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A shake-up in the porphyry world?

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A shake-up in the porphyry world?

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Abstract

Porphyry Cu deposits form in the shallow crustal parts of arc magmatic systems, which root in the mantle wedge, evolve in lower crustal MASH zones (melting, assimilation, storage, homogenization) and lower-to-mid crustal hot zones, and accumulate in mid-to-upper crustal batholiths at depths of 5–10 km. A small proportion of the magma and most of the volatile load rises due to buoyancy towards the surface, and may erupt as volcanic or fumarolic emissions. Low levels of volcanism and fumarolic activity, as well as subsurface hydrothermal flow and alteration, are normal and semi-continuous features of active arc magmatic systems, which may operate for millions of years. Porphyry Cu deposits, on the other hand, form rarely (typically ≤1 per batholith) and rapidly (≤100,000 years) in the subsurface (2–5 km depth), where hydrous volatiles exsolved from the underlying batholith are channeled into structurally controlled cupola zones and cool before reaching the surface. The explosively brecciated character of early mineralization stages (breccia pipes and stockworks) suggests that the initiation of fluid flow

may be essentially instantaneous and catastrophic, with the longer total duration of hydrothermal activity reflecting slower kinetically controlled fluid exsolution processes, or draining of deeper parts of the system. These fluids generate intense subsurface hydrothermal alteration, and may precipitate economic concentrations of Cu-sulfide minerals in potassic alteration zones as they cool between ~400°–300°C.

The suddenness and infrequency of these ore-forming hydrothermal events suggests that they are triggered by an external process acting on otherwise normally evolving magmatic systems. Sudden depressurization or agitation of a large, primed, volatile-saturated or supersaturated mid–upper crustal magma chamber could lead to rapid and voluminous volatile exsolution and fluid discharge. This sudden volatile flux could result in either a large explosive volcanic eruption if the surface is breached, or a large magmatic-hydrothermal system that could form a porphyry Cu deposit if fluid flow is restricted to the subsurface. Candidates for triggers of these destabilizing events are catastrophic mass wasting such as volcanic edifice collapse, or mega-earthquakes, the latter possibly causing the former. The frequency of such catastrophic events occurring in proximity to active arc batholiths may approximate the recurrence rate of formation of large porphyry Cu deposits.

Introduction

Porphyry Cu±Mo±Au (hereafter simply porphyry Cu deposits) are one of the most studied and best understood mineral deposit types in the world, with benchmark studies by Lowell and Guilbert (1970), Gustafson and Hunt (1975), and Sillitoe (2010) being amongst the most cited papers in the economic geology literature. And yet a very fundamental question remains unanswered: What is the trigger for porphyry ore-forming events? We know that porphyry deposits are formed by the release of large volumes (≥10 km^3) of hydrothermal fluid from mid-to-upper crustal batholithic-scale magma chambers (≥100 km^3; Burnham, 1979; Cline and Bodnar, 1991; Richards, 2003, 2005), and numerous studies have shown that these magmatic systems can have lifetimes of several million years (e.g., Matzel et al., 2006). However, individual porphyry Cu deposits have lifespans that rarely exceed 100,000 years (e.g., Chiaradia et al., 2013), and are commonly unique events within the history of their associated batholith. In addition, the well-established vein paragenesis from high-temperature A-veins to low-temperature D-veins (*sensu* Gustafson and Hunt, 1975) is rarely repeated or reversed within individual deposits, indicating a fluid flow regime that evolves from hot to cold (e.g., Reed et al., 2013). Finally, although porphyry-type hydrothermal alteration systems are relatively common in arc volcanoplutonic complexes, mineralized systems, and especially large *economic* porphyry Cu deposits, are rare (by economic definition). So what sometimes triggers these large magmatic-hydrothermal ore-forming events at singular points in the much longer histories of arc magmatic systems?

Here I first review constraints on the *duration* of arc magmatic and ore-forming processes, and then consider the key question of *timing*. It turns out that, whereas the duration and mechanics of these processes are reasonably well understood, predicting *when* major events such as mega-volcanic eruptions or the formation of large porphyry Cu deposits will occur is extremely difficult. This is because these geologically sudden, singular events are stochastic, being the products of multiple cumulative and coincidental events, none of which are individually rare, but whose correct combination has a low probability of occurrence (Richards, 2013). In particular, it appears that large volcanic eruptions and large magmatic-hydrothermal events require an external trigger to push an otherwise fairly passively evolving mid-to-upper crustal magmatic system into a state of instability and sudden voluminous fluid exsolution. If that fluid flux drives magma to the surface, a violent explosive volcanic eruption will ensue; if the fluid flux is contained and channeled below surface, an ore deposit might be formed. Candidates for this triggering process include the impact of a mega-earthquake or sudden mass wasting events such as volcanic edifice collapse (the former perhaps prompting the latter), which could cause sudden depressurization and agitation of a volatile-saturated or supersaturated magma chamber, resulting in voluminous fluid exsolution and expulsion.

Porphyry Cu deposits: Timing is everything

Large (billion tonne) porphyry Cu deposits are globally associated with large batholithic scale arc plutonic systems (Burnham, 1979; Richards, 2003; Rohrlach and Loucks, 2005; Rezeau et al., 2016). They do not form within the batholith itself, which represents the source of hot fluids and metals at mid-to-upper crustal depths of 5–10 km, but rather they form in the shallow apical parts of the system at depths of 2–5 km, where fluids and bubbly magma are channeled towards the surface (the cupola zone; Burnham, 1979; Shinohara and Hedenquist, 1997; Cloos, 2001; Weis, 2015). Accurately measuring the age and duration of these magmatic and hydrothermal systems has been the focus of many recent field and analytical studies, using combinations of U-Pb zircon dating of igneous rocks, K-Ar and $^{40}Ar/^{39}Ar$ dating of hydrothermal minerals, and Re-Os dating of molybdenite in ore assemblages. The results are incontrovertible:

Magmatic duration: Batholiths of the size necessary to supply the volume of Cu in large porphyry deposits (≥100 km^3; 10 Mt Cu; Richards, 2005) are assembled in the mid- to upper crust on timescales of millions of years (Matzel et al., 2006; Miller et al., 2007; Fiannacca et al., 2017), while the magmatic lifetime of a single upper crustal pluton may extend for up to 10^6 yr (Matzel et al., 2006; Chiaradia et al., 2013; Kaiser et al., 2017). These durations are comparable to timescales for deep crustal heating and melting by underplating of basaltic magmas from the subduction zone (Petford et al., 2000; Annen and Sparks, 2002; Hawkesworth et al., 2004).

Porphyritic pluton duration: The typically small volume plutons and dikes found at shallower crustal (subvolcanic) levels and within the porphyry deposits themselves would have cooled and crystallized much more quickly ($\leq 10^5$ yr) and even more rapidly if convectively cooled by groundwater circulation (Norton, 1982; Cathles et al., 1997; Weis et al., 2012).

Porphyry ore-formation duration: The duration of an individual porphyry ore-forming event appears to be shorter than the precision of most of our geochronological techniques, and has been repeatedly shown to be on the order of 10^5 yr or less (Arribas et al., 1995; Marsh et al., 1997; Shinohara and Hedenquist, 1997; Weis et al., 2012; Chiaradia et al., 2013; Chelle-Michou et al., 2017; Mercer et al., 2015). Furthermore, many porphyry deposits are characterized by early high-temperature breccia and stockwork events that likely formed explosively (Sillitoe, 1985; Skewes et al., 2002; Vry et al., 2010; Harrison et al., 2018). This suggests that the initial event involved the essentially instantaneous escape of previously exsolved but physically trapped fluids (e.g., Christopher et al., 2015; Boudreau, 2016; Parmigiani et al., 2016; Edmonds and Wallace, 2017), with prolonged or pulsed flow of high-temperature fluids reflecting the slower kinetics of magmatic fluid exsolution, and the draining of fluids from deeper or peripheral parts of the underlying magmatic system (Candela, 1997; Wallace et al., 1999).

Lower temperature alteration may extend the overall hydrothermal duration to ~10^6 yr, likely reflecting the thermal (subsolidus) life of the underlying pluton (Goff et al., 1992; Cathles et al., 1997).

Cyclicity of ore-formation: Detailed paragenetic studies of numerous porphyry deposits worldwide (e.g., Cannell et al., 2005; Redmond and Einaudi, 2010; Vry et al., 2010; Sepp and Dilles, 2018) have confirmed the vein paragenesis originally defined by Gustafson and Hunt (1975) for the El Salvador porphyry Cu deposit in Chile. Early dark mica (EDM) and sinuous, deformed (ductile) quartz-K-feldspar-anhydrite-sulfide A veins are formed from high temperature fluids associated with intense potassic (K-feldspar, biotite) alteration, and are generally only weakly mineralized. These early veins are crosscut by linear, parallel-sided (brittle), quartz-anhydrite-chalcopyrite-molybdenite B veins deposited from lower temperature fluids (450°–350°C) with potassic alteration, which, together with disseminated sulfide mineralization, account for the bulk of ore in most deposits (Redmond et al., 2004; Landtwing et al., 2005; Klemm et al., 2007).

A and B veins are in turn cut by pyritic D veins, with minor quartz and anhydrite, and feldspar-destructive sericitic (phyllic) alteration halos that may link to affect large volumes of rock. This alteration stage forms around and above the potassic zone, and overprints it downwards and inwards as the hydrothermal system begins to cool and collapse back on itself. The phyllic zone is generally barren except where some residual Cu from earlier potassic alteration is preserved, or where molybdenite precipitates in muscovite-stable assemblages (Westra and Keith, 1981). Fluids in D-veins are typically lower temperature (350°–250°C; Landtwing et al., 2005; Harris et al., 2005).

The phyllic zone grades upwards into clay-stable assemblages (argillic and advanced argillic) which are formed from the condensation of increasingly acidic low density magmatic vapors, and are generally barren (except where high-sulfidation sulfide assemblages occur in advanced argillic alteration zones, but these are relatively rare) (Deyell et al., 2004; Harrison et al., 2018).

The entire system is surrounded by a propylitic alteration envelope (chlorite-epidote-carbonate) formed by ground water circulation heated by the underlying intrusions (Norton, 1982; Weis, 2015), with variable contributions from magmatic fluids (Cooke et al., 2014; Wilkinson et al., 2015). Propylitic alteration begins with initial magmatic emplacement, and continues to the end of the thermal life of the igneous system. As such it is coeval with (but peripheral to) early high-temperature potassic alteration, but also overprints this alteration as the externally circulating fluid invades the collapsing magmatic-hydrothermal system (Sheppard, 1977; Beane and Titley, 1981).

Arribas et al. (1995), Hedenquist et al. (1998), and Reed et al. (2013) have shown that cooling and depressurization of a single high-temperature magmatic-hydrothermal fluid can account for all of the alteration styles observed in typical porphyry deposits as described above, and in theory a single vein could be traced upwards through these alteration styles. However, there is also a temporal aspect to the evolution, such that as the initial magmatic-hydrothermal fluid flux wanes, the acidic alteration fronts collapse back inwards and downwards, overprinting earlier higher-temperature alteration and mineralization. At the same time, cooler external ground waters also invade the system.

Reed et al. (2013, p. 1379) asserted that there is a “universally observed sequence of vein cutting relations in porphyry copper deposits”. Reversals in this sequence are almost never observed, and repetitions of the paragenesis are relatively uncommon. Exceptions where repetition of the high-temperature vein stages have been recorded, perhaps related to pulsed magmatism, include the Boyongan and Bayugo porphyry Cu-Au deposits, Philippines (Braxton et al., 2018), El Teniente, Chile (Cannell et al., 2005), and Northparkes, Australia (Lickfold et al., 2003). The normal sequence is interpreted to represent a single hydrothermal event that starts hot and progressively cools as fluid flow wanes. Repetitions occur in the form of multiple mineralized centers in some of the largest porphyry districts, but these are typically separate intrusive–hydrothermal events with their own unidirectional paragenetic sequences, occurring at different times and mostly in different places within the overall life and spatial extent of the magmatic system (e.g., Richards et al., 2001; Lickfold et al., 2003; Cannell et al., 2005; Vry et al., 2010; Braxton et al., 2018).

Arc magmatic systems: Longer is better

Large porphyry Cu deposits tend to form relatively late in the history of arc magmatic cycles (Hine and Mason, 1978; Richards et al., 2001; Richards, 2003; Rohrlach and Loucks, 2005; Rezeau et al., 2016). This observation likely reflects the need for the arc to evolve towards more felsic, volatile-rich (H_2O, S, Cl), and oxidized magmatic compositions, all of which are prerequisites for upper crustal emplacement and exsolution of a voluminous ore-forming magmatic-hydrothermal phase (Burnham, 1979; Burnham and Ohmoto, 1980; Candela, 1992; Richards, 2003, 2011; Rohrlach and Loucks, 2005). Recent studies have shown that it may take up to 10 m.y. to establish a trans-crustal magmatic system that can deliver evolved magmas in volume to upper crustal levels (Whattam and Stern, 2016), and perhaps as long as 50 m.y. before steady-state thermal conditions are established in the arc crust (Rees Jones et al., 2018). Ardill et al. (2018) describe this process in terms of magmatic focusing, whereby transcrustal magma flow, mid–upper crustal pluton assembly, and volcanism become spatially focused (10^2–10^5 km^2) over timescales of 10^5–10^7 yr. These scenarios require that the axis of magmatism remains fixed, but subduction zones are dynamic systems, with rapid and frequent changes in the angle of subduction or subduction polarity, leading to arc migration or cessation. Early termination of arc magmatism is likely one of the most common reasons for the lack of development of large porphyry systems in any given arc segment.

The need to emplace a large volume (≥100 km^3) of fertile (volatile-rich, moderately oxidized) magma in the mid-to-upper crust without excessive eruption and venting to the surface also implies specific tectonic conditions that favor plutonism over volcanism. Transpressional strain can localize vertical magma ascent and mid–upper crustal pooling (batholithic plutonism), whereas compressional stress will tend to trap magmas in the deep crust, and extensional structures will allow magmas to rise directly to the surface (Brown, 1994; Tosdal and Richards, 2001; Richards, 2003; Chaussard and Amelung, 2014). Thus, changes in subduction zone dynamics leading to changes in upper plate stress conditions, following a prolonged period of compressional tectonics that built up a large deep-crustal magma volume, may be a first-order control on voluminous mid–upper-crustal plutonism that could source a large magmatic-hydrothermal system (e.g., Barton, 1996; Richards et al., 2001; Skewes et al., 2002; Cooke et al., 2005; Rohrlach and Loucks, 2005; Rezeau et al., 2016). However, this is still a *precondition* for ore-formation, and not necessarily a trigger for the actual ore-forming event.

Mid–upper crustal magma chambers: Stability to instability

The evidence from studies of batholithic plutonism and related volcanism is that, provided the flux of magma from depth is sustained, mid–upper crustal magmatic activity can continue over periods of many millions of years. Pulsed or semi-continuous magma influx activity is suggested by peaks in pluton emplacement and volcanic flare-ups, which may repeat on million-year cycles (Glazner et al., 2004; Matzel et al., 2006; Fiannacca et al., 2017; Kaiser et al., 2017; Pritchard et al., 2018). In between these pulses of activity, when recharge by hotter more primitive magma from the deep crustal MASH zone (melting, assimilation, storage, and homogenization; Hildreth and Moorbath, 1988) or deep-to-mid-crustal hot zones (Annen et al., 2006) wanes, mid-to-upper crustal plutons will either completely solidify (if small, or if the hiatus is too long), or will stagnate as largely crystallized mushes with small volumes of residual interstitial melt (Eichelberger et al., 2006; Klemetti, 2016). Recharge of these magma chambers by more primitive magma may trigger explosive eruption of residual felsic melts (Snyder, 2000; Eichelberger et al., 2006; Schubert et al., 2013; van Zalinge et al., 2017), or lead to hybridized intermediate-composition magmatism (diorite–granodiorite plutonism, andesite–dacite effusive volcanism; Eichelberger et al., 2006; Zellmer et al., 2012; Bergantz et al., 2015).

Volcanologists debate the processes that trigger massive explosive eruptions of felsic magma versus those that lead to more passive plutonism and effusive volcanism (Cashman and Sparks, 2013; Sparks and Cashman, 2017; Wilson, 2017). Overpressuring of a subvolcanic magma chamber by fluid exsolution (Eichelberger, 1995; Stock et al., 2016; Chelle-Michou et al., 2017; Edmonds and Wallace, 2017; Tramontano et al., 2017) or gas injection (Caricchi et al., 2018), perhaps linked to mafic magma recharge (Caricchi et al., 2014; Putirka, 2017), depressurization by catastrophic mass wasting, such as volcanic edifice collapse (Pinel and Jaupart, 2003; Voight et al., 2006; Roman and Jaupart, 2014), or seismic shaking (Walter, 2007; Namiki et al., 2016; Avouris et al., 2017; Nishimura, 2017) have all been suggested as triggers for cataclysmic explosive volcanism.

The build-up of dissolved volatile content to the point of saturation and exsolution is inevitable in already volatile-rich arc magmas as they depressurize, cool, and crystallize anhydrous minerals during ascent through the crust. Mafic to intermediate composition magmas containing 4–6 wt.% H_2O will saturate in water at pressures of ~2 kb, or depths of ~5–10 km under lithostatic pressure conditions (Burnham, 1979; Burnham and Ohmoto, 1980). This is the depth at which mid–upper crustal batholiths assemble, and implies that the magmatic volatile phase originates from batholithic volumes of magma, well below the levels of small subvolcanic plutons and dikes, and also below the level of porphyry ore formation (2–5 km).

The separation of a volatile phase will either result in expansion of the melt-crystal-bubble mixture, lowering its density and driving it towards an explosive eruption at the surface (Eichelberger, 1995), or will increase the magma chamber pressure if trapped below surface (Burnham, 1979; Burnham and Ohmoto, 1980; Snyder, 2000).

Christopher et al. (2015), Sparks and Cashman (2017), and Cashman et al. (2017) have suggested that mid–upper crustal magma chambers cycle between dormancy, unrest, and instability (eruption), controlled by gravitational instabilities caused by localized accumulation of volatiles (e.g., Turner et al., 1983; Pritchard et al., 2018). Overturn and coalescence of lenses of bubble-rich crystal mush may lead to the rapid ascent and expulsion of large volumes of low-density, bubbly magma. Sparks and Cashman (2017) suggest that this may explain the periodicity of large explosive volcanic eruptions. However, if the fluids and magma cannot vent readily to the surface, then an alternative possibility is that these instability events could generate pulses of subsurface hydrothermal fluid flow, which may lead to ore formation if other factors align.

Subsurface venting, hydrothermal alteration, and (sometimes) ore-formation

Volcanic rocks and shallowly emplaced porphyry systems (<1 km depth) are typically poorly mineralized, because fluids exsolved at low pressure are mostly low density vapors (with only small volumes of high density brine or even solid salt) that have little capacity to transport metals (Cline and Bodnar, 1991; Muntean and Einaudi, 2000). In contrast, most porphyry Cu deposits form at depths of 2–5 km, from supercritical saline fluids or liquids coexisting with higher density vapors that can efficiently transport Cu, Mo, Au, and other metals (Cline and Bodnar, 1991; Redmond et al., 2004; Landtwing et al., 2005; Rusk et al., 2008). These fluids in turn are derived from volatile-saturated magmas in the underlying source magma chamber at depths of 5–10 km, as described above. An efficient magmatic-hydrothermal system that could precipitate economic porphyry-type mineralization at 2–5 km depth will therefore only form if fluid released from a deeper source magma chamber is not vented directly to surface in an explosive eruption, but is instead channeled towards the surface under a confining pressure and along a steep geothermal gradient, cooling to below 300°C at shallow depths (≤1 km; Shinohara and Hedenquist, 1997). However, the question remains as to what conditions might allow a sudden, voluminous release of fluid (on time scales of $\leq 10^5$ yr) from this deep magma chamber, which has otherwise been evolving and degassing relatively passively on timescales of $\geq 10^6$ yr? Could the conditions, and triggers, be similar to those described above for large-scale explosive volcanic eruptions, with the difference being that the fluid emission was largely trapped below surface instead of venting?

Under the dormant or passive states described by Sparks and Cashman (2017), volatile exsolution in the batholith can be expected to be a slow, continuous process as magmas progressively cool and crystallize. Bubbles of low-density supercritical magmatic-hydrothermal fluid will coalesce and slowly escape upwards towards the surface, initially by forming channels between crystals in the magma mush (Candela, 1997; Boudreau, 2016), then along hydraulic fractures propagated by their own fluid pressure, and finally through joints and fracture networks in the brittle overlying cover rocks. If there is no focusing of this fluid flow, it can be expected to cool rapidly due to wallrock interaction, and cause widespread but weak hydrothermal alteration (perhaps indistinguishable from groundwater-generated propylitic alteration; e.g., Pritchard et al., 2018), and little or no mineralization (Fig. 1a). But if instead fluids are released in a sudden pulse, perhaps in response to an instability of the type described by Sparks and Cashman (2017), and if the ascent of this fluid pulse is focused into an apical part of the batholith and then into a narrow overlying cupola zone, then conditions may be favorable for ore formation (Shinohara and Hedenquist, 1997; Weis et al., 2012; Weis, 2015).

Such sudden fluid flow events may be marked initially by diatreme or breccia pipe formation, as initial high fluid pressures blast a pathway through overlying rocks (Fig. 1b; Sillitoe and Sawkins, 1971; Norton and Cathles, 1973; Burnham, 1985; Sillitoe, 1985; Vry et al., 2010; Harrison et al., 2018). Bubbly magma and large volumes of fluid will then flow along these conduits as the entire underlying magma chamber decompresses and more slowly devolatilizes (Fig. 1c), consistent with the observation that diatremes and breccia pipes are commonly mineralized immediately *after* initial emplacement, and intruded by dikes of the same generative magma (Fig. 2; Sillitoe and Sawkins, 1971; Sillitoe, 1985; Anderson et al., 2009; Richards, 2011; Large et al., 2018). This implies that the onset of hydrothermal activity in porphyry systems may be a sudden, even seismic event, triggered by explosive release of magmatic fluid pressure into shallower, lower-pressure environments above the brittle–ductile transition zone (400–350°C; Fournier, 1999; Gorczyk and Vogt, 2018). It is important to note that many hydrothermal breccia pipes rooted in magmatic breccias do not extend to the surface, indicating that fluid flow was restricted to the subsurface (Sillitoe and Sawkins, 1971; Sillitoe, 1985; Anderson et al., 2009). Gorczyk and Vogt (2018) also suggest that this process may explain the narrow cylindrical or pipe-like shape of many porphyry systems ("pencil-porphyries"; Norton, 1982; Skewes et al., 2002; Lickfold et al., 2003).

A seismic trigger for ore-forming magmatic-hydrothermal events?

Sparks and Cashman (2017) did not speculate on what actually tips a dormant magmatic system into an unstable state, although they implied that it may result from progressive build up of low-density volatile-rich lenses within the magma chamber, which eventually overturn gravitationally (Turner et al., 1983). They also imply that recharge by fresh, volatile-rich mafic magma contributes to this process. Indeed, recharge is essential if the magma chamber is not to simply solidify. However, recharge alone is unlikely to be the actual trigger for instability, not least because recharge must occur almost continuously throughout the history of the batholith (as noted above), and is not a singular, short-lived event like a major volcanic eruption or the formation of a porphyry ore deposit (e.g., Putirka, 2017).

It seems more likely that a sudden external event triggers this process, by acting serendipitously on a primed, volatile-saturated or -supersaturated mid-upper crustal magma chamber (Tramontano et al., 2017). This event could lead either to a major explosive volcanic eruption, as envisaged by Sparks and Cashman (2017), or to a major subsurface magmatic-hydrothermal event as proposed here. These two processes are closely related and may occur at different times in the same magmatic system, but are antithetical in the sense that surface venting is the opposite of subsurface hydrothermal circulation (e.g., Buret et al., 2017).

Candidates for this triggering process must be sudden, large-scale events, which are normal but infrequent occurrences in volcanic arcs, happening only a few times over the life of a batholith (i.e., with a recurrence rate of once every 10^5–10^6 yr). Two possibilities, already identified as potential triggers of explosive volcanism, are catastrophic mass wasting events, such as volcanic edifice sector collapse, and M>9 mega-earthquakes (which may themselves trigger mass wasting events). Sudden unloading of a magma chamber by edifice collapse (e.g., Voight et al., 2006), or seismic shaking of supersaturated melt or a melt-crystal-bubble mush (Davis et al., 2007; Cannata et al., 2010; Namiki et al., 2016; Avouris et al., 2017) may trigger a sudden pulse of fluid exsolution and hydrothermal activity. Analogues are found in the way that seismic shaking can trigger mud volcanoes and geothermal fluid flow (Manga and Brodsky, 2006), and suddenly depressurizing a shaken can of beer results in explosive effervescence (Rodríguez-Rodríguez et al., 2014). Seismic pressure release has also been proposed as an important mechanism for gold deposition in epithermal and mesothermal gold deposits, due to vapor exsolution (Sibson et al., 1988).

Volcanic sector collapse is a relatively common feature of arc volcanoes (Francis and Wells, 1988), historically occurring at ~4 events per 100 yr. globally (Siebert, 1984), and perhaps at least once over the life of a typical multi-million-year-old stratovolcano and underlying batholith (e.g., Cantagrel et al., 1999). Similarly, M>9 earthquakes occur 1–3 times per 100 yr. globally (McCaffrey, 2008). Given that these are individually stochastic events, and furthermore that they would need to coincide spatially and temporally with a primed batholith, their probability of occurrence may be comparable to the observed frequency of formation of large porphyry Cu deposits globally in the Mesozoic–Cenozoic (1–2 per m.y.; Singer et al., 2008). (Note that

Wilkinson and Kesler, 2009, predict a higher rate of 244 per m.y. if undiscovered and eroded deposits are included.)

Thus, it is concluded that an external trigger such as a mega-earthquake and/or catastrophic mass wasting, acting on a primed, volatile saturated or oversaturated magma chamber, may explain the sudden, random, and generally singular formation of large magmatic-hydrothermal systems and associated porphyry Cu deposits in otherwise unmineralized arc magmatic systems. Similar ideas have been mooted in the past by Sillitoe (1994) and Mpodozis and Cornejo (2012), and it is suggested that a fruitful avenue of research would focus on the effects of seismic devolatilization of hydrous magmas leading to explosive volcanism and/or subsurface hydrothermal ore formation. These are truly stochastic events, with the successful eruption of a large volcanic plume, or the formation of a large subsurface porphyry Cu deposit, depending not only on the efficient and maximized operation of a sequence of processes, from mantle magma generation to fluid exsolution and focusing, but also on the serendipitous timing of an external trigger that will tip an otherwise relatively passively evolving magmatic system into a state of sudden devolatilization.

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Figure captions

Figure 1. From steady state degassing to unstable fluid expulsion. (a) A primed (vapor-saturated or supersaturated) magma chamber undergoing steady state volcanism and degassing. (b) An external trigger such as volcano sector collapse or mega-earthquake tips the system into an unstable state, resulting in massive fluid exsolution and expulsion. (c) If fluid flow is largely contained below surface, it will cause intense hydrothermal alteration and possibly ore deposition.

Figure 2. Magmatic-hydrothermal breccia textures from the Pachapaqui Ag-Zn-Pb-Cu mine, Huaraz province, Peru. (a) Magmatic-hydrothermal breccia with clasts of wallrock and dacite porphyry, subsequently intruded by the same porphyry magma which generated the breccia; late cavity space partially filled by vuggy hydrothermal quartz. Photograph of adit wall. (b) Magmatic-hydrothermal breccia with “live” clasts of dacite porphyry magma, showing rounded, cuspate, or “amoeboid” shapes indicating that the clasts were still partially molten when incorporated into the breccia, presumably by explosive disaggregation of the same magma. The breccia in this location is unmineralized (DDH101). (c) (Slightly) later hydrothermal alteration and sulfide mineralization overprinting a different part of the same magmatic-hydrothermal breccia pipe (DDH28).

Figure 1

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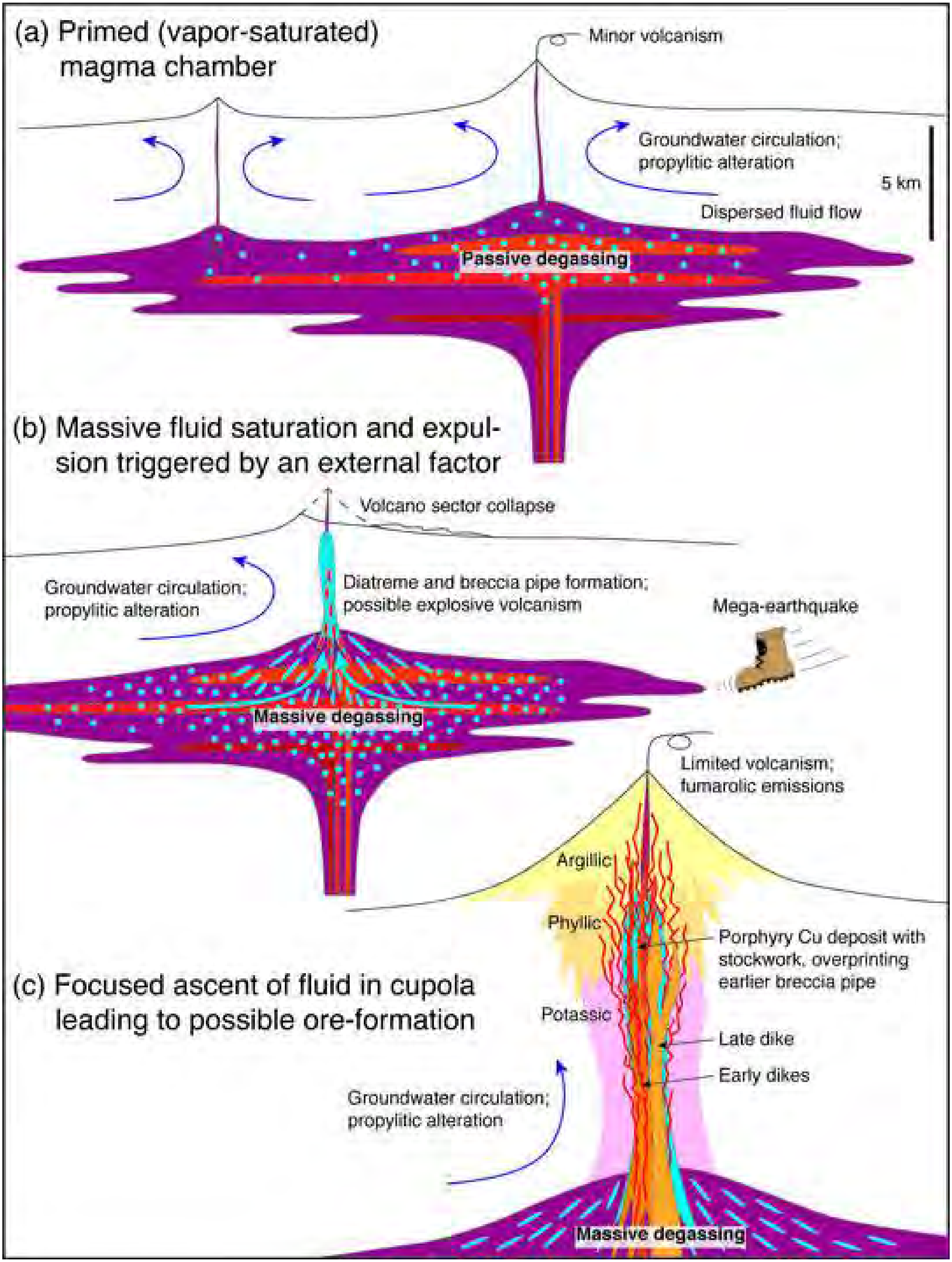
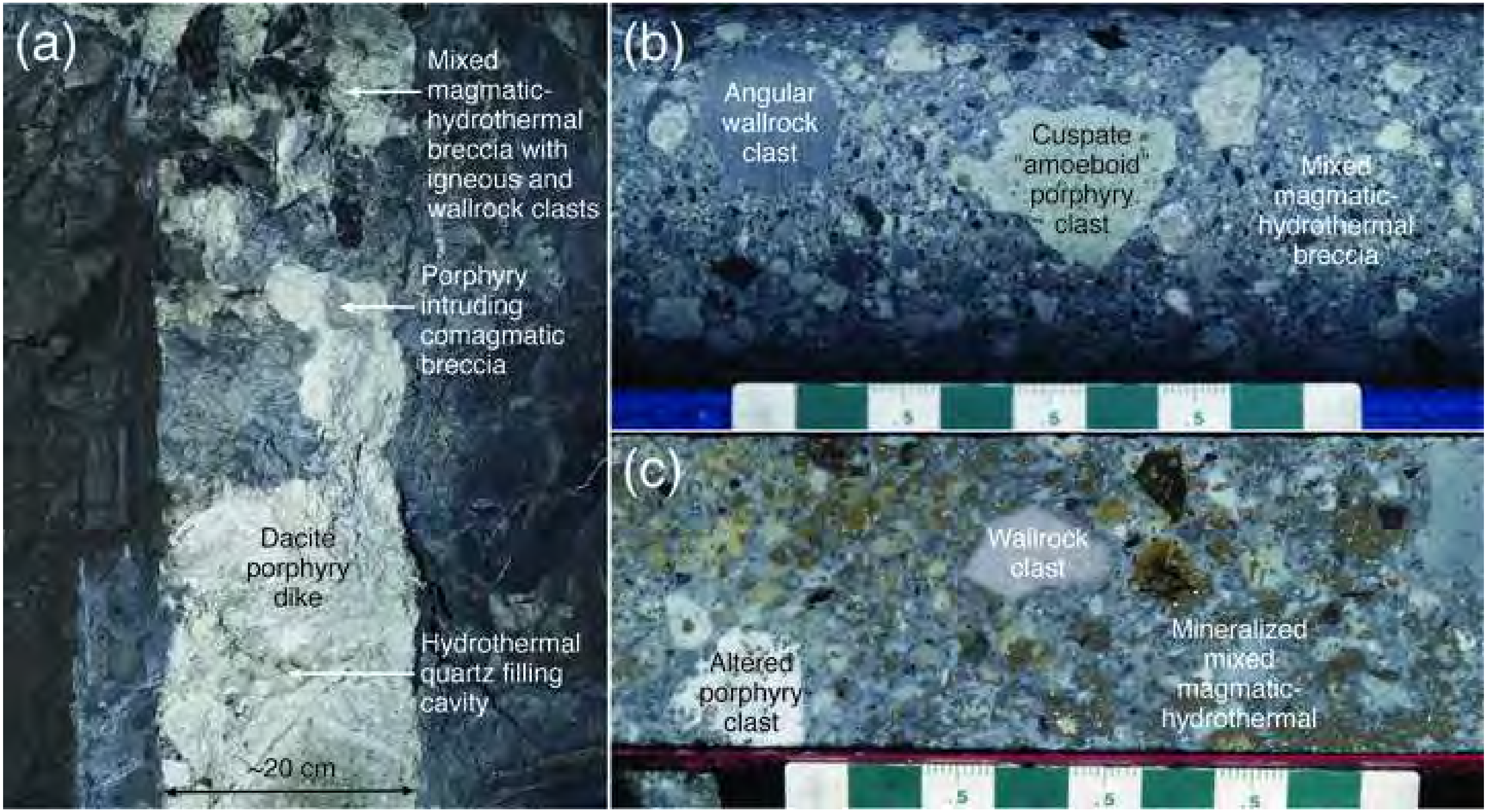


Figure 2


(a)
Mixed magmatic-hydrothermal breccia with igneous and wallrock clasts
Porphyry intruding comagmatic breccia
Dacite porphyry dike
Hydrothermal quartz filling cavity
~20 cm
(b)
Angular wallrock clast
Cuspate "amoeboid" porphyry clast
Mixed magmatic-hydrothermal breccia
(c)
Wallrock clast
Altered porphyry clast
Mineralized mixed magmatic-hydrothermal

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