that the earliest magmas in nascent island arcs are tholeiitic, reduced, and have relatively low volatile and lithophile element contents (Arculus, 1994; Schmidt and Jagoutz, 2017): it takes a finite amount of time (several million years) for the flux of volatiles from the newly subducting slab to metasomatize and oxidize the mantle wedge to the point that it can produce oxidized, hydrous, calc-alkaline magmas that are fertile for porphyry ore formation (Reagan et al., 2008; Whattam and Stern, 2016; Hickey-Vargas et al., 2018).

**STEP 2: ARC GEODYNAMICS**

Subduction zone geodynamics have a fundamental control on arc magmatism (Fig. 3). At one extreme, shallow or flat subduction will terminate magmatism by eliminating the asthenospheric mantle wedge (Fig. 3C; Manea et al., 2017; Axen et al., 2018), whereas at the other extreme, a retreating slab may lead to arc rifting and bimodal magmatism.
that mantle-derived magmatism will cease or will migrate inland to where the slab eventually steepens (Fig. 3C). The Late Cretaceous Laramide porphyry Cu deposits of southwest North America are believed to have formed in this way, spread over a wide area extending hundreds of kilometers inland from the trench where the shallowly subducting Farallon slab finally dipped down into the mantle (Barton, 1996; du Bray, 2007; Leveille and Stegen, 2012). Although flat subduction itself does not seem to be conducive to porphyry formation except at its terminal edge, it has interesting implications for later tectonics and magmatism.

During flat subduction, the downgoing plate will dehydrate as before, but instead of the released fluids interacting with a hot asthenospheric mantle wedge, they will directly encounter the relatively cool base of the upper-plate lithosphere, causing hydration and metasomatism (Fig. 3C; Peacock, 1993; Sommer and Gauert, 2011; Porter et al., 2012). If flat subduction continues for a significant amount of time, the upper-plate lithospheric mantle will become extensively hydrated, lowering its solidus temperature. When subduction eventually steepens again and hot asthenosphere returns, the hydrated base of the upper plate will either undergo dehydration melting or will effectively change into asthenosphere by weakening, resulting in lithospheric erosion (Fig. 3D; Green et al., 2010; Sommer and Gauert, 2011). This effect may be seen in the
Puna-Altiplano region of Argentina-Bolivia, where a period of flat subduction followed by slab steepening triggered a major felsic volcanic flare-up (the Altiplano-Puna volcanic complex; de Silva, 1989), removal of the subcontinental mantle lithosphere (SCLM), crustal thickening, and plateau uplift (Kay and Kay, 1993; Kay et al., 1999; Gregory-Wodzicki, 2000; Beck and Zandt, 2002; Kay and Coira, 2009).

Flat subduction and associated magmatic flare-ups generally result in voluminous felsic volcanism rather than plutonism and are not typically associated with porphryy Cu deposits. However, in the Miocene–Pliocene El Indio–Pascua belt of Chile-Argentina, porphyry and epithermal Au-Ag-Cu deposits have been linked to either deep crustal dehydration melting during flat subduction and crustal thickening (Kay et al., 1999; Kay and Mpodozis, 2001; Muñoz et al., 2012), direct release of flat slab-derived fluids into the lower crust, causing partial melting (Bissig et al., 2003), or tectonic changes immediately prior to or during the initial stages of subduction flattening (Skewes et al., 2002). It is the opinion of this author that these deposits mainly reflect stress changes in the upper plate associated with slab flattening acting on a mature arc, rather than the actual condition of flat subduction, which terminates magmatism (and any associated ore formation). This may also explain the observation of a temporal correlation between some porphyry Cu deposits and subduction of aseismic ridges and other buoyant features on the downgoing plate (Cooke et al., 2005). Subduction of such anomalies may cause transient periods of shallow subduction and increased inter-plate stress coupling. These changes acting on a mature arc, particularly where they involve a transition from compressional to transpressional stress, may lead to a brief pulse of upper-crustal plutonism and porphyry formation, prior to shutdown of the arc (Tosdal and Richards, 2001).

**STEP 3: UPPER-PLATE DEEP CRUSTAL MAGMA PROCESSING**

Hildreth and Moorbath (1988, p. 484) stated, “Not a single primitive basalt (having high enough Mg, Ni, and Cr for equilibrium with peridotite) is known to us in Chile,” and noted that they are rare in continental arcs globally. They concluded that a simple, reproducible mechanism was required to transform the flux of basaltic magma generated by partial melting in the mantle wedge into the intermediate composition calc-alkaline (andesitic) magmas that characterize the Andes and other volcanic arcs worldwide. They termed this the MASH process, standing for melting, assimilation, storage, and homogenization.

The MASH model considers mantle-derived basaltic magmas to be too dense to pass through the upper-crust in the absence of a hydraulic head (Walker, 1989), restricting their eruption at surface to extensional tectonic settings (back-arc rifts; Fig. 3B) or areas of thin or mafic crust (nascent island arcs). Elsewhere, such magmas will tend to pool at the Moho, where their latent heat of crystallization will be transferred to crustal rocks, leading to partial melting (Figs. 1 and 3A). Hildreth and Moorbath (1988, p. 483) envisaged the MASH zone to be “a plexus of dikes, sills, pods, small chambers, and mushy differentiated intrusions,” where basaltic magmas evolve by fractional crystallization and mix with felsic crustal melts to form hybrid magmas of intermediate composition and leave ultramafic to mafic lower-crustal cumulate zones (as observed in the Talkeetna arc, Alaska, the Kohistan arc, Pakistan, and the Sierra Vallee Fértil arc, Argentina; Greene et al., 2006; Jagoutz et al., 2007; Walker et al., 2015). The lower density of the derivative andesitic melts would then allow them to rise to shallower levels in the crust, where they accumulate again in mid- to upper-crustal magma chambers (batholiths) or erupt due to the further density decrease caused by vesiculation.

The MASH concept has been tested widely since its introduction and modified only in detail. Annen et al. (2006) introduced the concept of deep crustal hot zones (Fig. 1) to reflect the interaction of infiltrating magmas with different levels of the crust, explaining the significant geochemical and isotopic heterogeneity commonly seen in magmas erupted from a single arc volcanic complex, and numerous studies have confirmed the origin of andesitic magmas through magma mixing and assimilation processes (e.g., Eichelberger, 1978; Kay, 1980; Ussler and Glazner, 1988; Straub and Martin-Del Pozzo, 2001; Halter et al., 2004; Dufek and Bergantz, 2005; Price et al., 2005; Zellmer et al., 2005; Eichelberger et al., 2006; Richards et al., 2006, 2013; Reubi and Blundy, 2009; Koteas et al., 2010; Schiano et al., 2010; Solano et al., 2012; Nandedkar et al., 2014). The result is that, although these common petrogenetic processes result in relatively uniform major-element compositional ranges for arc magmas globally (from calc-alkaline basaltic andesites, andesites, dacites, and rhyolites), their mineralogy, trace-element, and radiogenic isotopic compositions record the details of these processes in terms of interactions with upper-crustal crustal rocks, overlain by fractional crystallization effects (assimilation-fractional crystallization [AFC]; DePaolo, 1981).

Fractional crystallization of hydrous, moderately oxidized arc magmas results in extensive early fractionation of ferromagnesian silicate minerals (olivine, pyroxene, and amphibole) and spinel, prior to cotectic crystallization of plagioclase, with the result that alkali contents in fractionated melts rise initially relative to Fe (Sisson and Grove, 1993). In contrast, earlier crystallization of plagioclase from dry tholeiitic magmas leads to alkali depletion and Fe enrichment (Bowen, 1928; Chin et al., 2018). These differences are reflected in the calc-alkaline and tholeiitic fractionation trends, respectively, on alkali-FeO-MgO (AFM) diagrams. A further effect observed at trace-element levels is the enrichment of hydrous andesitic magmas in Sr and depletion in middle and heavy rare-earth elements (MREEs and HREEs, including Y), again resulting from delayed plagioclase crystallization (Sr enrichment) and abundant early crystallization of amphibole (which preferentially partitions MREEs and Y; Castillo et al., 1999; Macpherson et al., 2006; Richards and Kerrich, 2007; Nandedkar et al., 2016). Such magmas tend to display lustric-shaped normalized REE patterns, which flatten in the MREE and may increase again slightly in the HREE (Sisson, 1994; Nandedkar et al., 2016). Garnet fractionation or its presence as a residual phase in the source will also strongly fractionate Y and HREE but does not preferentially partition MREE, resulting in monotonically decreasing patterns from LREE to HREE (Davidson et al., 2013). Andesitic magmas with resultant high Sr/Y and La/Yb values are commonly referred to as adakites,
but this term has been specifically associated with the melting of subducted oceanic crust (Kay, 1978; Defant and Drummond, 1990), which can lead to similar trace-element characteristics. Although it is acknowledged that “slab melting” does occur in Phanerozoic arcs under certain restricted circumstances (Defant and Drummond, 1990; Kay et al., 1993; Guivel et al., 2003), the majority of Phanerozoic adakite-like rocks, especially in continental arcs, reflect MASH and AFC processes in hydrous magmas (Feeley and Hacker, 1995; Castillo et al., 1999; Garrison and Davidson, 2003; Richards and Kerrich, 2007). High Sr/Y values (>20) in igneous rocks have been widely used as an indicator of magmatic fertility for porphyry Cu formation (Thiéblemont et al., 1997; Sajona and Maury, 1998; Oyarzun et al., 2001; Chiaradia et al., 2012; Loucks, 2014; Bissig et al., 2017), most likely reflecting high magmatic water content as a prerequisite for forming magmatic-hydrothermal systems (López, 1982; Dilles, 1987; Lang and Titley, 1998; Richards et al., 2001, 2012; Rohrlich and Loucks, 2005; Schutte et al., 2010; Richards, 2011b).

In Figure 2, I suggest that the probability that any given upper plate arc system will develop a mature MASH zone is ~50%, taking the cumulative probability down to ~10% (steps 1–3). The relatively high probability of successfully negotiating step 3 reflects the global uniformity of the MASH process in Phanerozoic arcs. Except in nascent oceanic arcs, or arcs undergoing rifting, a MASH zone will inevitably develop if a sustained flux of mafic mantle-derived magma underplates the thickening arc crust. This process may not lead to any particular increase in metal endowment in the magmatic system (perhaps even the opposite; Chiaradia 2014), but it may act to enrich derivative melts in volatiles (H2O, S, and Cl) and further increase oxidation state (Burnham, 1979; Candela, 1992; Streck and Dilles, 1998; Rohrlich and Loucks, 2005; Chambeafort et al., 2008; Richards, 2015; Hutchinson and Dilles, 2019) and forming relatively viscous, intermediate composition magmas. Such magmas will have a higher probability of rising buoyantly from the lower crust but stilling in mid- to upper-crustal plutons, rather than being erupted at surface (Richards, 2003, 2015; Chiaradia and Caricchi, 2017). In addition, the process of storing magma in lower-crustal MASH zones serves to build up the volume of magma available for upper-crustal plutonism, and therefore for shallow porphyry ore formation.

One additional factor can undermine the fertility of arc magmas at this point in their evolution to a more reduced state. The relatively high oxidation state of Phanerozoic arc magmas (AFMQ ≅+1 to +2) limits the tendency of these S-rich magmas to undergo early sulfide saturation with resultant removal of chalcopyrite and sidereophile elements from the melt (Mitchell and Keays, 1981; Hamlyn et al., 1985; Peach et al., 1990; Li and Audétat, 2015). As discussed by Richards (2011a, 2015), the high overall sulfur content of arc magmas means that small amounts of sulfide melt or minerals may fractionate from these magmas even under moderately oxidized conditions, which may explain their strong depletion in sparse highly siderophile elements (Au and PGE). However, unless large amounts of sulfide minerals separate from the magma (e.g., Lee et al., 2012; Chiaradia, 2014), the contents of Cu may not be significantly affected, with the result that the magmas remain fertile for Cu ore formation, but not for PGE, or to variable degrees, Au.

Reduction would cause voluminous sulfide exsolution from S-rich arc magmas, and this could occur either at source or by interaction with reduced lithologies in the deep crust. Subduction of reduced sediments has been proposed to generate locally reduced conditions in arcs (Wang et al., 2007; Richards and Şengör, 2017), whereas reducing conditions in deep oceans throughout the Precambrian have been proposed to explain the rarity of porphyry Cu deposits in older rocks (Evans and Tomkins, 2011; Richards and Mumin, 2013a, 2013b). These conditions would result in the retention of the bulk of chalcopyrite and siderophile elements in the mantle source region, decreasing magma fertility for later porphyry-type ore formation. Tomkins et al. (2012) have also suggested that the presence of carbonaceous materials in lower-crustal sequences may cause reduction of arc magmas in the MASH zone, leading to voluminous sulfide precipitation and retention of chalcopyrite and siderophile elements in cumulates, potentially forming orthomagmatic Ni-Cu-PGE ore deposits (e.g., Manor et al., 2016). Although this process reduces the fertility of derivative magmas, the metals in these sulfide phases can be remobilized during later tectonomagmatic events such as collision or crustal thickening, potentially forming post-subduction porphyry Cu ± Au deposits, as found in Tibet and other parts of the Tethyan collided orogen (Richards, 2009; Lee et al., 2012; Hou et al., 2015).

### STEP 4: UPPER-CRUSTAL MAGMA EMPLACEMENT

Silicate magmas rise toward the surface due to their buoyancy relative to the surrounding rocks. In situations where hydraulic connectivity between the source region and surface is established, such as in extensional terrains, magma density is matched against the density of the entire rock column, and relatively dense basaltic magmas can be erupted through lithosphere that includes dense mantle and lower crust (Walker, 1989; Takada, 1994). However, in contractional settings, hydraulic connectivity is lost, and magma ascent is controlled by its density relative to the immediately surrounding rocks. Dense basaltic magmas will tend to be trapped in the lower crust, as discussed above, but more evolved andesitic magmas can rise to mid-crustal levels, where their lower density matches crystallized granodioritic rocks (Herzberg et al., 1983). This represents a second crustal accumulation level, and if magma supply is sustained, can result in the construction of large arc batholiths (Fig. 1).

Magma ascent through the crust is thought to be controlled by visco-elastic processes, with flow predominantly along fractures (Lister and Kerr, 1991; Clemens and Mawer, 1992; Rubin, 1993, 1995; Pettford et al., 2000). Disregarding extensional settings (where normal faults can allow primitive magmas to ascend directly to surface, with little potential to form upper crustal plutons and porphyry deposits; Takada, 1994), contractual and transpressional settings will control magma ascent in different ways. Simple horizontal compression will favor vertical opening and horizontal propagation of fractures, resulting in the emplacement of sills (Ida, 1999; Simakin and Talbot, 2001; Chaussard and Amelung, 2014). Such conditions will tend to trap mafic magmas in the deep
crust, in the sill complexes envisaged by Hildreth and Mooibath (1988). Any magma that escapes into the upper crust is likely to do so under high pressure, leading to volcanism (e.g., Tibaldi, 2008). Consequently, many contractional “orogenic” episodes in arcs are characterized by mafic-intermediate volcanism but little plutonism (except in the deep crust; see review in Richards, 2003).

In contrast, transpressional tectonic settings, which are common in obliquely convergent arcs, feature orogen-parallel strike-slip fault systems along which extensional (or lower confining pressure) vertically oriented pathways can develop at step-overs and jogs (de Saint Blanquat et al., 1998; Tosdal and Richards, 2001). These structures can act to focus magma ascent from deep reservoirs toward the upper crust, restricting direct hydraulic connectivity to surface, thereby promoting plutonism over volcanism (D’Lemos et al., 1992; Tikoff and Teyssier, 1992; Vigneresse, 1995; Benn et al., 1998; Brown and Solar, 1999; Klepeis et al., 2003; Romeo et al., 2006; Olivier et al., 2016; Bedrosian et al., 2018). Transpressional conditions are therefore ideally suited to the formation of mid- to upper-crustal batholiths and associated porphyry deposits (Fig. 1; Tosdal and Richards, 2001; Carranza and Hale, 2002; Drew, 2006; Cloos and Sapiie, 2013). The transition from a protracted period of compression, leading to the build-up of a large lower-crustal MASH zone, to transpression, would facilitate the rapid ascent and emplacement of a large volume of evolved magma (i.e., a batholith). These two criteria, large volumes and rapid emplacement of magma, are critical for subsequent ore formation (Richards, 2003; Chiaradia and Caricchi, 2017; Schöpa et al., 2017). However, these tectonic conditions, and especially the transition from contractional to transpressional strain, are relatively uncommon during the history of arcs, reducing the overall probability to perhaps 1% (steps 1–4 in Fig. 2).

### STEP 5: FOCUSED FLUID EXSOLUTION FROM UPPER-CRUSTAL MAGMA CHAMBERS

A key step in the formation of porphyry Cu deposits is the transfer of metals from silicate melt to an exsolving volatile phase. Copper and many other base and precious metals partition strongly into saline aqueous fluids relative to silicate melt (Candela, 1992; Candela and Holland, 1984; Candela and Piccoli, 1995; Heinrich et al., 1999; Frank et al., 2011; Simon et al., 2005, 2006; Williams et al., 1995), and so, if this fluid phase equilibrates with a large volume of magma, it can be expected to efficiently extract metals from the melt. Recently it has been suggested, based on observations of magmatic sulfide inclusions in igneous rocks associated with some porphyry deposits, that partitioning of metals into sulfide minerals or melts separating from the magma prior to fluid exsolution may serve as an important pre-enrichment step, with those sulfide phases subsequently breaking down (oxidizing) and releasing their metal contents during later fluid exsolution (Spooner, 1993; Keith et al., 1997; Borrok et al., 1999; Larocque et al., 2000; Halter et al., 2002, 2005; Stanavski et al., 2006; Nadeau et al., 2010, 2016; Wilkinson, 2013). However, it is not clear to this author why this extra step is necessary or increases the probability of ore formation, notwithstanding that some sulfide saturation is probably inevitable in these S-rich magmas as they fractionate and crystallize. In fact, if sulfide separation was voluminous and its redissolution was not almost complete, then this process would be expected to decrease the efficiency of ore formation. My assessment is that this process is probably common in upper-crustal arc magmatic systems, but it may not be a critical step in ore formation (Audétat and Pettke, 2006; Richards, 2015).

Exsolution of volatiles from magma during depressurization and/or cooling is an inevitable consequence of decreasing solubility in silicate melts with decreasing pressure and temperature, as well as the low concentration of volatiles in most crystallizing phases relative to silicate melt (which increases the volatile content of the residual melt; Burnham, 1979; Candela, 1989). This has led to a distinction between “first boiling” during ascent (depressurization) and “second boiling” during cooling and crystallization (Burnham, 1979; Candela, 1989), although technically this is not boiling (the transformation of a liquid to a vapor), and the process of volatile exsolution is probably semi-continuous during the ascent and crystallization of volatile-rich magmas (Holloway, 1976; Hedenquist and Lowenstein, 1994; Lowenstein, 2001; Baker and Alletti, 2012). Volatiles exsolve in inverse order to their solubility, with CO₂ being released first, probably during early ascent from deep crustal levels, followed by H₂O, S, Cl, and F (Holloway, 1976; Stix et al., 1993; Lowenstein, 2001; Caricchi et al., 2018). Arc magmas may contain more CO₂ than commonly assumed (3000–4000 ppm CO₂; Wallace, 2005; Blundy et al., 2010), such that an exsolved fluid phase may be present throughout their ascent from deep crustal or even mantle levels (Holloway, 1976). Because of the relatively low solubility of CO₂ in silicate melts compared to H₂O, the initial deeply exsolved fluids will have high CO₂/H₂O values (Newman et al., 2000), becoming H₂O-dominant by the time magmas reach mid- to upper-crustal levels (consistent with the aqueous, relatively CO₂-poor composition of magmatic-hydrothermal fluids sampled in most shallow crustal porphyry Cu systems). Little is known about the composition or fate of deeply exsolved fluids in arc magmatic systems, or whether these fluids physically separate from or ascend with magma rising into the shallow crust. Their much lower viscosity and density compared to silicate magmas suggest that they may rise through the crust as a separate plume, but it is also possible that much of this deep fluid is absorbed by reaction with hot wall rocks to form carbonate and hydrous silicate alteration minerals (e.g., Rosing and Rose, 1993).

By the time the magma has reached the mid–upper crust and evolved to more felsic compositions, much of its CO₂ will have been lost, and exsolved fluids will be dominantly hydrous, with high S-Cl concentrations. This is the depth at which porphyry ore-forming fluids are thought to originate, as indicated by the compositions of the deepest fluids sampled (in fluid inclusions) from porphyry systems. For example, Rusk et al. (2004, 2008) report that fluid inclusions from deep quartz veins from the Butte, Montana, porphyry Cu–Mo deposit trapped 550–700 °C supercritical aqueous fluids at depths of 6–9 km, containing 2–5 wt% NaCl equivalent, minor amounts of CO₂ (2–8 mol%), and up to 1.3 wt% Cu (typically ~1000 ppm Cu). Similar deep magmatic-hydrothermal fluids were reported...
from the Bingham Canyon porphyry Cu-Au-Mo deposit, Utah (Redmond et al., 2004). The high Cu contents (as well as other elements such as Fe, Mo, Pb, Zn, Ag, Au, As, and Sb) reflect the strong partitioning of metals into high-temperature Cl-S–bearing hydrothermal fluids, relative to silicate melt (Candela and Holland, 1984, 1986; Candela, 1992; Williams et al., 1995; Simon et al., 2006; Frank et al., 2011; Tattitch and Blundy, 2017; Zajacz et al., 2017).

The probability that such fluids will be exsolved from hydrous arc magmas emplaced in the mid–upper crust is 100% (i.e., it is inevitable). Simple calculations of the volume of typical andesitic magma carrying ~60 ppm Cu that could supply 10 Mt Cu to a large porphyry Cu deposit return values ≥100 km$^3$ (Dilles, 1987; Cline and Bodnar, 1991; Richards, 2005; Steinberger et al., 2013), whereas similar calculations of the volume of fluid containing ~1000 ppm Cu that could transport this amount of Cu come to 10–20 km$^3$ (which could be sourced from 100 to 200 km$^3$ of magma containing 4 wt% H$_2$O; Plank et al., 2013). Such magma and fluid volumes are not unrealistic in mid- to upper-crustal magma chambers, as suggested by the size of typical arc plutons and the scale of geodetic and geophysical anomalies around active arc volcanoes (e.g., Schilling and Partzsch, 2001; ANCORP Working Group, 2003; Comeau et al., 2015; Perkins et al., 2016; Laumonier et al., 2017). However, under normal conditions, this fluid will be released slowly and progressively as the pluton cools and crystallizes, and will mostly be dispersed along fractures and joints into the overlying country rocks (Fig. 4A). Diffuse fluid flow will not form ore deposits and will likely only be detectable as broad low-grade hydrothermal alteration zones, with no significant metal enrichments. Such systems are very common in arc volcanoplutonic complexes and are probably the default product of batholithic devolatilization.

In contrast, porphyry Cu deposits form where the flow of large volumes of magmatic-hydrothermal fluid has been focused into narrow (typically ≤1-km-wide) cylindrical or elliptical zones rising above the batholith, termed “cupolas” (Fig. 5). These zones may be the equivalent of feeder systems below large composite volcanoes (e.g., Stillitoe, 1973), although surface volcanism does not appear

![Diagram](https://example.com/diagram.png)

Figure 4. Evolution of a mid- to upper-crustal batholith. (A) Progressive batholith growth by semi-continuous recharge, crystallization, and degassing, with limited volcanism. (B) Triggering of explosive fluid flow through cupola zone, potentially leading to breccia pipe or diatreme formation in shallow systems, and explosive eruption, if the system breaches the surface.
to be necessary for porphyry formation (Sillitoe, 2010). In detail, cupola zones in porphyry systems are characterized by complex overlapping histories of dike and stock emplacement, magmatic and hydrothermal breccia pipe formation, hydrothermal alteration, and vein and stock-work formation, which are described in more detail below. The origin of these structures is unclear, but their common pipe-like form (as opposed to planar dikes or sills) suggests either that they are controlled by vertically oriented structural intersections in the tensionally stressed cover rocks above the inflating batholith, and/or that they originate as breccia pipes or diatremes generated by explosive fluid release from the magma chamber (possibly also utilizing vertical structural intersections; Fig. 4B; Richards, 2011a, 2018).

Cupola zones develop at apical irregularities in the roof of the underlying batholith; these irregularities in turn develop in response to the force of intrusion and differential uplift or foundering of brittle cover rocks (e.g., Guillou-Frottier and Burov, 2003; Montanari et al., 2017). These apical regions will focus the accumulation of lower density bubbly magma and fluids, whose buoyant, high-pressure ascent toward the surface will then be channeled along extensional structures in the overlying brittle rocks (or carapace; Burnham, 1979; Dilles, 1987; Shinohara and Hedenquist, 1997; Cloos, 2001; Tosh dal and Richards, 2001). Because there will always be apical irregularities at the top of any magma chamber, accumulation of low-density magma and fluid at these points is also inevitable. However, the degree to which this accumulation occurs, and the proportion of the total fluid volume that becomes focused here, will depend on the number and prominence of such apices. Numerous coeval apical zones will disperse fluid flow and bubbly magma ascent from the batholith, whereas a single dominant apex could focus the entire magma-fluid flux. Therefore, where there is a greater degree of fluid focusing, there is a concomitant increase in the probability of subsequent ore formation during ascent and cooling of the magmatic-hydrothermal fluids.

The formation of a singular large apical zone in the roof of a batholith is less likely than the development of multiple small apices, or simple fluid dispersion into the brittle cover rocks, but the exact probability is not known. Nevertheless, there is some evidence that, once such a central structure is developed, it becomes self-reinforcing as continued flow of hot fluids and melts establish a heated, permeable pathway toward the surface (Weis, 2015; Ardill et al., 2018). I have therefore estimated the probability of forming a single, large, well-focused cupula zone above a batholith to be ~10%, bringing the cumulative probability of reaching the next step to 0.1% (Fig. 2).

STEP 6: SUSTAINED, RAPID FLUID FLOW IN CUPOLAS

The total amount of metal that could potentially be deposited by fluids flowing through the cupula zone is limited by the volume of fluid, which in turn is limited by the volume of devolatilizing magma. As noted above, a batholith containing 100–200 km³ of typical andesitic magma (4 wt% H₂O) could in theory source all the fluid and Cu in a giant porphyry Cu deposit (10 Mt Cu), but this would require almost complete degassing of the magma chamber, highly efficient extraction of Cu from the melt, and channeling of the entire fluid volume up a single cupula zone. A decrease in the efficiency of any one of these processes, or dispersion along multiple pathways, correspondingly either reduces the amount of metal that could be deposited (forming a smaller, probably subeconomic deposit), or increases the volume of magma required to source the ore deposit. Furthermore, there is a requirement that this fluid exsolution event should be extremely rapid in terms of the overall lifespan of the batholith.

Porphyry Cu deposits form on geologically very short timescales of 100,000 yr or less (Arribas et al., 1995; Marsh et al., 1997; Shinohara and Hedenquist, 1997; Barnes, 2000; von Quadt et al., 2011; Weis et al., 2012; Chiaradia et al., 2013;
Chelle-Michou et al., 2015, 2017; Mercer et al., 2015; Buret et al., 2016; Correa et al., 2016; Li et al., 2017; Cernuschi et al., 2018). In contrast, a typical arc batholith has a magmatic lifespan of one to several million years, reflecting incremental construction by quasi-continuous magma supply from lower-crustal MASH or hot zones (Matz et al., 2006; Miller et al., 2007; Schaltegger et al., 2009; Paterson et al., 2011; Fiannacca et al., 2017). An individual batholith rarely generates more than one large economic porphyry Cu deposit (commonly none), and so such deposits must be considered to be rare (albeit reproducible) and short-lived events within the overall life of an arc batholith. The short duration of ore formation reflects a fundamental characteristic of porphyry deposits, which is that ore deposition is primarily controlled by the very steep temperature gradient along the fluid pathway, from magmatic temperatures (600°–700 °C) at ~4–5 km depth (rarely to 9 km; Rusk et al., 2008), to <300 °C at 1–2 km depth (Shinohara and Hedenquist, 1997). Copper is mainly deposited over a narrow temperature interval between ~550°–350 °C along this pathway, at depths of 1.5–4 km (Fig. 5; Redmond et al., 2004; Landtwing et al., 2005; Klemm et al., 2007; Cernuschi et al., 2018). Sustaining these high temperatures in shallow crustal rocks requires continuous (likely pulsed) and rapid flow of hot fluids (Weis et al., 2012; Mercer and Reed, 2013; Weis, 2015; Cernuschi et al., 2018). If the flow stops or slows, the cupola zone will rapidly cool down, ore deposition will cease, and any new pulse of magmatic-hydrothermal fluid flow will need to establish a new pathway, effectively restarting the process of ore formation; in such cases, the same total tonnage of metal might be transported by the combined hydrothermal systems as by a single flow event, but it will likely be deposited in different places, and therefore at lower grades. These characteristics place significant constraints on the processes that give rise to ore formation. For example, it will not be sufficient simply to progressively devolatilize a 200 km³ magma chamber over its typical million year history, because it will not be possible to sustain a single hot fluid pathway toward the surface over this long period, with the result that the exsolved fluids will cool quickly at depth before their flow can be focused, leading to reprecipitation of metals at background levels. Instead, it appears that the bulk of the fluid must be released suddenly and the source magma chamber essentially devolatilized within a few tens of thousands of years, if ore formation is to be successful (Huber et al., 2012; Chelle-Michou et al., 2014; Schöpa et al., 2017). Indeed, the initial release of high-temperature fluid may be explosive, characterized by the formation of breccia pipes and diatremes, and occurring essentially instantaneously (on the order of a few tens or hundreds of years; Sillitoe, 1986; Vargas et al., 1999; Landtwing et al., 2002; Skewes et al., 2002; Cathles and Shannon, 2007; Vry et al., 2010). Lower-temperature fluids may continue to flow over longer periods (10⁶–10⁷ yr) as the system wanes and fluid drains from distal parts of the magma chamber (Candela, 1997; Cathles and Shannon, 2007).

The timing of these fluid-release and ore-forming events appears to be random over the life of a batholith (Chiaradia and Caricchi, 2017), although there is a tendency for large porphyry deposits to form relatively late in the evolution of arc magmatic systems (Richards, 2005; Rohrlach and Loucks, 2005; Chiaradia et al., 2009a) reflecting the time required to build up a sufficient flux of volatile-rich magma and heat reaching the upper crust, as noted above. In terms of the history of individual batholiths, there are some examples where ore formation is early (Chang et al., 2017; Li et al., 2017), but the majority are intermediate to late (e.g., Steinberger et al., 2013), albeit with no clear pattern. These events are therefore considered to be stochastic: that is, although the processes involved are largely understood and predictable, their actual occurrence is unpredictable (Chiaradia and Caricchi, 2017; Richards, 2018).

Mid- to upper-crustal batholithic magma chambers (5–10 km depth) are thought to exist for most of their super-solidus lives as crystal mushes (Gelman et al., 2013; Klemetti, 2016), in which significant volumes of fluid may be trapped as interstitial bubbles (seen as vesicles or miarolitic cavities in granitic rocks; Candela and Blevin, 1995; Candela, 1997; Edmonds et al., 2014; Edmonds and Wallace, 2017) or in supersaturated viscous silicate melt (Gardner et al., 2000). Normally, such fluids would slowly bleed off along intergranular pathways and fractures as the magma cools and crystallizes (Candela, 1991; Dunbar et al., 1996; Boudreau, 2016; Parmigiani et al., 2016; Edmonds and Woods, 2018), leading only to diffuse hydrothermal alteration in the crystallized carapace and cover rocks (Fig. 4A). In contrast, ore formation requires a mechanism for rapid exsolution and focused escape of fluids from a large volume of the batholith.

Building on ideas relating to the triggering of large explosive volcanic eruptions (Christopher et al., 2015; Cashman et al., 2017; Sparks and Cashman, 2017), Richards (2018) suggested that an external trigger, such as sudden magma chamber depresurization or seismic shaking, might be required to prompt the onset of wholesale devolatilization in a batholithic-sized magma chamber (cf. Cathles and Shannon, 2007). Shearing is known to increase volatile permeability in magma flowing in subvolcanic conduits (Gardner et al., 1996; Cannata et al., 2010; Carey et al., 2012; Kushnir et al., 2017), but a large stagnant magma chamber may require an external shock, such as a mega-earthquake, to achieve the same effect (Davis et al., 2007; Namiki et al., 2016). Shearing produces transient pressure reductions that promote volatile exsolution, and also mechanically reducing the percolation threshold along grain boundaries, allowing channelization and rapid fluid escape. Similarly, pressure reduction on the magma chamber by sudden unroofing or mass wasting (e.g., volcanic sector collapse; Voight et al., 2006) could have the same effect.

Christopher et al. (2015) and Sparks and Cashman (2017) have proposed that overturn of density instabilities arising from differential accumulation of fluids within a large heterogeneous magma chamber might lead to explosive volcanic eruptions where bubbly magma reaches the surface. A similar mechanism, but in which fluids separate from magma at depth to form a subsurface hydrothermal system, is envisaged here (Fig. 4B; cf. Edmonds and Woods, 2018). Dunbar et al. (1996) anticipated this in a conceptual model (their fig. 4) that links coalescence of interstitial bubbles to form porous zones in the crystal mush, escape by fracture flow into the brittle carapace, and stock-work vein and
breccia formation in the shallow cover rocks (see also Candela, 1991). Once such a fluid pathway develops, it is likely to become self-reinforcing as a structurally weak and heated conduit, in the same way that the entire magma column is thought to become self-focusing by sustained magma flow (McCuaig and Hronsky, 2014; Weis, 2015; Ardill et al., 2018). Thus, although the initial, explosive fluid flow event may be marked by a breccia pipe or diatreme (Fig. 4B), this permeable structure will subsequently be intruded by more slowly flowing bubbly magma (forming dikes and stocks) interleaved with multiple sets of veins and stock works, each potentially depositing metals to build up an ore deposit (Figs. 5, 6A, and 6B; Sillitoe and Sawkins, 1971; Norton and Cathles, 1973; Burnham and Ohmoto, 1980; Burnham, 1985; Sillitoe, 1985; Carrigan et al., 1992; Maimon et al., 2012). The largest ore deposits will likely be those that feature the most complex overlapping and overprinting magmatic intrusion, hydrothermal brecciation, veining, and alteration events, to the extent that the earliest breccia pipe stages may be almost unrecognizable (Figs. 5 and 6C; Skewes and Stern, 1994; Vargas et al., 1999; Vry et al., 2010; Harrison et al., 2018).

The key difference between a weakly mineralized hydrothermal alteration zone and a giant ore deposit (or a small versus giant explosive eruption) may simply be one of scale and timing, with a sudden and violent external trigger leading to the latter, and quasi-steady-state fluid exsolution leading to the former. Overall, the probability of a major event affecting a large batholith, albeit one already primed to exsolve hydrothermal fluids, is relatively low, perhaps 1 in 10, reducing the overall probability of achieving this step to 0.01% (Fig. 2).

STEP 7: ORE DEPOSITION

The previous steps in this analysis have focused on the source and transportation of ore metals in arc magmas and exsolved fluids. It has been shown that, although such systems commonly have the potential to deliver metals into upper-crustal hydrothermal systems, few (perhaps 0.01%) do this efficiently. Of those systems that achieve this step,
the transported metal load still needs to be deposited in high enough concentrations (typically >0.4% Cu equivalent, or >1.2% for deeper deposits that can only be mined underground) to be considered economic (i.e., an ore deposit, as opposed to a mineral deposit or geochemical anomaly). Many large porphyry systems exist, but perhaps only one in ten of these prove to be economic to mine (under present-day conditions), reducing the overall probability of forming a large ore deposit to perhaps 0.001% (Fig. 2). The greatest risk to successful ore formation at this stage is that the fluids vent directly to surface through breccia pipes, diatremes, or large explosive volcanic eruptions (perhaps due to shallow emplacement depth), rather than cooling and circulating below surface where their metal load can be precipitated (Figs. 4B and 5). At the Lihir epithermal gold mine in Papua New Guinea, an earlier stage of porphyry Cu-Au-Mo deposit formation was abruptly terminated by an explosive caldera eruption or volcanic sector collapse (Moyle et al., 1990; Müller et al., 2002; Carman, 2003; Sykora et al., 2018), whereas at the El Teniente porphyry Cu-Mo deposit in Chile, ore formation was followed by emplacement of the large (1-km-wide) Braden Breccia pipe (Skewes et al., 2002; Vry et al., 2010).

Assuming that the bulk of the magmatic-hydrothermal fluid flux remains below surface, high-grade zones of hypogene mineralization in porphyry Cu ± Mo ± Au deposits are commonly characterized by either highly focused fluid flow or the presence of reactive wall rocks that efficiently precipitated ore-forming sulfide minerals (Sillitoe, 2010; Richards, 2013).

In the first case, high-grade ore zones are characterized by sheeted vein sets or dense vein stockworks, in which each vein has introduced additional metal into the rock volume (e.g., Ridgeway, Australia, Wilson et al., 2003; Red Chris, Canada, Rees et al., 2015). In some of the most extreme examples, such as the Oyu Tolgoi porphyry Cu-Au deposit in Mongolia, quartz veins exceed 90% of the rock volume, and grades of 3%–4% Cu plus 1–2 g/t Au were encountered over 100 m intervals during drilling (Figs. 7A and 7B; the Hugo Dummett North zone of the deposit contains a measured resource of 41 Mt at 1.56% Cu and 0.41 g/t Au, an indicated

Figure 7. Samples of porphyry Cu ore. (A and B) Drill core samples from Oyu Tolgoi, Mongolia, showing intense stock work and sheeted veining in quartz monzodiorite intrusions, with abundant chalcopyrite (Cp; yellow) and bornite (Bn; purple). The sample in (B) is from drill hole OTD 976 at ~12270 m, and contains 3.18% Cu and 1.18 g/t Au. The scale bar is marked in centimeter intervals. (C) Disseminated chalcopyrite (Cp; yellow) in mafic andesitic wall rock, El Teniente, Chile (reflected light photomicrograph). (D) Quartz vein stock work with chalcopyrite (Cp; yellow) and molybdenite (Mo; blue-gray) in biotite-altered (potassic) monzonite intrusion, Bingham Canyon, Utah.
STEP 8: POSTMINERALIZATION HISTORY

Porphyry Cu deposits typically form 1.5–4 km below the surface (rarely down to 9 km), whereas high-sulfidation type epithermal Cu-Au deposits may form coevaly in the shallower parts of the system (Arribas et al., 1995; Einaudi et al., 2003; Sillitoe and Hedenquist, 2003; Simmons et al., 2005). This means that, at the time of formation, most porphyry Cu deposits will be too deep to mine from open pits at surface, and only the richest deposits could be mined as underground operations. However, porphyry systems form in active convergent margin settings, where uplift and erosion are typically rapid. Consequently, after only a few million years, these systems may have been eroded down to levels where the mineralization can be accessed from surface. It is for this reason that the Eocene–Miocene arcs of the South American Andes host many of the largest porphyry Cu deposits in the world (Sillitoe, 1992). Younger stratovolcanic centers of the current High Andes may well overlie porphyry systems, but until the volcanic cover has been eroded down, they are unlikely to be economic (e.g., Sillitoe, 1973, 1975). Only where exceptionally high uplift and erosion rates occur will younger deposits be exposed at surface, such as the Grasberg (3 Ma; Pollard et al., 2005) and Ok Tedi (1 Ma; Page and McDougall, 1972) porphyry Cu-Au deposits in Highlands of New Guinea. In some cases, erosion or unroofing may be so rapid that the porphyry system is uplifted during formation, leading to “telescoping,” where shallow-level epithermal alteration overprints the marginally earlier porphyry system (Sillitoe, 1994; Houston and Dilles, 2013; Catchpole et al., 2015).

Continued uplift and erosion will progressively remove the porphyry deposit itself, exposing deeper and deeper levels until the batholithic roots are revealed below 4–5 km depth. With the exception of young porphyry systems that have been emplaced into older uplifted and eroded batholiths (e.g., the 158–153 Ma San Carlos–Panantz-Satsu porphyry Cu-Mo deposits hosted by the mid-Jurassic Zamora batholith, Ecuador; Chiaradia et al., 2009b), batholithic terrains are generally unprospective for porphyry Cu deposits, because they represent the deep, hot, source levels of the system, not the shallow, cool, deposit level. Thus the Mesozoic Sierra Nevada and Peruvian Coastal Batholiths are largely unprospective for porphyry deposits. Erosional loss of upper-crustal porphyry Cu deposits increases with geological age (Kesler and Wilkinson, 2006, 2008; Wilkinson and Kesler, 2007, 2009) and in large part explains the increasing rarity of these deposits in Mesozoic, Paleozoic, to Proterozoic rocks, and their almost complete absence in the Archean. Alternative explanations for this rarity in older rocks in terms of different tectonic processes and atmospheric and/or oceanic oxidation state have been proposed by Evans and Tomkins (2011) and Richards and Mumin (2013a, 2013b).

These various factors further reduce the probability of discovering a mineable porphyry Cu deposit (reaching step 9 in Fig. 2) in any given arc segment to 0.0001%, with this probability decreasing further with age to almost zero in Archean rocks.

Supergene Enrichment

One additional geological factor can improve the odds of forming an economic porphyry Cu deposit, commonly spectacularly. During uplift and weathering, sulfide minerals such as pyrite and chalcopyrite oxidize and dissolve to form highly acidic ground-water. Normally this will result in the destruction of the orebody and dispersion of the metals. However, under some circumstances, the dissolved Cu may be reprecipitated below the water table as secondary chalcocite (Cu$_2$S), leading to supergene enrichment of the orebody below this level (Brimhall et al., 1985). This process operated particularly efficiently in the Paleogene arcs of northern Chile and southwest North America, where rapid erosion and deep weathering of Paleocene–Eocene porphyry systems uplifted them to near-surface conditions where thick supergene enrichment blankets formed in the Oligocene and Miocene (Alpers and Brimhall, 1988; Sillitoe and McKee, 1996; Arancibia et al., 2006). Continued uplift and erosion would likely have destroyed these transient weathering profiles, but in the early Miocene, uplift of the current High Andes began to the east, and, along with a change in ocean circulation patterns, led to the extreme...
aridification of the Atacama Desert region, termination of the weathering and enrichment process, and partial burial of the deposits by thick blankets of Atacama gravel (Dunai et al., 2005; Rech et al., 2006). Most porphyry Cu deposits in this region of Chile have undergone some degree of supergene Cu enrichment, but the Escondida orebody, discovered below gravel cover in 1981, was the most spectacular, when drill hole DDH61 intersected 250 m of chalcopyrite-enriched material averaging 3% Cu (pre-production reserves estimated to be 1.76 Mt at 1.59% Cu; Lowell, 1991; Ortiz, 1995; current measured sulfide resource is 5350 Mt at 0.63% Cu; BHP Annual Report, 2017).

Non-Geological Factors

One of the great frustrations for exploration geologists is that it is possible to find large and high-grade mineral deposits, but for them to prove impossible to mine (currently) for various logistical (access to transportation, power, and water), environmental, or political reasons. This is not the place to examine these problems in detail but just to note that such issues further reduce the probability of discovering a mineable ore deposit. Two recent examples illustrate the point. Large tonnages of high-grade mineralization were discovered in 2005 at the Pebble porphyry Cu-Au-Mo deposit in Alaska, with a measured and indicated resource of 5942 Mt at 0.42% Cu, 0.35 g/t Au, and 250 ppm Mo (Kelley et al., 2013); however, development has been blocked on environmental grounds. Similarly, a large high-grade resource was delineated in 2011 at the Reko Diq porphyry Cu-Au deposit in western Pakistan (Razique et al., 2014), with a mineable resource of 2200 Mt at 0.53% Cu and 0.30 g/t Au (Tethyan Copper Company, http://www.tethyan.com/the-reko-diq-project/reko-diq-resource/, accessed 1 November 2018), but development has been blocked on political and legal grounds.

■ SUMMARY

Convergent plate margins host a wide range of magmatic and hydrothermal ore deposits because of the mobilizing effect of subducted volatiles, especially $\text{H}_2\text{O}$, $\text{S}$, and $\text{Cl}$. Under moderately oxidizing conditions ($\Delta\text{FMQ} = +1$ to $+2$, typical of most arc magmas) chalcopyrite and siderophile metals dissolve preferentially as sulfide species in silicate magmas (forming “fertile” magmas for ore formation) but then strongly partition into Cl-rich hydrothermal fluids that inevitably exsolve from these hydrous magmas upon emplacement in the shallow crust. Under the right conditions, metals such as Cu, Mo, and Au may be deposited in concentrations that are economic to mine (porphyry Cu and related epithermal and skarn deposits), and these are referred to as ore deposits. However, such deposits represent end members of a spectrum of deposits that range through marginally subeconomic (at today’s metal prices, using current technology, and under current geopolitical conditions) to background concentrations. Such subeconomic deposits and geochemical anomalies are far more common than economic ore deposits and represent the normal products of arc magmatism.

In this analysis, I have reviewed the various steps in the process of generating an economic porphyry Cu deposit, and I show that Phanerozoic subduction zones are well suited to the formation of fertile magmas that have the potential to form such deposits (estimated to be 90% of arcs). However, a first-order control on ore-forming potential is the volume of arc magma generated, and this is a function of arc geodynamics. Long-lived, static arcs with normal subduction angles (~30°–45°) and moderately compressional stress states will generate large volumes of basaltic arc magma that become trapped by density contrasts at the base of the upper-crust plateau (MASH zones). Here the magma undergoes fractional crystallization and interaction with lower-crustal lithologies, evolving to intermediate calc-alkaline compositions. Perhaps 10% of arcs form large lower-crustal fertile MASH zones.

Evolved andesitic magma will tend to rise buoyantly until it pools again in the mid–upper crust as a second density barrier, although exsolution of volatiles to form low-density bubbly magma can drive it to erupt at the surface. Porphyry Cu deposits form from magmatic hydrothermal fluids exsolved below surface, and large deposits require large volumes of fluid and source magma (to supply metals). Consequently, conditions that promote the emplacement of large volumes of evolved magma in mid- to upper-crustal batholiths are essential for ore formation. Transpressional strain can localize magma ascent and promote plutonism, but this condition is transient and limits favorability to perhaps 10% of arcs, reducing overall probability of ore formation to 1%.

Assembly of a large mid- to upper-crustal batholith is not sufficient to form an ore deposit, however, because even though exsolution of most of the $\geq 4$ wt% $\text{H}_2\text{O}$ dissolved in the contained magma is inevitable as it cools and crystallizes, unless the flow of this fluid is focused and its release is rapid, metals will not be efficiently extracted from the batholith and channeled into narrow volumes of rock where their reprecipitation could generate an ore deposit. The formation of a structurally focused cupola zone and the sudden release of the bulk of the fluid contained in the batholith into this conduit are rare and stochastic events, perhaps triggered by a sudden random external process such as a mega-earthquake or volcano sector collapse. The low probability of such an event affecting a “primed” (volatile saturated or supersaturated) batholith with a well-formed cupola is low, perhaps less than 1% of such systems, which reduces the overall probability of ore formation to ~0.01%.

Multiple factors affect the efficiency of metal precipitation from hydrothermal fluids, and the probability of depositing economic concentrations of Cu ± Mo ± Au is likely less than 10% of well-plumbed hydrothermal systems. Such mineralization is typically precipitated 1.5–4 km below surface, which is too deep to mine (except for exceptionally high grade deposits with $>1.2\%$ Cu equivalent). Uplift and erosion is therefore required to bring these deposits nearer to the surface where they can be mined economically. This additional step reduces the overall probability of finding an economic porphyry Cu deposit in any given arc batholithic system to perhaps 0.0001%. However, continued uplift and erosion will begin to remove these shallow crustal ore deposits, such that their preservation in the geological record falls rapidly with age, to almost zero in the Precambrian.
LIMITATIONS OF THIS STUDY AND FUTURE DIRECTIONS

I fully acknowledge that the estimates of probability provided in this paper are little more than best guesses based on my personal experience—perhaps “educated guesses” would be a better description. It is not possible at this time to provide more accurate assessments, but I nevertheless think it is useful to place numerical estimates on the probabilities of the various steps, rather than simply saying “common” or “rare,” because this highlights the relative importance (or probability) of each step in the porphyry ore-forming process. In so doing, it also highlights which steps might deserve closer scrutiny to assess how they control or limit various processes, leading to more accurate quantification.

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