Vertically extensive and unstable magmatic systems: a unifying view of igneous processes associated with volcanoes

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ABSTRACT

Volcanoes are an expression of their underlying magmatic systems. Over the past three decades, the classical focus on upper crustal magma chambers has expanded to consider magmatic processes throughout the crust. A trans-crustal perspective must balance slow (plate tectonic) rates of melt generation and segregation in the lower crust with new evidence for rapid melt accumulation in the upper crust prior to many volcanic eruptions. Reconciling these observations is engendering active debate about the physical state, spatial distribution and longevity of melt in the crust. Here we review evidence for trans-crustal magmatic systems and highlight physical processes that might affect the growth and stability of melt-rich layers, focusing particularly on conditions that cause them to destabilize, ascend and accumulate in voluminous but ephemeral shallow magma chambers.

ONE SENTENCE SUMMARY: Magmatic systems traverse the crust, are dominantly crystalline and are inherently unstable; this view has great explanatory power but requires new conceptual models for addressing questions about magma evolution and the behaviour of volcanic systems.

INTRODUCTION

For more than 100 years, a classic concept for volcanism and igneous processes has been a melt-dominated magma chamber (Fig. 1a) that supplies volcanoes during eruption (*1*). The model commonly considers the magma chamber to be long-lived, usually shallow, and to solidify eventually to form igneous plutons such as the silicic Sierra Nevada batholith and the mafic Stillwater, Skaegaard and Bushveld intrusions. The magma chamber is reproduced in simplified forms in introductory textbooks and popular science conceptions of volcanic activity, as illustrated by the Wikipedia entry: "A magma chamber is a large underground pool of liquid rock found beneath the surface of the Earth".

Key igneous phenomena that have been viewed through a magma chamber prism include geophysical investigations of magmatic systems, igneous differentiation, volcano evolution, volcanic unrest, caldera collapse and ore deposit formation. From a hazards perspective, the magma chamber paradigm remains central to many conceptual models of volcanic eruptions and, as a consequence, is a common starting point for interpreting geophysical signals of unrest (*2*). Over the past few decades, however, accumulating evidence from geophysical, geochemical, petrological, volcanological and geological observations is proving enigmatic, puzzling or inconsistent with the classic magma chamber paradigm. At the same time, physical models of igneous processes based on thermal, mechanical and dynamical principles suggest that classical magma chambers are surprisingly difficult to form and to maintain. For this reason, igneous processes are increasingly viewed as occurring throughout the crust, with the magma chamber occupying only the top of a much larger trans-crustal magmatic system.

Processes in the larger system lead to magma accumulation in the shallow crust. This perspective has developed into the concept of open, rather than closed, magma chambers that can be modified by both crustal assimilation (*3*) and magma replenishment from greater depths (*4*). Magma replenishment results in various mixing phenomena (*5*) and provides one possible triggering mechanism for volcanic eruptions (*6*). Looking deeper, thermal models suggest that the lower crust is the most favourable location for initial stages of fractionation and melt segregation (*7, 8*), and exposed crustal sections show a continuum from mafic cumulates in the deep crust to granitic intrusives in the upper crust (*9*). In this way, conceptual models of magmatic systems have evolved to encompass the entire crust (Fig. 1b).

Another emerging consensus is that trans-crustal magmatic systems (TCMS) are dominated by crystal mush. Here we use the term mush to refer to any system of crystals and melt in which the crystals form a continuous framework through which melt is distributed (Fig. 1c). Thus mushes are by definition at or above the solidus and are synonymous with partially molten rock. Mushes in their entirety are not eruptible (*10*) because their rheology is controlled by deformation of the crystalline framework. For this reason, the deformation behaviour of mushes is sensitive to the absolute melt fraction. This is particularly true below about 0.07, which defines the melt connectivity transition, below which the strength of the crystal-melt aggregate increases strongly with decreasing melt fraction (Fig. 1c inset; (*11*)). Mush fragments, however, can be erupted as crystal clots (glomerocrysts), cumulate nodules or restite (Fig. 2). Eruptible magma, in contrast, is defined as a suspension of crystals in melt (+/- exsolved volatiles). The transition from magma to mush takes place over a narrow range in crystal content (*12*), where the bulk viscosity increases by several orders of magnitude and the crystal-melt suspension develops strongly non-Newtonian properties (Fig. 1c). Typically the rheological transition occurs within the window of 50-65%, but may occur over a narrower range for a specific system (*13, 14*).

A mush-dominated view of igneous systems has great explanatory power, but requires new conceptual models for addressing questions related to both magma evolution and the behaviour of volcanic systems. In particular, it shows that TCMS models need to address both the chemistry and physics of multiphase (crystal-melt-volatile) materials. Although such systems are quite easy to conceptualise, they are remarkably difficult to model numerically, not only because of the number of phases, but also because those phases react with each as the system evolves: melts crystallise, crystals dissolve and gases exsolve. We describe components of such a model to illustrate that, by connecting processes throughout the entire TCMS, it should be possible to (1) relate physical processes of crystal-melt-fluid segregation to magmatic differentiation, (2) anticipate interactions between melts and fluids generated and stored at different depths, and (3) determine the physical mechanisms by which magma and associated fluids move through the crust, accumulate in shallow chambers, and then erupt. These processes represent internal controls on volcanic eruptions. Although external controls (e.g., tectonic, isostatic, hydrothermal interactions) also influence melt accumulation (*15*) and eruption (*16, 17*), here the focus is on internal processes.

EVIDENCE FOR COMPLEX AND VERTICALLY EXTENSIVE MAGMA STORAGE REGIONS

The concept that magmatic systems extend beneath upper crustal magma reservoirs is not new; in fact, the geochemical case for magma accumulation and processing in the deep crust was made elegantly almost thirty years ago *(7)*. More recent thermal and physical models *(8)* suggest extensive igneous differentiation in the deeper and hotter parts of the crust. Melt storage in the lower crust may partially explain a persistent puzzle in studies of magmatic systems, which is the inability of geophysical imaging techniques to identify large volumes of melt in the subsurface, even in regions that are currently active or have produced very large eruptions in the Quaternary. These volcanic systems are commonly underlain, however, by electrically conductive zones of low seismic velocity that are certainly hotter than the surrounding crust, and probably contain melt. At Yellowstone caldera, USA, for example, tomographic studies have identified a large (~ 10,000 km^3) low velocity body at 5-17 km depth, which is underlain by an even larger (46,000 km^3) lower crustal anomaly (*18, 19*). With estimated melt contents of 9% and 2%, respectively, the upper and lower mush zones may each contain hundreds of km^3 of melt, more than enough to generate another super-eruption. Even more extensive is the Altiplano Puna Magma Body in the central Andes, where recent geophysical surveys have identified a low velocity zone in the mid-crust (<25 km depth) that may comprise up to 500,000 km^3 of melt-rich (≤25%) mush (*20, 21*).

The geometry of melt distribution in lower and mid-crustal reservoirs is not well constrained. Melt fraction estimates based on tomography or MT images are averaged at resolutions that are typically no better than about 1 km, so melt-rich regions smaller than this scale cannot be detected. However, a plausible notion, consistent with a large body of physical theory (*22-24*), is that melt is heterogeneously distributed in the lower crust and includes (microscale) melt distributed along grain boundaries, mesoscale variations in melt concentration created by compaction, and large-scale, vertically stacked melt-rich lenses. The latter have been inferred from a recent tomographic study of the very large Toba caldera, Indonesia (*25*) and observed in regions of extension, such as mid-ocean ridges (*26*), Iceland (*27*) and continental rifts (*28*).

The inability of geophysical methods to identify large melt-rich bodies in the upper crust, in contrast, suggests that large volumes of upper crustal melt are likely ephemeral. In extensional environments such as mid-ocean ridges, identified melt lenses are thin and sill-like in shape. In arc environments, large regions of possible melt accumulation appear limited to the mid-crust, where they may feed multiple volcanoes (*20, 29, 30*). Between the mid-crust and individual volcanoes, observed low velocity zones are typically narrow and vertically elongated, averaged melt estimates are <10%, and exsolved volatiles may be important at shallow levels (*31-33*).

Conditions of magma storage can also be inferred from the compositions and textures of the erupted material (lava and pyroclasts). Phase compositions and proportions can be matched to pre-eruptive magma storage conditions using phase equilibria experiments (*34-36*), and bulk magma compositions are commonly used to track magma evolution by crystal fractionation and/or assimilation *(37-39)*. Growing evidence for extensive entrainment of crystals throughout the spatial extent of the magmatic system (*40*), however, suggests that these methods do not provide sufficient information to fully characterize most magmatic systems. The diversity of crystal 'cargo' contained within many volcanic rocks is nicely illustrated by the most common igneous mineral plagioclase, where the crystallization history is preserved in strikingly complex compositional zoning patterns that appear as shades of gray in the backscattered electron image shown in Figure 3a. Measured compositional and trace element variations in a single crystal can then be modeled using data from an experimentally determined phase diagram (Fig. 3b) to show that, in this example, the crystal core resided in a cooler but deeper part of the magmatic system before being entrained by hotter melt that transported it to a shallow temporary storage region and, ultimately, to the Earth's surface (*41*). More generally, abundant evidence for diverse pre-eruptive histories of neighbouring crystals within individual samples requires crystals to be assembled from different parts of the subvolcanic system (*41-45*), often shortly before eruption (*46-48*). These crystals may derive from either cooler marginal zones (*49-51*) or deeper and hotter parts of the system (*52, 53*). The complex history of the crystal cargo is further illustrated by variations in isotopic compositions within and between individual crystals (*40, 54-57*). These data show that different crystals in a sample, or even different zones within a crystal, must have grown from isotopically distinct melts. It is hard to envisage such small-scale isotopic heterogeneity existing within a large and continuous body of melt.

Further insight into the nature of magmatic storage systems can be found in measured timescales of magmatic differentiation, crystal growth and (pre-eruptive) residence time in the transporting melt (*58-61*). Timescales of magma differentiation can be estimated by dating zircon crystals, which are sufficiently resilient to be recycled between individual magma batches. Zircon data suggest differentiation timescales of 10^3 to $>10^5$ years, which contrast sharply with the much shorter timescales (< 1 to $\sim 10^3$ yrs) calculated for magma accumulation in the upper crust prior to volcanic eruptions (*62-67*). This dichotomy is illustrated by the example from Mount St. Helens shown in Figure 3. Here isotopic constraints show that entrained plagioclase crystals grew over thousands of years (Fig. 3c; (*68*)), while zircon crystals derive from magmatic activity tens or hundreds of thousands of years prior to the 1980 eruption (Fig. 3d; (*69*)). Diffusion time scales for magma assembly prior to the 1980 eruption, in contrast, are on the order of months to years (*70*), which are commensurate with the months of observed pre-eruptive unrest. These timescales can be reconciled only if crystals with different histories, stored in different parts of the magmatic system, are transported to the growing upper crustal magma chamber and amalgamated shortly prior to eruption.

The depth range of pre-eruptive magma storage can be estimated using the dissolved volatile content of crystal-hosted melt inclusions (*53, 71, 72*). Crystal-hosted melt inclusions from extensional environments show shallow magma storage over a limited pressure range (Fig. 4); this is true for both a young submarine basalt flow from the Galapagos ridge (GAL) and a large silicic eruption in the Taupo Volcanic Zone, New Zealand (OR). In contrast, melt inclusions from arc volcanoes commonly record a larger pressure range, as illustrated by data from Etna volcano, Italy (ETNA) and Popocatepetl volcano, Mexico (POPO), Soufrière Hills Volcano, Montserrat (SHV) and Mount St. Helens, USA (MSH). Additional pressure estimates obtained from geobarometry of deep-crystallizing pyroxenes (SHV) and amphiboles (*73*) suggest that many magmatic systems extend at least to the mid-crust. Additional evidence for magmatic systems that extend from the upper to mid-crust comes from seismic data collated from eruptions over the past few decades (*74, 75*), which cover a similar depth range (Fig. 4a).

Measurements of dissolved volatiles provide only a minimum estimate of magma storage pressure. First, melt inclusions record only the part of the magmatic history over which crystals are growing and trapping inclusions. Second, not all volatiles are dissolved in the melt prior to eruption. CO_2, for example, is relatively insoluble in silicate melts and will start to exsolve at high pressures (*76*). Additionally, sulphur dioxide, which is much more soluble, also shows emissions accompanying explosive volcanic eruptions that greatly exceed the volume of gas that could have been dissolved in the erupted magma (*77*). One plausible explanation for this excess gas is the presence of an exsolved volatile phase in the magma chamber. This explanation does not work, however, when excess gas is emitted over long time periods, and/or is not correlated with extrusive activity. For example, the 1995-2010 eruption of Soufrière Hills volcano, Montserrat, was accompanied by elevated gas emissions over the entire eruptive episode despite long hiatuses in eruptive activity. These observations require the volatile phase to be decoupled from the magma; they also appear to require segregation and accumulation of exsolved fluids at depth during the long period (centuries) of dormancy that preceded the onset of eruptive activity in 1995 (*78*).

Taken together, geophysical, geochemical and petrologic observations combine to present a picture of magmatic systems that extend through the crust, are largely crystalline, and are characterized by distributions of melt, crystals and exsolved volatiles that are heterogeneous in both space and time. This conceptual model of TCMS raises new questions about the development and stability of these complex systems and the processes that control their chemical and physical evolution.

TEMPORAL EVOLUTION OF MELTS AND MUSHES

Most crustal magmatism is driven, ultimately, by the influx of mantle-derived basaltic melt. Rates of magma generation are governed by plate tectonics and mantle convection, and modulated by melt segregation and transport within the mantle. These rates are assumed to be slow, and intrusion into the lower crust incremental. Volcanism is, in stark contrast, episodic, often involving large magma volumes erupted over very short times (*79*). This dichotomy suggests that a balance between slow magma accumulation and rapid magma release is inherent to magmatic systems, and that the physical processes that control chemical evolution of magma also control the episodicity of volcanism. Thus a key question relates to threshold behaviours that control the physical transitions from slow melt accumulation to rapid melt transfer. These transitions are most likely modulated by processes that control the distributions of melts, crystals, and volatiles within the crust.

Primitive basaltic melts are supplied incrementally to the crust from the mantle. Their initial fate is determined by thermal constraints, and by temperature-dependent physical processes. If individual magma additions are sufficiently small that they thermally equilibrate with their surroundings at the level of emplacement (*80*), the ambient temperature will impart one of three possible outcomes: (1) complete solidification, (2) formation of (non-eruptible) mush or (3) formation of (eruptible) magma.

In the lower crust, high ambient temperatures approach the solidus and are thermally favoured to keep primitive basalt additions above the solidus; for this reason, the flux required for magma accumulation is low (*8*). It is also an environment where dense basalts stall, cool and crystallize to form both melt-rich mushes and evolved melts. Here the mushy state is easily achieved and melt-bearing regions may be sufficiently long-lived to allow extensive melt segregation and fractionation (*23*). Segregation occurs by mush compaction, which squeezes out evolved melts with compositions that reflect the input melt composition (particularly the volatile content) and the temperature contrast between the melt and the ambient environment. The segregated melt evolves by cooling-driven crystallization, which is typically much faster than compaction. The derivative melts can then ascend to higher levels in the crust, leaving behind refractory cumulate rocks.

At shallower levels in the crust, temperatures can be maintained above the solidus between melt inputs only when intrusion rates are sufficiently high or intrusion volumes are sufficiently large (*23*). In these circumstances, melt-bearing mush can persist for long periods of time and chambers of eruptible magma can grow (*8*). The rate of cooling depends on both melt volume and geometry. Large melt intrusions can cool rapidly if they pond and convect, mix with cooler (stored) melts, assimilate partially molten wall rocks, or are associated with major hydrothermal systems in overlying crust. Small intrusions may be too thin to convect, but lose heat rapidly by conduction. Cooling rates are thus minimized when the temperature contrast between melt and host is small, or when the crystal content is sufficiently high to prevent convection (*81*). In the upper crust, individual magma batches moving from hotter to cooler regions achieve local thermal equilibrium in $<10^3$ years (*19*). Large igneous systems (spatial scales of 10 to 100 km), however, may have lifetimes of 10^5 to 10^7 years. Accommodating large volumes of eruptible magma in the middle and upper crust over these time spans requires a sustained and localised magma flux at rates that are substantially (one to two orders of magnitude) higher than average (*82, 83*).

Thermal requirements for maintaining *non-eruptible crystal mushes* in the upper crust are less severe. Here the temperature threshold is the solidus rather than the temperature at which the system is ~50% molten. This means that mushes can exist at temperatures a few hundred degrees lower than eruptible magmas, particularly when they have high water contents (*14, 83*). By definition, melt-rich crystal mushes will have a yield strength that inhibits initiation of convection (*84*), and a sufficiently high viscosity that convection, even if started (e.g., by heating and crystal resorption), will be sluggish (*81*). Thus cooling will occur predominantly by conduction, and on the basis of heat transfer alone we infer that upper crustal mush systems are characteristically long-lived. Whether the mush is maintained at melt contents close to that required for eruptibility (~40%; (*85, 86*)) or approaches the solidus (*87*), will depend on magma composition, input rates, local stress fields and heat transfer to the surroundings.

Thermal conditions, however, provide only one constraint on mush evolution. An additional consideration is the changing spatial distribution of melt and crystals, including melt segregation by compaction (*22, 88*) or shear (*89*). At low melt fractions, the time and length scales of compaction are strongly controlled by the melt viscosity and matrix properties (permeability, bulk and shear viscosity; (*22*)). Compaction time scales can vary from thousands of years (for basaltic melts) to hundreds of thousands of years (for wet silicic melts); corresponding compaction length scales vary from hundreds of meters (basalt) to only a few meters (wet rhyolites). Broadly this explains why many crustal ultramafic and mafic plutonic rocks are refractory adcumulates from which melt has been largely extracted (*90*); granitoid compositions, in contrast, suggest less efficient melt-crystal separation (*91*) notwithstanding long cooling times. Compaction-driven melt segregation and evolution can also explain both the overall density and compositional stratification of the crust (*9*) and, in a general sense, why long-lived mush systems are favoured at higher levels in the crust, where the bulk and melt compositions are more evolved and compaction time scales are long.

Melt segregation processes are well described in shallow mafic bodies. Relatively thin (<200 m) mafic lava flows, lava lakes and shallow sills, for example, experience rapid cooling; here the extent of compaction-driven melt segregation is modest, generally ceasing at crystallinities of 30-50% (50-70% melt; (*92-95*)). These melt retention values overlap with melt estimates for basal (rapidly cooled) sections of large mafic intrusions, such as the Skaergaard *((96-99);* Fig. 5a). Higher in these intrusions, textures are adcumulate and melt segregation is efficient, with trapped liquids representing only 1-5% of the resulting rocks (*90*). In these systems, melt segregation efficiency can be modeled as a balance between rates of crystal settling, melt convection through the mush, compaction and cooling, and changes in the crystallizing mode (which controls the density difference $\Delta\rho$ between melt and matrix; Fig. 5b). Convection through the mush is controlled by melt viscosity and matrix permeability, in turn a function of melt fraction and mean crystal size (Fig. 5c). The strong size dependence of permeability is interesting and demonstrates that the permeability reduction expected because of reduced porosity accompanying crystal growth is countered, at least in part, by increasing crystal size.

The efficiency of melt segregation in upper crustal silicic magma bodies is less well constrained, and lies at the center of a controversy concerning links between plutonic and volcanic rocks (*91, 100, 101*). Under favorable circumstances (hydrous silicic melts with matrix and melt viscosities of 10^{15} - 10^{17} and 10^{4} - 10^{5} Pa s, respectively; (*102*)), models suggest that compaction-driven melt segregation could occur over geologically reasonable time scales (10^{4}-10^{5} years). Efficient melt extraction in these systems, however, requires either maintenance of upper crustal mush zones at high temperatures (high magma influx rates) and/or sufficiently low melt viscosities (high dissolved volatile contents; (*83, 103*)).

The models of melt compaction and segregation reviewed above consider only the physical behaviour of the melt and matrix. Chemical interactions will also be important if the melt and crystals in the mush are not in equilibrium. Segregating melts that travel through porous networks in a temperature gradient may heat and melt matrix crystals, which, in turn, can focus melt into high permeability channels (*104, 105*). When upward moving melt is reactive chemically as well as thermally, progressive reactions also change the composition of the both the percolating melt and the matrix crystals (*106, 107*). Although reactive models have long been applied to grain boundary flow of melts in mantle (*105, 108*) and crustal metamorphism (*24, 109, 110*), there are very few studies of reactive flow within mush-dominated magmatic systems. Exceptions include descriptions of reactive melt infiltration in mafic sills (*111*) and intrusions (*112, 113*). Here melt is driven towards the cooler margins by crystallization-driven volume reduction and resulting pressure gradients. Lateral melt infiltration causes incompatible element concentrations to increase in the marginal facies (*111*). Similarly, observations of unusually high concentrations of incompatible elements in the lower oceanic crust have been used as evidence of reactive flow (*114-117*). Evidence for reactive flow may also be cryptic, as illustrated by the inferred reaction of melt and clinopyroxene to form amphibole in the source regions of arc magmas (Fig. 2b, (*118*)).

Pervasive reaction during flow requires melt to percolate along grain boundaries. Melt geometries in high crystallinity mush zones have not been extensively studied, except in cumulate nodules brought to the surface with volcanic rocks (*119*). These nodules commonly preserve melt along crystal boundaries and in the triple junctions between crystals, suggesting that processes related to reactive flow are likely to exert strong controls on the temporal, spatial, thermal and compositional evolution of mushy magmatic systems. Importantly, percolative, reactive flow through mush piles can cause geochemical variations that are not captured by arithmetical descriptions of fractional or equilibrium crystallisation (*120*). As a consequence, the tools commonly used to interpret geochemical variations in igneous rocks may be ill-suited to understanding the processes that created them.

Volatiles also migrate through active magmatic systems, as demonstrated by high volatile fluxes measured in arc volcanoes (*72*). Physically, dissolved volatiles contribute to melt buoyancy and enhance upward magma ascent. The chemistry of the exsolved volatile phase varies with the depth of exsolution (*76*). Volatiles with low solubilities will exsolve first; in the case of carbon dioxide this may occur in the lowermost crust or mantle if initial CO_2 contents are sufficiently high (*71*). Volatile exsolution further increases magma buoyancy, unless the low-density volatile phase segregates and decouples from the crystal-melt suspension (*78*). In three phase (melt-crystal-fluid) systems, the geometry of the exsolved volatile phase is controlled largely by the crystal fraction (*86, 121, 122*). In particular, as the crystal concentration approaches the packing limit (~ 50-65%; Fig. 1c), concentration of the volatile phase into narrow channels may greatly enhance segregation efficiency. Pervasive upward migration of exsolved volatiles, in contrast, may disrupt loosely bound crystal networks and has been postulated to trigger very large eruptions of crystal-rich intermediate-composition magmas (*123*). At high crystal contents (>~60%), the low-density volatile phase forces melt through the locked crystal network in a process known as filter pressing (*92, 122, 124*). At very high crystallinities (very low melt permeabilities), gas-generated overpressures can fracture and brecciate the solidifying mush (*125*). Thus the efficacy and style of volatile extraction from magma reservoirs may change significantly as they cool and crystallise.

5. MAGMATIC PROCESSES FROM A MUSH PERSPECTIVE

The review presented above shows that controlling processes in crystal-rich mushes are different from those in melt-dominated magma chambers. Mushes are multiphase systems in which buoyant melt (and sometimes fluids) are distributed within a deformable crystalline matrix; the dynamics of such systems are characterised by compaction-driven segregation of melts and fluids from the crystalline matrix (*22-24, 105*). Melt-rich segregations, in turn, are inherently unstable and can amalgamate to form larger magma accumulations. These magma chambers are susceptible to processes of convection, cooling and crystallization that are commonly considered to control magma evolution (e.g., (*81*)). Thus a mush perspective both extends the way in which we think about igneous and volcanic processes, and presents challenges for future work.

Key mush processes are illustrated in Figure 6. In the lower crust, the high mean temperatures and low melt viscosities combine to promote highly efficient compaction-driven melt segregation and, in thick cumulate piles, reactive flow. Evolved melts segregated from lower crustal intrusions will be enriched in volatiles and incompatible elements and will migrate upwards. Whether they reach the surface directly or stall and accumulate at intermediate levels will depend on the thermal state of the melt and intruded crust, as well as its thickness, rheology and state of stress. Ascending melts in arc environments commonly accumulate in the mid-crust (*20, 53, 126*), perhaps because of an abrupt increase in rigidity, and/or a decrease in density, at the brittle-ductile transition. At this level, open system processes in both the melt and mush will modulate the compositional evolution and stability of the resulting magma. Destabilization of melt lenses is probably common, and may be responsible for episodic and rapid ascent of evolved melts to upper crustal reservoirs. During this process, mass balance may be achieved by convective return of dense refractory cumulates to the lower crust through negative diapirism (*64, 127, 128*). In this way, the dual processes of (1) ascent of evolved melts and (2) sinking of refractory cumulates can explain not only the abrupt mid-crustal change in composition and density observed in crustal arc sections (*9*), but also growing evidence for mid-crustal magma intrusion and storage (*20, 30, 75, 129*).

Episodic 'recharge' of upper crustal magma chambers is widely recognized as necessary to grow large magma bodies, and is commonly invoked as a trigger for volcanic eruptions (*6*). In a trans-crustal model, this recharge magma derives from melt accumulations deeper in the system (Fig. 6), and commonly brings with it volatiles, crystals (antecrysts), crystal clots (glomerocrysts) and cumulate nodules derived from the (more crystalline) mush (Fig. 2). Although recharge magma may be preserved as enclaves with distinct compositional and textural features, the ubiquity of antecrysts and glomerocrysts in many erupted magmas attests to efficient mixing of successive recharge batches during the magma accumulation process. This mixing may be aided by volatile exsolution during both magma ascent (*71*) and enclave cooling (*130*). When ascending separately from the melt, the volatile phase may travel along grain boundaries to cause either reactive melting (if it increases P_{H2O}) or reactive crystallisation (if it decreases P_{H2O}; Fig. 3b), or may be released en masse to interact with higher levels of the magmatic system (Fig. 6); loss of those volatiles to the surface by passive degassing has been suggested as an eruption trigger (*131*). Alternatively, volatiles may segregate and become trapped as a separate layer (*78*), or may feed a hydrothermal brine cap in ways that are critical for ore formation ((*132, 133*); Fig. 6). It is possible that rapid release of magmatic volatiles may also trigger phreatic eruptions from hydrothermal systems.

Physical decoupling of the crystal, melt and volatile phases will also affect their chemical signatures. In this way, different processes are recorded by the non-eruptible mush, the mobile melt, and the magma (which may combine both mobile and non-mobile elements). Segregation of melt from crystal mush explains extensive geochemical evidence for fractionation as a major process in the evolution of igneous rocks (*39*), as well as general characteristics of arc crusts (*134*). At the same time, reactive flow associated with melt/volatile segregation means that the bulk compositions of magmas and gases are the integrated result of processes that have acted over space and time. As a result, trace elements affected by reactive flow may be difficult to reconcile with bulk crystallization models (*116, 135*). Bulk analyses of crystal separates from a single sample, or even analyses of individual crystals, may similarly integrate a wide range of spatio-temporal information, such that a single rock sample can preserve a rich record of the magmatic history. Interrogation of these records driven by rapid advances in microanalytical tools have provided much of the evidence for subvolcanic mush zones; future interrogation of the highly variable crystal cargo transported in many erupted magmas (Fig. 2,3) will help to assemble a comprehensive picture of the evolution of individual magmatic systems in space and time.

A mush perspective also provides new insight into mechanisms by which eruptible magma bodies form. A widely applied conceptual model of magma chamber formation involves melt inputs into an upper crustal magma chamber at a rate that is sufficiently fast to grow a single large magma body (Fig. 1a). This incremental model of melt accumulation generally assumes that aliquots of melt are emplaced into subsolidus country rock. The universal applicability of incrementally grown magma chambers is being challenged, however, by evidence that many igneous processes occur over timescales of decades to millennia, much faster than incremental growth models suggest. Are there other ways to grow magma chambers that can account for this paradox? One alternative mechanism is to rapidly redistribute melt from vertically stacked lenses within a mush into a single magma chamber (*64, 78, 136*). Rapid destabilisation explains not only the dichotomy of slow melt segregation followed by rapid magma assembly, but also addresses the space problem, which is that the country rock must be either displaced or assimilated to make room for the growing magma chamber. Destabilization of a vertically extensive crystal mush-melt system, however, involves rearrangement of crystal and melt layers; because the melt and crystal phases are incompressible, this can occur without substantial volume changes except when accompanied by upward transfer of volatile-saturated melts. At the same time, mush destabilization explains the occurrence of crystals with very different histories in a single sample of erupted rock (Fig. 6).

Finally, a mush model has important implications for understanding volcanic eruptions. First, it offers a way to merge, conceptually, slow (plate rate) processes of melt generation with rapid eruption of large magma volumes. This slow-fast dichotomy is nicely described by Hildreth and Lanphere (*137*) in their description of dormancy as "an anthropocentric notion and generally only an upper-crustal condition; suspensions of eruptive activity… need signify no fundamental change in deep-level magmatic processes." It also offers a cautionary note for efforts to anticipate future large eruptions. On the one hand, if distributed melt lenses can destabilize rapidly to accumulate in the shallow crust, or are tapped syn-eruptively to produce large cumulative erupted volumes (*138*), then it may be difficult to recognize systems poised to generate a large eruption. On the other hand, the association of very large eruptions with crystal-rich (≥35% crystals) magma suggests that these eruptions are triggered when a near-eruptible mush (~60% crystals) is mobilized by inputs of either volatiles (*123*) or melt (*139*). This model raises an important question that relates to the standard state of the mush: near-solidus or near-eruptible? Thermal and rheological constraints suggest that compaction and melt segregation in the lower crust should efficiently segregate melt and crystals; as a result, we expect that the lower crust comprises near-solidus mush with interspersed melt lenses. These same constraints suggest that it should be easier to maintain near-eruptible mush in the upper crust, where melts are more evolved and the system is cooler. Whether maintaining large volumes of near-eruptible magma also requires unusually high magma inputs is an active topic of debate (*83, 103, 140*). Answers to these questions will require integration of geophysical, geochemical and petrological studies of active volcanic systems.

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Figure Captions

Figure 1. Magmatic systems. (a) Upper crustal magmatic system where silicic melts segregate from underplated mafic magma that is intermittently resupplied from deeper levels. Redrawn from (*141*). (b) Trans-crustal magmatic system, where melt processing in the deep crust produces melts that are transferred to mid- and finally upper crustal levels. The potential for transient vertical connectivity in this system presents the possibility of successive destabilization of melt lenses (*78*). (c) Changes in magma (orange) and mush (gray) rheology as a function of particle volume fraction. Blue curve is calculated assuming a maximum packing fraction of 0.6 and a classical Roscoe-Einstein formulation (*10*). Red curve uses the formulation of (*12*). Inset shows changes in mush strength as function of particle volume fraction; green dashed curve is experimental data using Westerly granite, purple dashed curve is experimental data using Delegate aplite. Redrafted from (*11*).

Figure 2. (a) Photomicrograph of dacite lava from Dominica, Lesser Antilles, showing wide diversity of crystal phases, size and zoning patterns. Crystals are present as single crystals and crystal clusters (glomerocrysts) and have a wide range of size and shape caused by entrainment of mush and mixing of magmas with different crystllisation histories. This textural complexity creates challenges for conventional petrographic characterization. Photo courtesy of R. Arculus. (b)

Figure 3. Complex crystal histories in magma from Mount St. Helens volcano, USA. (a) Backscatter SEM photo (inset) and areal distribution of plagioclase composition (as X_{An}) in a single crystal from the December 1980 eruption of Mount St. Helens. Note that the core composition (mode ~An_{45}) is significantly more evolved than that

of the broad inner rim zone (mode ~An_{65}). (b) Experimentally determined phase diagram (temperature vs. water pressure P_{H2O}) for plagioclase in a dacitic magma composition appropriate for Mount S. Helens. Fields of solid (S) and melt (Melt) are labelled. The field of melt + crystals is contoured for anorthite content and melt fraction (F). Highlighted in green and blue are the compositions of the plagioclase core and rim, respectively. Both (a) and (b) are redrafted from (*41*). (c) Range of plagioclase ages measured in plagioclase mineral separates from Mount St. Helens eruptions over the past 2000 years. Note that crystal ages span time scales of thousands of years prior to eruption (data from (*68*)). (d) Age range of individual zircon crystals from Mount St. Helens eruptions over the past 3900 years. Here pre-eruptive crystal ages exceed 200,000 years, often within single crystals (redrafted from (*69*)).

Figure 4. Pre-eruptive magma storage depths. (a) The maximum pre-eruptive earthquake depth and (b-g) the volatile content of crystal-hosted melt inclusions from individual volcanic eruptions. Earthquake data are from (*75*). Melt inclusion data are from Soufriere Hills Volcano, Montserrat (SHV: (*53*)); Galapagos ridge (GAL; (*142*)); Mt. Etna, Italy (ETNA; (*143*)); Popocatepetl, Mexico (POPO; (*144, 145*)); Mount St. Helens, USA (MSH; (*71*); May 18 data only); and the Oruanui eruption (OR), Taupo volcano, New Zealand (*146*). Arrows denote melt inclusion entrapment pressures that exceed 400 MPa. The low-pressure mode in the MSH data records the shallow magma intrusion that preceded the 1980 eruption. SHV data also show depth range estimated from pyroxene compositions.

Figure 5. Compaction and melt segregation in mafic intrusions. (a) Changes in trapped liquid fraction through the Skaegaard intrusion, as calculated from U contents. An initial trapped melt content of ~40% is inferred from the crystallinity required to form a touching framework. (b) Variations in compaction rate with thickness of the compacting layer, calculated for conditions appropriate for the lower (purple) and upper (blue) layers of the Skaergaard intrusion, as shown in (a). Both (a) and (b) redrafted from (*99*). (c) Variations in permeability with changing melt fraction for different crystal sizes; dashed and solid lines show different proposed versions of the Kozeny-Carman equation used to calculate the curve. Solid lines use the lower equation, where κ is permeability, d is average grain size, φ is the melt fraction and C_1 is a constant equal to 180 (*147*). Dashed lines use the upper equation with C = 10.

Figure 6. Schematic diagram showing the types of processes likely to operate within a TCMS. In the lower crust, high ambient temperatures and low viscosity melts promote efficient compaction and segregation of dense mafic cumulates from more evolved melts. Destabilization of ponded magma in the mid-crust is the likely source of 'recharge' magma that supplies upper crustal magma chambers; the process of destabilization may create and entrain cumulates, glomerocrysts and antecrysts, while magma decompression will cause volatile exsolution. Volatiles introduced to the shallow magma chamber may accumulate at the chamber roof, segregate to form an isolated volatile layer, or feed into an overlying hydrothermal system.

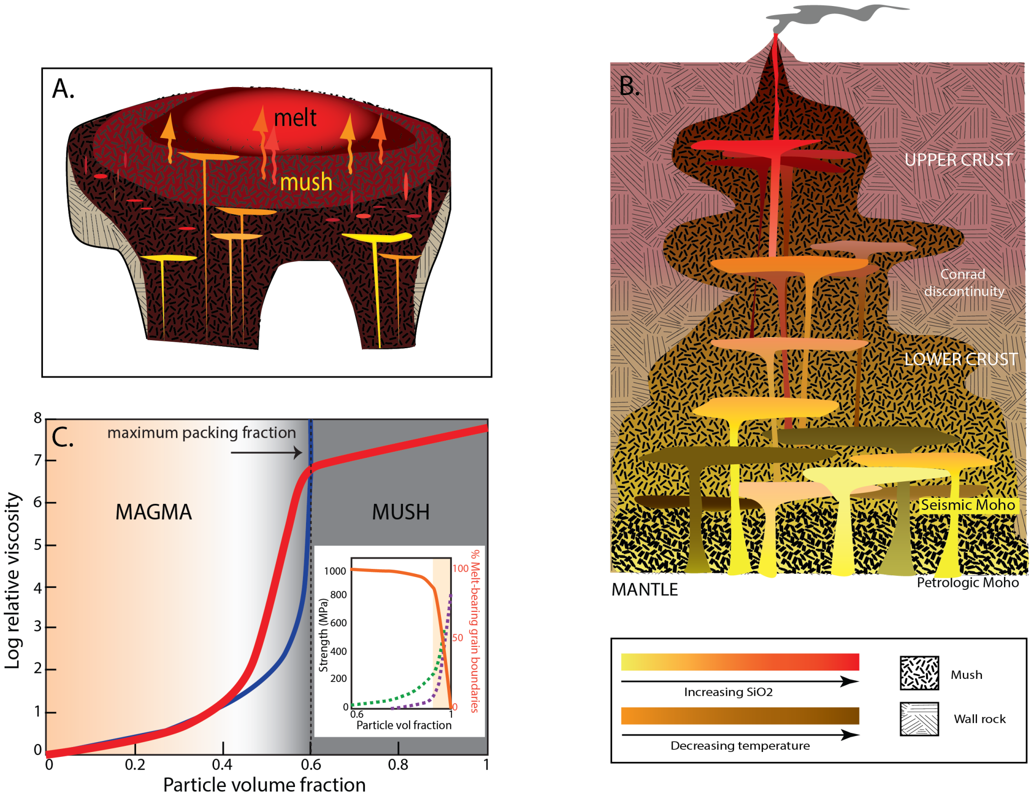

A.
melt
mush
B.
UPPER CRUST
Conrad discontinuity
LOWER CRUST
Seismic Moho
MANTLE
Petrologic Moho
Increasing SiO2
Decreasing temperature
Mush
Wall rock
C.
maximum packing fraction
MAGMA
MUSH
Log relative viscosity
Particle volume fraction
Strength (MPa)
% Melt-bearing grain boundaries
Particle vol fraction

FIGURE 1

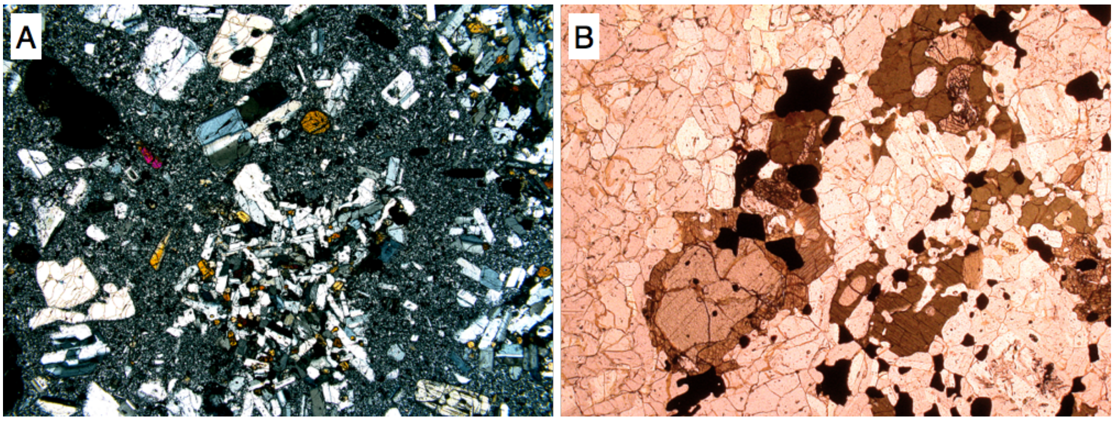
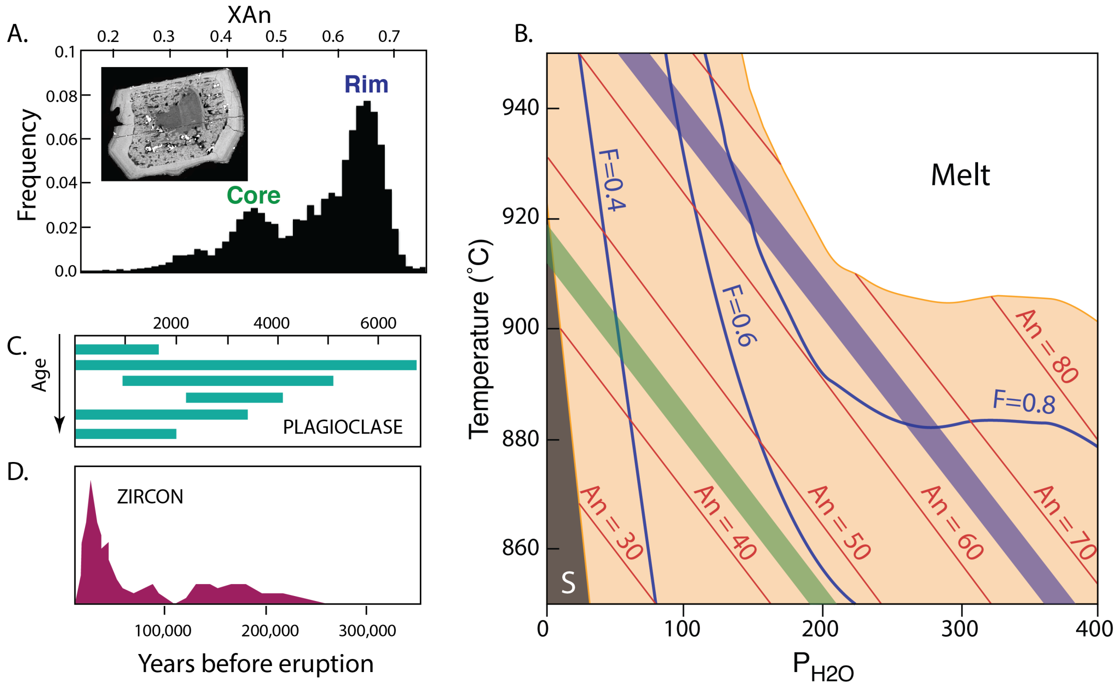
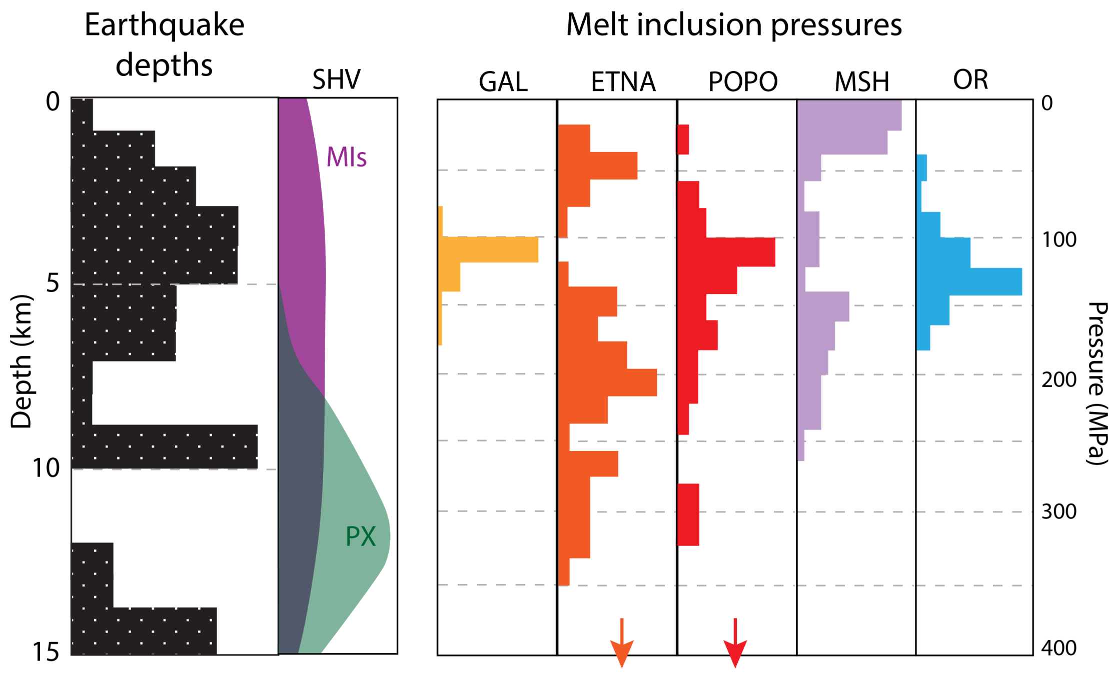


FIGURE 2



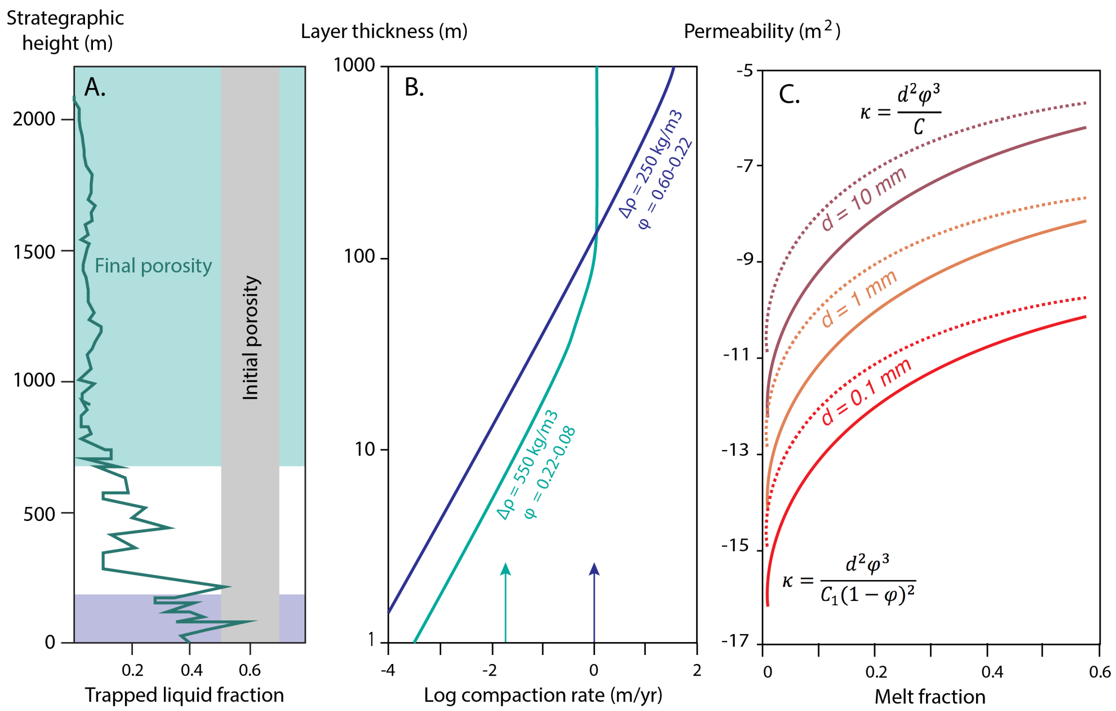
A.
XAn
0.2
0.3
0.4
0.5
0.6
0.7
Frequency
0.1
0.08
0.06
0.04
0.02
0.0
Rim
Core
C.
Age
2000
4000
6000
PLAGIOCLASE
D.
ZIRCON
100,000
200,000
300,000
Years before eruption
B.
Temperature (˚C)
940
920
900
880
860
Melt
F=0.4
F=0.6
F=0.8
An = 30
An = 40
An = 50
An = 60
An = 70
An = 80
S
0
100
200
300
400
P H2O

FIGURE 3



Earthquake depths
SHV
Melt inclusion pressures
GAL
ETNA
POPO
MSH
OR
0
5
10
15
Depth (km)
MIs
PX
0
100
200
300
400
Pressure (MPa)

FIGURE 4



Strategraphic height (m)
A.
2000
1500
1000
500
0
Final porosity
Initial porosity
0.2
0.4
0.6
Trapped liquid fraction
Layer thickness (m)
B.
1000
100
10
1
Δρ = 250 kg/m3
φ = 0.60-0.22
Δρ = 550 kg/m3
φ = 0.22-0.08
-4
-2
0
2
Log compaction rate (m/yr)
Permeability (m2)
C.
-5
-7
-9
-11
-13
-15
-17
$\kappa = \frac{d^2\varphi^3}{C}$
d = 10 mm
d = 1 mm
d = 0.1 mm
$\kappa = \frac{d^2\varphi^3}{C_1(1-\varphi)^2}$
0
0.2
0.4
0.6
Melt fraction

FIGURE 5

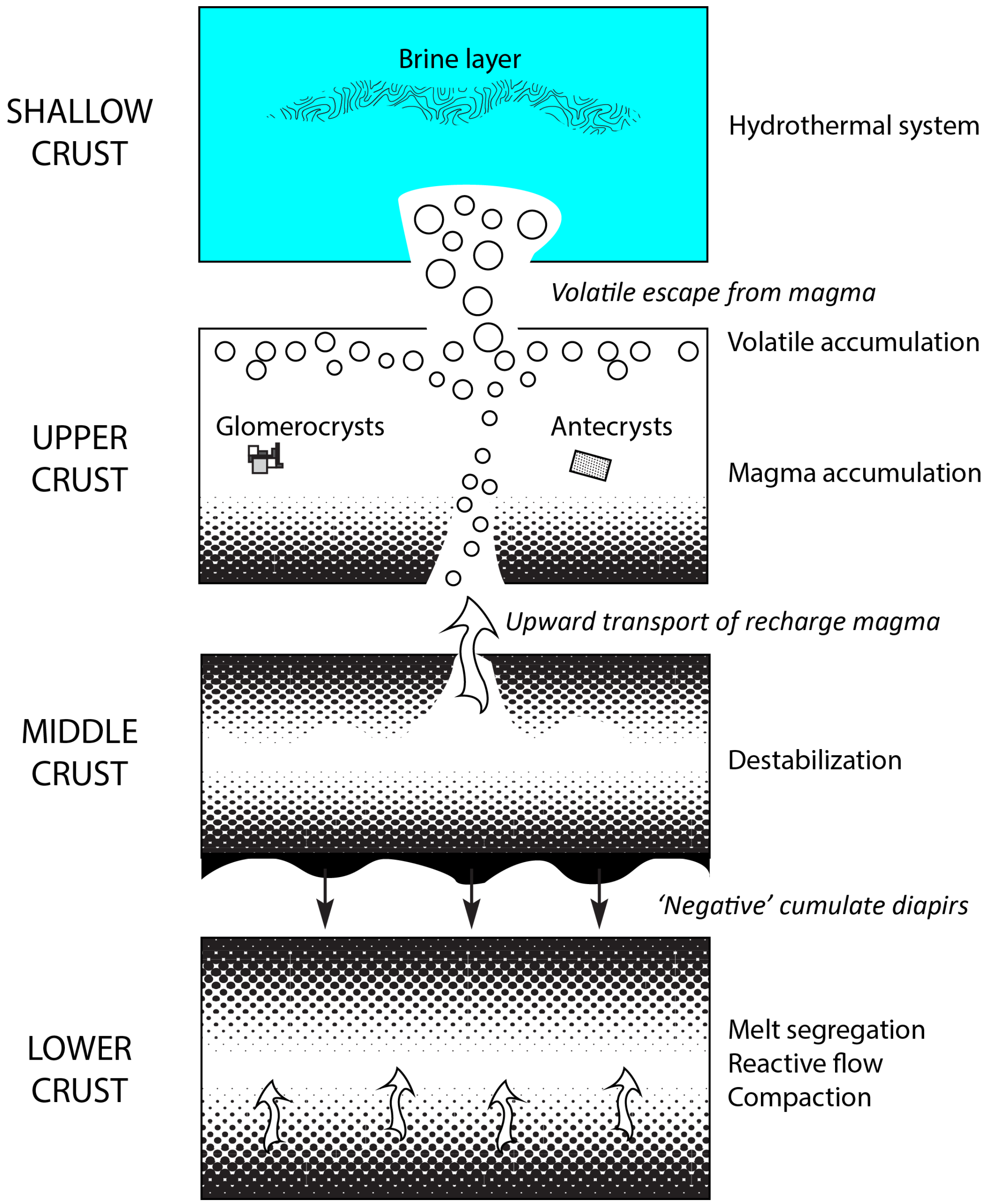

SHALLOW CRUST
UPPER CRUST
MIDDLE CRUST
LOWER CRUST
Brine layer
Hydrothermal system
Volatile escape from magma
Volatile accumulation
Glomerocrysts
Antecrysts
Magma accumulation
Upward transport of recharge magma
Destabilization
'Negative' cumulate diapirs
Melt segregation
Reactive flow
Compaction

FIGURE 6