PHILOSOPHICAL TRANSACTIONS A

Phil. Trans. R. Soc. A.
doi:10.1098/not yet assigned

Architecture and dynamics of magma reservoirs

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Keywords: Magma reservoir, mixing, mush, volatiles, melt, crystals

Summary

This introductory article provides a synopsis of our current understanding of the form and dynamics of magma reservoirs in the crust. This knowledge is based on a range of experimental, observational and theoretical approaches, some of which are multidisclipinary and pioneering. We introduce and provide contextual background for the papers in this issue, which cover a wide range of topics, encompassing magma storage, transport, behaviour and rheology, as well as the timescales on which magma reservoirs operate. We summarise the key findings that emerged from the meeting and the challenges that remain. The study of magma reservoirs has wide implications not only for understanding geothermal and magmatic systems, but also for natural oil and gas reservoirs and for ore deposit formation.

Aims and scope of the issue

This special issue arose from a Theo Murphy Discussion meeting held in November 2017. The meeting brought together petrologists, geologists, geodesists, geophysicists, fluid dynamicists and modelers to discuss how, where and in what form magmas are stored in the crust; and the implications for geohazards, for understanding fundamental Earth processes such as crustal growth, and for ore deposit formation. The meeting was sparked by the large number of new discoveries in the field, across a range of disciplines, and by the emergence of a number of critical localities that have been the focus of key breakthroughs. In this issue we present chapters led by the speakers in that meeting, aimed at capturing the state-of-the-art in our understanding of the form and storage of crustal magma bodies, while exploring critical unknowns and hence identifying fruitful targets for future interdisciplinary work. Understanding storage of magma in the crust has implications not only for reconstructing crustal growth through emplacement of plutons, and understanding the precursory signals to volcanic eruptions, but also for understanding the rheological behaviour of complex multiphase systems in crustal reservoirs, knowledge that might be laterally applied to other systems e.g. oil

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and gas reservoirs and ore deposits.

What is a magma reservoir?

A *magma reservoir* may be defined as a region of partially or wholly molten rock with varying proportions of melt, crustals and exsolved volatiles (1). In this issue, Sparks et al. define a magma reservoir as '*the domains within the magmatic system that contain melt (+/- exsolved fluid) and by definition are above the solidus*' (2). Some authors have suggested that parts of the magma rservoir may be entirely below the solidus temperature, in 'cold storage', and are later rejuvenated (3). The parts of the reservoir which contain sufficiently high crystal contents to form a semi-rigid framework may be termed 'mush' (4). Likewise, pockets or bodies of melt with a very low (or zero) crystal fraction may exist in magma reservoirs, formed through melt expulsion from a mush, through compaction (5, 6), convective or diffusive exhange or gas-driven filter pressing (7, 8), or as super-liquidus melts injected into more crystal-rich material (a mush or solid country rock), which then cool and crystallise. In some systems, the '*reservoir*' may span the vertical extent of the crust (and perhaps beyond, into the mantle); and may be termed a '*trans-crustal mush system*' (2, 9, 10). In other magmatic systems e.g. Iceland; (11-14) there may be multiple, 'stacked' storage areas for magma (15), separated by sub-solidus material, which has been well constrained both petrological and geophysically (13, 16).

In this volume, Sparks et al. review the history of the 'magma chamber' concept, whereby a liquid vat of magma slowly cools and crystallises to produce layers of igneous minerals reflecting the liquid line of descent, to a greater or lesser degree affected by mixing and/or assimilation processes. As our observations of magmatic systems have improved, the community has acquired abundant evidence to suggest that large volumes of melt are not currently present in the crust and instead, melt may be disseminated in mush regions, and extracted shortly before eruptions (2, 9, 17). Much of this paper, and the papers which follow, lay out this evidence.

The different types of regions in a magma reservoir (mush, melt, rock) have vastly different rheological and physical properties, including their viscosity (18), their response to applied stress (19-21), and their conductivity (22, 23). Sparks et al. (this issue) delineate rheological domains for the magmatic system components and show that their effective viscosities may span 25 orders of magnitude, and consequently the timescales on which reservoir processes operate range from seconds to millenia (2). The inherent instabilities generated from the heterogeneity of the magma reservoir may give rise to scenarios whereby the reservoir may re-organise (17) or overturn (24, 25) on short timescales prior to, or during, eruption. This reorganisation may take place in response to a tipping point caused by the long term processes of fractionation, settling, compaction, reactive flow and second boiling (26); or it may be triggered by recharge of the reservoir by mafic magmas from the lower parts of the crust, or from the mantle (27-30). In some instances magma reservoir perturbations, and sometimes eruptions, may occur in response to tectonic forcing (31).

What is the ultimate fate of the melt in magma reservoirs? Magma reservoirs may feed eruptions; or they may freeze and become plutons. Between 80 and 90% of magma supplied to the crust is emplaced endogenously (32, 33), making pluton formation the most likely end point for much of the magma reservoir mass (34, 35). Plutonic rocks exposed at the surface provide a wealth of information about the emplacement mechanisms and differentiation processes of magma reservoirs in the crust (36, 37).

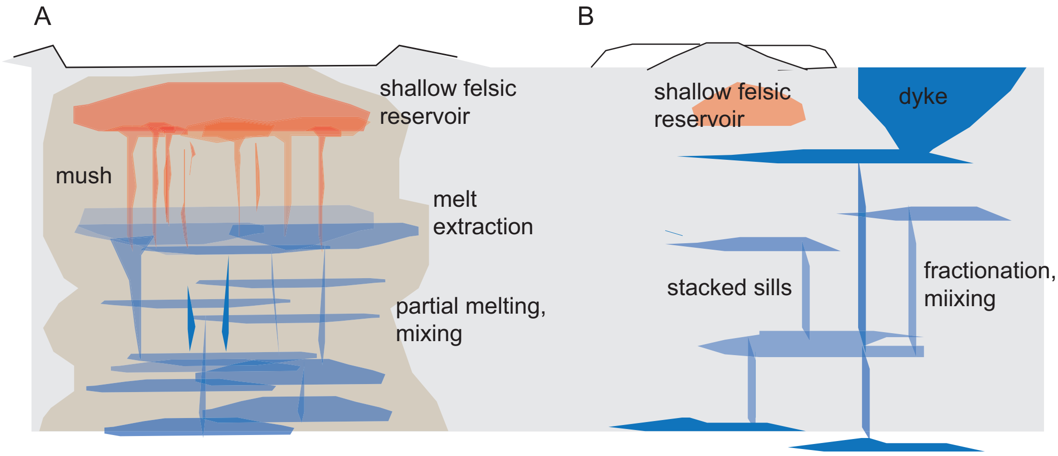

Figure 1: Schematic cartoon of the magma reservoir end member types that may exist in A: continental and arc settings (38); and B: in Iceland (discussed in text) (39).

Evidence for the vertical extent and form of magma systems

To understand magma storage in the crust, we might wish to know the depth at which magma is stored, and for how long. What fundamental processes are occurring inside a magma reservoir as it cools and is recharged? Can we find evidence of these processes both in the erupted rock and in the geophysical signals we can measure at the Earth's surface and buried in boreholes?

In the brittle crust, magma may be emplaced along structural features such as large-scale faults (40, 41), forming aligned, large composite batholiths, e.g. Rum, Arran, Adamello (42, 43). Examples of volcanic systems which are influenced by the tectonic stress field are many: e.g. lava domes in the southern parts of Monserrat are aligned parallel to the bounding faults of the half graben extending the southern part of the island (44); lava domes outside the Long Valley caldera emplaced parallel to regional extensional faults; and volcanoes aligned along transcurrent faults in the NW Bicol volcanic arc, Philippines (45, 46). Extreme cases of tectonic control on magma emplacement are provided by rift volcanoes, where magmas ascend through brittle faults in the continental crust associated with rifting, forming scoria cones along the rift margin and central volcanoes along the rift centre (47). At many volcanoes, CO_2 and He emissions are concentrated along faults around volcanoes, such as caldera faults but are also emitted from cross-cutting large-scale faults related to regional tectonics (48). In this issue, Biggs and Annen explore the tectonic controls on magma emplacement (49).

Seismology is an important tool to study the storage and transport of magma, in real time, reviewed by White *et al*. in this issue. Microseismicity occurs when rock fractures in a brittle fashion in cool crust. Not all magma movement is captured by earthquakes, however: once a magma-filled channel is 'opened' it may permit aseismic flow; likewise hot crust is ductile and does not fracture under commonly-accessible strain rates (11). In Iceland, crustal thickness ranges from 20 to 40 km (50) and the brittle part of the crust extends down to 6-8 km (11). Seismic tomography studies of central volcanoes in Iceland have discovered magma storage areas at depths of 3-6 km and deeper. For example, at Askja volcano, a series of high V_P/V_S ratio bodies indicate the presence of melt situated at discrete locations throughout the crust to depths of over 20 km (16), consistent with the model of stacked sills *(15)*. A dominant anomaly at 6 km depth cannot be the signature of crystal-poor melt, however, because it transmits shear waves, raising the possibility that this body may be mush-dominated (11). Deeper in the crust, at up to 25 km, earthquakes have persistently occurred in the same regions for over a decade, indicating persistent magma supply into sills, where the earthquakes occur at the sill tips, where the stress is highest (16).

Deeper in the crust, microseismicity can pick out the path of magma movement. Several years prior to the Icelandic Bardabunga eruption in 2014, for example, earthquakes occurred at depths of 7 and 22 km on the SE flank of the caldera (not directly beneath it) (51). Although the storage region beneath the caldera is thought to be at 5-12 km bsl (52), earthquakes do not occur deeper than ~8 km (53). These observations show that, at least sometimes, magmas may bypass the shallow magma reservoir as has also been suggested for many other volcanic systems, based on petrology e.g. Kīlauea and Mauna Loa, Hawaii (54, 55) and on ground displacements (56, 57).

Seismicity provides an enormous amount of information, from the distribution, size and timing of the earthquakes and from the focal mechanisms, and has led to new insights about how magma is stored in sills and migrates upward in deep dykes beneath Iceland (11). For example, a non-eruptive seismic swarm beneath the volcano of Upptyppingar occurred on a fault plane parallel to the dip and strike of the dyke (as opposed to 30 degrees to the dyke tip, which was expected). This observation, and the presence of both reverse and normal fault plane solutions, has produced a model for this swarm of repetitive melt freezing and fracture in the brittle part of the crust (58, 59). Furthermore, bursts of seismicity above the dyke may record the release of CO_2 from the ascending magma (60). Volatile release associated with magma transfer may be a common process ** (11), and has also been observed at Mammoth Mountain, inside Long Valley Caldera, California (61).

Seismic tomography studies have imaged bodies with slow p wave velocities and low v_p/v_s ratios (which may represent melt-bearing regions of the crust). A region containing 10-12% melt has been imaged at depths of 4 and 6 km beneath Mount St Helens, inferred to be the top of a reservoir that extends down to 14 km depth (62). At Nevado del Ruiz volcano (Colombia), a recent study has imaged high v_p/v_s at depths of 2–5 km, which has been proposed to indicate degassing and the presence of an exsolved volatile phase (see section 5) (63). Average melt percentages of 5% to 15% have been suggested for the upper crustal magma reservoir that underlies the Yellowstone caldera (64, 65).

Importantly, tomography yields an averaged view of the seismic velocity structure and relies on broad seismic wavelengths of hundreds of meters to hundreds of kilometers, thereby limiting the spatial resolution. Even with dense seismometer networks, it is a challenge to image upper crustal anomalies <~5 km^3 or features much < 10 km across (38). Erupted magmas typically contain >50% melt, which raises the question: can eruptible magmas hide within a crystal mush that is 85%–95% solid (38)? A region containing 5-15% melt when viewed at the large scale could actually contain isolated lenses rich in melt instead of the melt being broadly disseminated in small pores throughout. Are isolated bodies of melt are separated by mush, or by subsolidus rock? Seismic tomography is rarely capable of resolving the difference between larger (but still less than 100 m) sill-like bodies set in a rock matrix, from a bulk mush containing a small fraction of melt in its pore space (66, 67).

Magnetotelluric surveys provide complementary information; they measure directly the conductivity of the crust (and perhaps upper mantle, depending on the geometry of the survey). This method is therefore highly sensitive to the presence of melt (67-69), due to the effect of temperature on conductivity (23), as well as the presence of low density fluids, including brines (70). Recent studies have identified signals that suggest elevated melt fractions in regions extending laterally 10s-100s km in the Cascades (71), Afar (72), Uturuncu Volcano (68), and Vesuvius-Campi Flegrei (73), perhaps suggesting that magma reservoirs may be linked up under volcanic regions, and significant magma storage may take place in the mid-crust, the lower crust or straddling the Moho.

Magma intrusion and eruption (i.e. magma addition to, and withdrawal from, the crust) is associated with inflation and deflation of the Earth's surface (74, 75), reviewed by both Segall *et al*. and Biggs and Annen, in this issue. Pressure changes in magma reservoirs may be inferred from displacements of the ground surface, which may be measured in a range of ways. The rate of decay of deformation with distance reflects the centroid depth of the magma source region (74-76). Furthermore, the amplitude of the ground displacement signal is related linearly to the pressure change, or volume change of the reservoir (74, 77). Independent constraints are required to quantify one or both of these parameters. The relative magnitude of vertical versus horizontal displacements gives information about source shape (76). A wealth of new data, particularly from space-based measurement systems such as InSAR (77) have allowed the construction of an unprecedented, detailed record of ground deformation around volcanoes (49, 56, 78-81). A number of important generalizations have been made from the overview of deformation data. The Mogi model (82) has been the mainstay of deformation modeling for several decades, and does a surprisingly good job at reproducing the deformation observed at many volcanoes. The assumption implicit in this model is that the crust is elastic. For many eruptions, however, patterns in surface displacements do not fit a simple Mogi model and multiple, complex sources (involving sills, dykes and multiple point sources) are required to fully reconcile all of the observations (75, 80, 83-85). Segall, this issue, explores whether models involving a visco-elastic crust may be more suitable in some cases for modeling volcanic deformation (see later). It may even be possible to begin to reconcile models of complex magma reservoirs comprising regions of mush hosting multiple melt lenses, with surface observations (77).

Petrological studies of erupted magmas use barometry to place constraints on the depth of magma storage. For mafic magmas, 'OPAM' barometry uses the melt composition that is in equilibrium with plagioclase, clinopyroxene and olivine to estimate a pressure of 'last equilibration' (86, 87). Yet another method uses the density of fluid inclusions trapped inside crystals, which is proportional to pressure (88). **Figure 2** contains a summary of data from clinopyroxene-melt barometry on basalts from a range of volcanoes (89-95), and shows that magmas may be stored (and last equilibrate with melt prior to eruption) at great depth in the crust, close to and beyond the seismic Moho in some cases.

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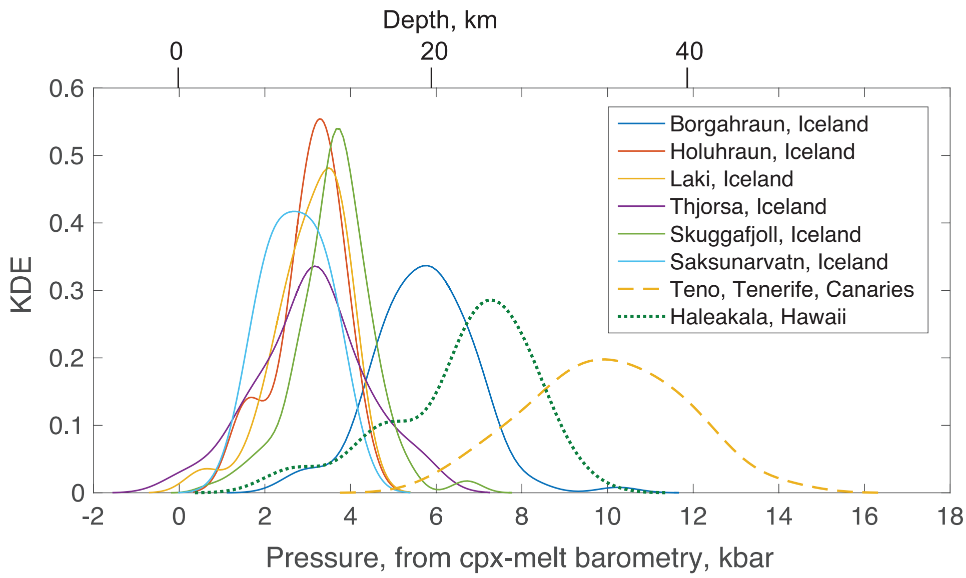

Figure 2: Last equilibration pressures of magma from clinopyroxene-melt barometry (95, 96) for a range of ocean island volcanoes (89-95).

Barometry of numerous Icelandic eruption products has shown that melts are stored over a large range of depths in the Icelandic crust (66, 95), reviewed by Maclennan in this issue. Petrological barometry estimates produce some agreement with the seismic constraints on magma storage (11). For example, clinopyroxene-melt pressures for Askja basalts typically average 2-3 kbar (6-9 km bsl) (92), a depth range that agrees well with the location of the main magma storage area deduced from seismicity (11, 16). Overall the data for Iceland are consistent with models of stacked sills throughout the crust and spanning the Moho (13, 15, 97, 98). During eruption, magmas may be tapped from either shallow storage areas, in the case of the central volcanoes, or from deeper for off-rift eruptions. It is important to note, however, that the petrological features of erupted rocks are a palimpsest of many processes acting over timescales of 10^0-10^4 years, perhaps entraining crystals from multiple reservoirs with different histories (66), as well as being on a much smaller spatial scale (microns to cm) than geophysical observations (which are over metres to kilometres): therefore it is extremely challenging to link these two data sources.

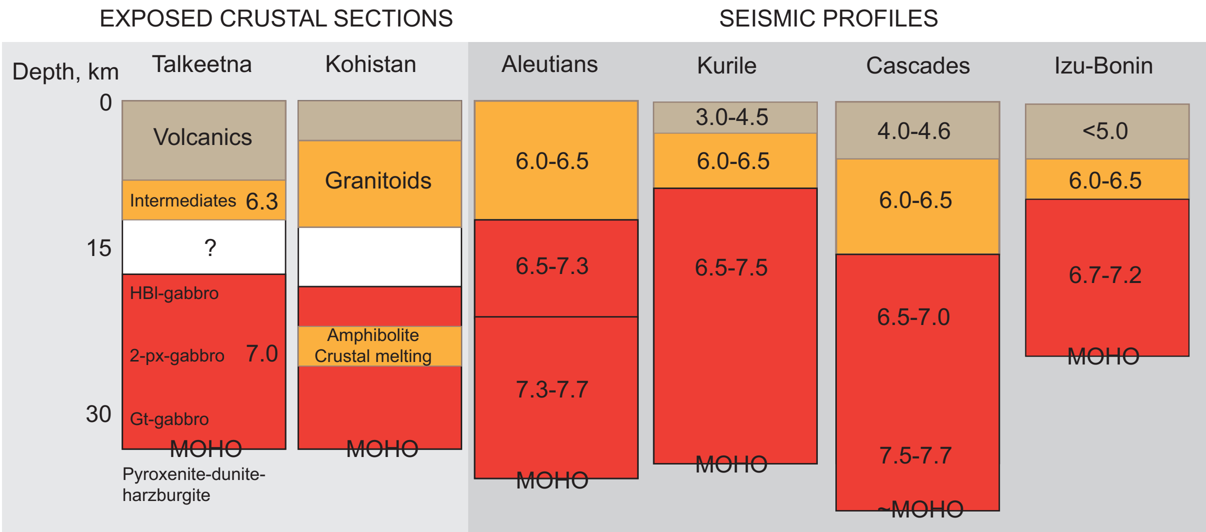

Figure 3: Schematic sketches of exposed arc sections of Talkeetna (Alaska, USA) and Kohistan (Pakistan) (100). Brown represents volcanic rocks, orange are intermediate to felsic plutonic rocks (53–72 wt% SiO_2), red is mafic lithologies (gabbro cumulates) (100). The Moho is defined as the transition between ultramafic rock (plagioclase absent) and gabbroic rock (plagioclase present). The righthand four sections are a summary of seismic velocities for a range of arcs (100): Aleutian (106), Kurile (107), Cascade (108) and Izu-Bonin (109).

Exposed crustal arc sections allow direct insight into the structure of the magmatic system preserved beneath volcanic arcs and clues about depths of magma storage and differentation (**Figure 3**). The most complete of these sections are Talkeetna (Alaska) and Kohistan (Pakistan). Talkeetna is an example of island arc crust, active between 200 and 160 Ma (99). The lower part of this crust is made of pyroxenites and garnet-bearing gabbros; hornblende gabbro-norites in the mid-crust, gabbroic rocks and intermediate-felsic plutonic rocks in the middle-upper crust, and volcanic rocks in the uppermost crust (~ 7 km thick), intruded by felsic to intermediate tonalites and quartz diorites (100). These upper crustal plutonic rocks crystallized at 0.13–0.27 GPa (5–9 km) (101). Kohistan is a section of island arc crust obducted during India-Eurasia collision (102). The lowermost part of the crust is ultramafic and consists of garnet-gabbros (granulites) that overlie and intrude the residual/cumulate rocks of the subarc mantle (103). The middle crust of the arc is separated into a northern section (middle crust represented by calcalkaline plutons of the Kohistan batholith) and a southern section (middle crust represented by metadiorites, metagabbros, metasediments, and metavolcanics of the Southern Amphibolite Belt) (100, 104). The shallowest part of the arc section consists of volcanic, sedimentary, and shallow-level granitic rocks in the northern part of the exposure (100). The Famatinian magmatic arc in western Argentina is a differentially exhumed section of a Late Cambrian-Middle Ordovician arc formed by plate subduction beneath the Gondwanan margin (105). Here, the source zones of tonalites and diorites are locted at the boundary of a lower, gabbro-dominated portion of crust and an upper intermediate unit. The intermediate rocks are sourced from mafic intrusions, melting of metasedimentary rocks, forming leucotonalitic vein and dyke system that coalesces to form leucotonalitic or tonalitic magma bodies. The preserved textures are migmatitic and may reflect melts caught in the act of segregation (105). Alternatively, these rocks may represent 'failed' source regions, from which the melt did not escape efficiently.

Figure 3 shows a summary of the main features of such sections, as well as some seismic velocity data from modern arcs. In all cases, high density mafic and ultramafic lithologies are present in the lower crust, which may represent cumulates; with intermediate and felsic plutonic and volcanic rocks in the mid and upper sections. In both cases, there is evidence for fractionation of melts beneath the Moho (defined as plagioclase-in), manifest as dunite and pyroxenite cumulates; the latter is the dominant lithology near the Moho (100). Above the Moho, gabbroic (olivine-free) cumulates dominate, which formed at 0.5-1.0 GPa (Hacker et al., 2008). Geochemically, the upper intermediate and felsic magmas can be related to the lower ultramafic portions by fractional crystallisation, suggesting that felsic liquids segregated from mafic cumulate mushes and ascended to shallower levels, creating stable density stratification. In the exposed sections, there is abundant evidence for crustal melting, with migmatites and tonalite melts produced from the partial melting of metasediments (105). In all of the exposed sections, the mid-crust is highly heterogeneous, with host amphibolite and multiple types of magmas present in close proximity and is likely a region of extensive mixing and mingling, and assimilation.

In general, the features of the exposed arcs are consistent with the results of seismic studies of modern arcs (110) (**figure 3**), which show distinct seismic velocities for upper, mid and lower crust. The most variability, in both velocity and thickness, occurs in the mid crust, where gabbroic as well as felsic rocks have been inferred. Seismic velocities of the lower crust in modern arcs are consistent with granulite facies rocks (111), but are thought to be highly dependent on the water content of melts in the lower crust (112). Models to describe how these arc sections evolve with time typically involve a young arc crust dominated by fractionation and mixing/mingling, followed by a second stage, once the arc crust is much thicker, which is characterised by extensive crustal assimilation (110).

Magmatic processes in reservoirs

Liquid vats versus crystal mush: evidence from microstructures

There is abundant evidence that liquid-rich magma bodies have existed in the past and may exist locally on small scales today. Such a melt-rich body may cool through inward solidification, generating a progressive layering (from mafic to more evolved bulk compositions) by efficicent separation of crystals from the remaining liquid (4, 113-115), perhaps interrupted by recharge events which reset the bulk composition and temperature, or 'sedimentary' features such as stoped and settled blocks and gravity currents involving crystal-rich suspensions (116, 117). Fractionation by gravitational settling of crystals can generate a sequence of modal types and crystal sizes on the floor of the 'chamber' (4, 113, 118). Crystals may orient in the settled layer from either compaction or alignment due to flow or shear at the crystal-melt interface (119-122). Examples of such features exposed at the Earth's surface include the Skaergaard intrusion (117), and the Rustenburg Layered Suite of the Bushveld Intrusion (115). For small melt bodies, such as sills, it is even possible to infer the presence or absence of, and the strength of, convection based on the stratigraphic distribution of grain size of the crystals at the upper and lower surfaces of the intrusion (123).

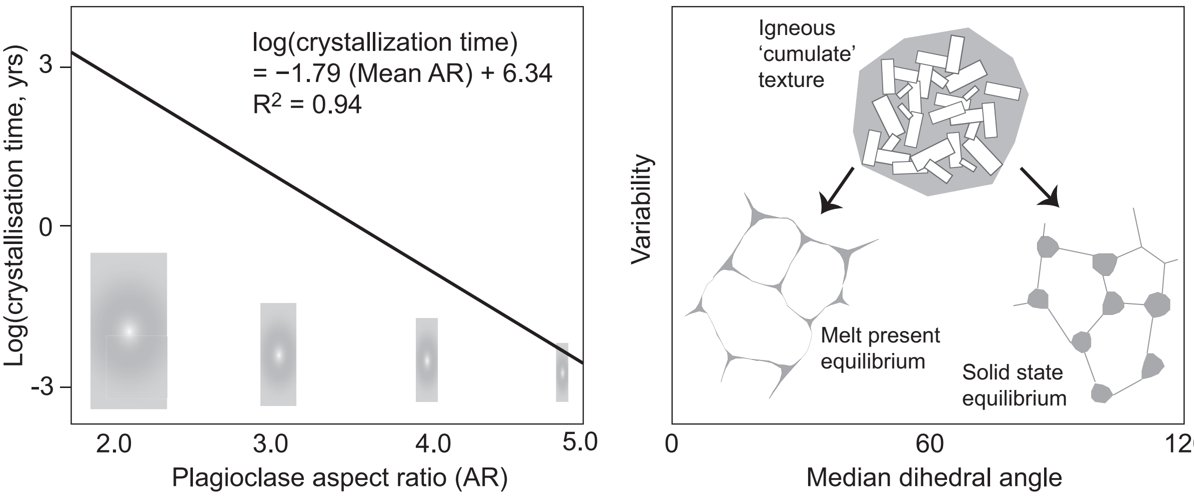

Figure 4: Petrological tools to decode microstructure. Left: plagioclase aspect ratios may be used to estimate cooling rates in volcanic and plutonic rocks (129). Dihedral angles of ~ 60° are characteristic of textures arising from igneous processes ('impingement'), arising from crystal settling, which may equilibrate with time in the mush pile (130, 131).

The other end member of magmatic storage is where magmas are stored on long timescales as a crystal-rich mush (1, 2, 9, 113, 124, 125). In this scenario, liquid-rich regions form by either expulsion of melt from the mush pile during compaction (5, 6) or gas-driven filter pressing (7), and may be enhanced by partial melting of the mush during reharge by hotter magmas (22, 29, 126, 127). Under these conditions, phenocrysts initially grow suspended in a liquid, but may experience protracted periods of storage and further crystal growth in the mush. Is there evidence for this latter scenario in erupted rocks? Holness *et al*, in this issue, show it is possible to 'decode' the microstructural features of plutonic rocks and crystalline enclaves or xenoliths erupted in volcanic rocks to infer the environment in which the crystals formed (118, 128). The aspect ratio of plagioclase crystals, for example, records the amount of undercooling experienced by the magma during growth (129) (**figure 4**). Dihedral angles, in contrast, are affected by cooling rate: during rapid crystallisation grains impinge upon one another with planar faces, which generates a wide range of disequilibrium dihedral angles and resulting a high total interfacial free energy (130, 131). At low cooling rates, crystal-melt boundaries can evolve towards the curved interfaces indicative of low interfacial free energy, with equilibrium dihedral angles at two-grain and three-grain junctions. In this way, the characteristic microstructures of plutonic xenoliths in volcanic systems can be interrogated to assess whether their constituent grains formed in a liquid-rich environment or *in situ* in a mush (132). Such microstructural evidence points towards the existence of liquid-rich fractionating sills of thicknesses up to ~800 metres for some Galapagos volcanic plumbing systems (Wolf and Ecuador volcanoes) (132).

A common feature of many volcanic deposits is the occurrence of polyphase crystal clusters, or glomerocrysts (133-136). These may be fragments of disaggregated crystal mush that has experienced open system mixing (see above; refs), or they may arise from synneusis (Schwindinger, & Anderson, 1985). The mechanisms and likelihood of crystal clustering arising from these processes are evaluated by McIntire *et al*. in this issue (137).

Mush processes: recharge, reactive flow, disaggregation, segregation and mingling

There is now abundant evidence from erupted magmas in all settings for both fractional crystallization and mixing; indeed it appears that mixing is near-ubiquitous (3, 138), reviewed by Cashman and Edmonds in this issue. Phenocrysts grown during fractional crystallization as grains suspended in the liquid are commonly euhedral and have rims, and perhaps cores, in equilibrium with the carrier liquid. Antecrysts, in contrast, may display disequilibium features such as rounding and resorption, reverse zoning or compositionally stepped rims. Antecrysts picked up from crystal layers, or mush, can be identified texturally by their deviations from linear on crystal size distribution plots (39, 139), and sometimes by deformation textures such as dislocations (140). Trace element systematics of antecrysts may show trace element concentrations that are not in equilibrium with the carrier liquid, but are in equilibrium with a related liquid from the same magmatic system (141, 142). The clear evidence that whole rock compositions merely represent an ‘average’ composition means that to really understand the processes of hybridisation and the tempo of magma assembly, one must turn to microanalysis of the components.

An example of extensive antecryst incorporation comes from Kīlauea Volcano, where the cores of olivine macrocrysts are commonly not in equilibrium with their carrier liquids (54). Instead, olivine cores are usually more primitive, indicating that they were ‘picked up’ from some (less evolved) part of the magma reservoir. Comparison of summit and rift magmas, erupted both effusively and explosively, shows that the Kīlauea magma reservoir is compositionally zoned, and that large volume eruptions excavate deeper into the reservoir system and therefore erupt more primitive olivine compositions, as well as more mafic (hotter) melt (55). The most primitive olivines at Kīlauea are also typically the largest, and show evidence of dislocation creep during deformation, suggesting they formed part of a mush system beneath the summit caldera subject to loading and/or shear (143).

Other volcanoes show some of the same features, suggesting that magma mixing, and perhaps disruption of crystal mushes to entrain their more primitive crystals, is a common process beneath active volcanoes (144).

Icelandic magmas, for example, have olivine-hosted melt inclusions that are related geochemically to the liquid that carried the crystals to the surface (145), reviewed by Maclennan in this issue. Thus even though the crystals themselves are out of equilibrium with the carrier liquid, the melt inclusions record an earlier history of melt ponding and cooling. This is further evidence that the crystals are antecrysts that are related to the liquid by fractionation but are not in equilibrium with that liquid. One interpretation of these observations is that magmas in Iceland do not interact with mush systems other than their own on the way to the surface (66), consistent with microseismic evidence of brittle fracture accompanying melt migration between sills (11).

Trace element signatures in olivine-hosted melt inclusions can further be used to constrain the dimensions of the magma bodies in which they were formed. If the variability in trace element ratios and isotopic ratios in the melt inclusions is inherited from mantle-derived heterogeneity (145, 146), then the melt inclusions were trapped during concurrent cooling and mixing (145). Thermal models for sill cooling, convection and mixing, coupled with thermodynamic models of fractionation, place constraints on sill thickness. It emerges that these magma bodies must be less than 10 metres thick to produce the observations (66), which involves production of 4 m thick layer of mush over a timescale of ~ 1 month, reviewed by Maclennan in this issue.

Evidence for antecryst entrainment is not restricted to basaltic volcanic centres, but instead are common features of rhyolites from caldera-forming eruptions in continental settings (147) as well as a range of magma types in arc settings (148, 149); similarly widely variable Ba contents of feldspars in East African Rift ignimbrites hint at the entrainment of feldspars which grew in a range of environments (pressure, melt composition) prior to eruption (Iddon et al., in press), suggesting that magmas are commonly assembled by mixing between mush components, intruding and resident liquids. In fact, arc magmatic systems preserve clear evidence for extensive mixing and mingling in the form of resorption textures, sieved plagioclase, antecrystic clots and clusters, mafic enclaves and other signs of disequilibrium (133, 134, 150-154). Cashman and Edmonds (this issue) explore the heterogeneity in major elements of arc melts (glasses) inherent in magma extracted from different pressures in the arc crust, which record different liquid lines of descent owing to the influence of H_2O on the crystallisation path.

The minerals in intermediate arc magmas also record information about magma petrogenesis that can be used to infer the compositions of melts that existed earlier in the magma's history. One example is evidence for 'cryptic' crystallization of amphibole at depth that is later 'erased' from the system through mixing and fractionation during magma ascent (155). Humphreys *et al.* (this issue) use amphibole compositions, which hold a large range of trace elements and are stable over a large pressure range, to infer deep crustal mixing and differentiation of andesites (138). The composition of 'amphibole equilibrium melts' (AEMs) may be calculated using trace element partitioning schemes (Humphreys *et al*., in review). The amphiboles preserve a record of multiple interactions (mixing episodes) between mafic melts and evolved melts (dacites); differences in the AEMs and whole rock compositions, in particular, reflect assimilation of plagioclase and zircon.

Mid-ocean ridge environments (both fast- and slow-spreading) may also be underlain by extensive regions of mush (156), although the presence and significance of mush beneath mid-ocean rudges is debated (66). Mid-ocean ridges are ideal settings to study magma storage and the importance of crystal mush: MOR basalt compositions are well understood and reasonably uniform (157, 158); there is abundant geophysical imaging data (156); and ophiolite sections allow characterisation of the layering and structure through the entire crust (97). These different data streams do not yet provide a single interpretation of magmatic processes. One view is that melt ascends through the mush through porous flow, driven by buoyancy and through expulsion by compaction (McKenzie, 1985), rather than by dyking. As melts ascend, they will inevitably flow through material with which they are not in chemical equilibrium, as reviewed by Lissenberg *et al*. in this issue. This will manifest (on a grain scale) by evidence of reaction such as dissolution, symplectites, reaction rims or complex mineral zoning. Geochemically, reactive flow will be recorded in the trace element zoning in crystals close to the channels of melt movement (159, 160), giving rise to heterogeneous and trace-element-enriched melt compositions. These melts may mix and homogenise during flow, leaving Mg# relatively unchanged; they may also entrain diverse phenocrysts and antecrysts from the mush during transport. Reactive flow also helps to explain the origin of the vertical fabrics observed in ophiolites (161). Reactive flow may involve both evolved and more primitive melts, and the extent of reactive homogenisation in a single mush body depends on allowable diffusion time scales. The resulting range of textures, zoning and compositions thus produced allow reconstruction of the range of melt compositions produced during reactive flow (159). An alternative view arises from examination of exposed ophiolite sections in Oman, which led Korenaga and Kelemen (15) to conclude that the cumulate rocks of the lower crustal section formed in small, open system melt lenses, and that melt flux through the mush was not important in supplying the total melt flux for the system. Taken together, outstanding questions surround the importance of reactive porous flow in mush systems beneath MORs.

Time-dependent nature of magma reservoirs

Important questions about the behaviour of magma reservoirs concern the timescales of various processes, reviewed by Cooper in this issue. These timescales are important in evaluating volcanic hazards, interpreting volcano monitoring data and, on longer timescales, developing models of likely repose periods for large volcanic systems. The timescales of magma supply, fractionation, mixing and recharge events also have relevance for broader questions relating to the timescale of crustal growth and of the creation of ore deposits in the crust. We might ask: how long does it take to make a magma reservoir? How long are magma reservoirs active? And of that time, how long are they super-solidus? On what timescale are magmas remobilised and ‘assembled’ prior to eruptions? How fast are melts extracted from mushy regions to form eruptible liquid volumes?

Progress in developing methods and models to generate timescale data in magmatic petrology, is reviewed, along with their limitations, by (162), as well as by other authors (163, 164). Diffusion modeling of compositional steps in volcanic crystals has emerged as a critical method to establish the age of 'young' (10^{-1}-10^{3} years) mixing events in magmas (165, 166), made possible by advances in microanalysis as well as in modeling diffusional processes. Most commonly modelled are the coupled diffusion of Fe and Mg as well as Li, Sr, Mg and Ba diffusion in plagioclase (167). Importantly, the timescale recorded by the diffusive relaxation of a compositional profile in a mineral is a function of temperature, and as such may record only a fraction of a mineral's total age (3). When combined with other petrological methods, diffusion chronometry becomes even more powerful. Diffusion chronometry on crystal cargoes erupted in Iceland, combined with barometry, shows that, at least in some cases, magmas may rise rapidly (~ days) from crustal storage areas, which may be deep (~ 20 km; Mutch et al., in review). These short timescales almost certainly require a brittle rock matrix and the opening of dykes, consistent with microseismicity observed in Iceland (11). These short timescales contrast with timescales of storage in these deep reservoirs, which may reach 1000 years (Gutai et al., in review).

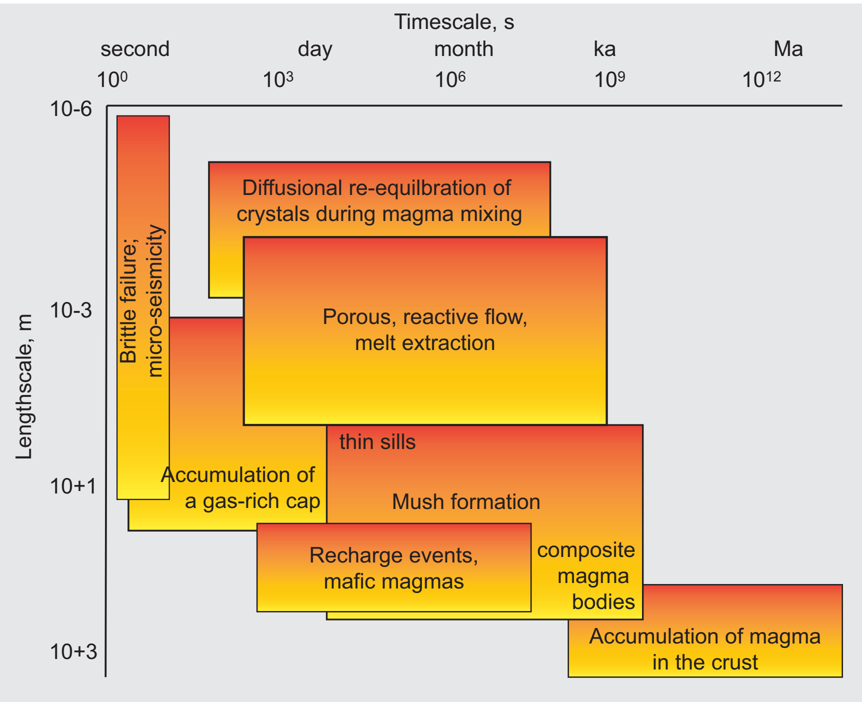

Figure 5: Time- and lengthscales associated with different magmatic processes in the crust; discussed in the text.

Radiometric dating systems, such as U-Pb and U-series of accessory minerals such as zircon and titanite (164, 168, 169) record longer ages, corresponding to the time since the mineral fractionated parent isotopes and became closed. Such a radiometric 'clock' operates only below its closure temperature. Crystallization ages of the minerals in magmas erupted from volcanoes are 10^{4}-10^{5} years (162). Remarkably, this entire age span can be recorded by the growth zones of individual crystals (commonly zircon), showing that magma reservoirs can be very long-lived.

These diverse sampling of timescales record different aspects of the system. The shorter, diffusional re-equilibration timescales tell us about mixing and magma ascent prior to eruption (170). Multiple subsets of crystal populations, and multiple zones inside crystals, may allow the reconstruction of complex series of mixing events that involve different populations of crystals (165, 171, 172). In general the shorter timescales are generated in the more mafic systems, suggesting that mixing occurs closer to eruption, perhaps consistent with the lower viscosity and higher flow rates of such systems. Older ages derived from the radiometric dating of crystals, in contrast, tell us about the longevity of the reservoir and the timescales of crystallistion. In tandem with thermal modeling, the integrated picture, from many volcanic deposits worldwide, shows that magmas spend much of their time close to or even below their solidus temperatures, and are heated in short bursts, which may or may not result in eruption (3).

Importance of the exsolved volatile phase

Volatiles typically make up only a few wt% of magmas in the crust, but they have enormous importance in controlling phase equilibria, buoyancy, and physical and rheological properties of magmas, including the mush (173-176). In magma reservoirs in the crust, dissolved volatiles exsolve if fluid saturation is reached (177), where the confining pressure becomes equal to or less than the sum of the partial pressures of volatiles in the melt. Fluid saturation may occur during decompression ('first boiling') or during isobaric crystallisation ('second boiling'). Second boiling dominates in magma reservoirs, producing an exsolved magmatic volatile phase (MVP) that is initially rich in CO_2 and becomes more hydrous as crystallisation proceeds (see Edmonds and Woods, in press, for a review). Evidence for the existence of an MVP in magma reservoirs includes: miarolitic cavities in plutonic rocks (177), magma compressibility inferred from muted ground displacements ((178-181), the existence of low V_P/V_S regions beneath calderas such as Yellowstone (182).

How does this exsolved volatile phase behave in magma reservoirs, particularly in the mush-dominated portions of such reservoirs? Degruyter *et al*. (this issue) review the various regimes that may allow the MVP to outgas from crustal magma reservoirs, building on previous work building numerical models at the pore-scale of how gas may migrate through crystal-rich magmas (22, 127, 175, 183, 184). Channel flow by viscous fingering may occur in a permeable crystal network at intermediate melt fractions (Parmiagiani et al., 2014; (185). For crystal fractions that exceed the 'locking point', high overpressures are required to overcome the strength of the crystal framework to propagate fractures through the mush to allow outgassing (186), which may be possible at low pressures. In melt-rich lenses in the magma reservoirs, in contrast, the MVP may get 'held up' during relatively slow Stokes' rise of bubbles (184, 187), such that melt lenses may be areas in which MVP can accumulate; this process may be the basis of the formation of gas-rich caps to magma reservoirs.

It follows then, perhaps rather counter-intuitively, that gas-rich regions may develop in the shallow reaches of magma reservoirs, because of the relatively fast migration of the MVP through the crystal-rich regions, and accumulation in the melt lenses (184). This mechanism might explain the large emissions of volcanic SO_2 that often accompany explosive eruptions (180, 188-192), the persistent degassing between eruptions (193) and the formation of hydrothermal ore deposits in the roof zones of magma reservoirs (194). 'Gas-rich caps' may be the primary cause of ground displacements, with outgassing causing subsidence, and inputs of magma fluids into a hydrothermal system generating cyclic uplift and subsidence e.g. for Aluto, East African Rift, and Campi Flegrei, Italy (195, 196). Outgassing of accumulated exsolved volatiles in the upper parts of volcanic systems may even generate downward-propagating instabilities which might induce mixing and mingling (17), or even eruption (197). The MVP may also play an important role in generating melt lenses in crystal mushes, particularly in the upper crust, where gas-driven filter pressing may cause expulsion of melt from a crystal framework because of expansion of the MVP phase during second boiling (7). Additionally, if an MVP develops in a crystal mush, it greatly alters its rheological properties, allowing it to be rapidly remobilised (8, 19, 20).

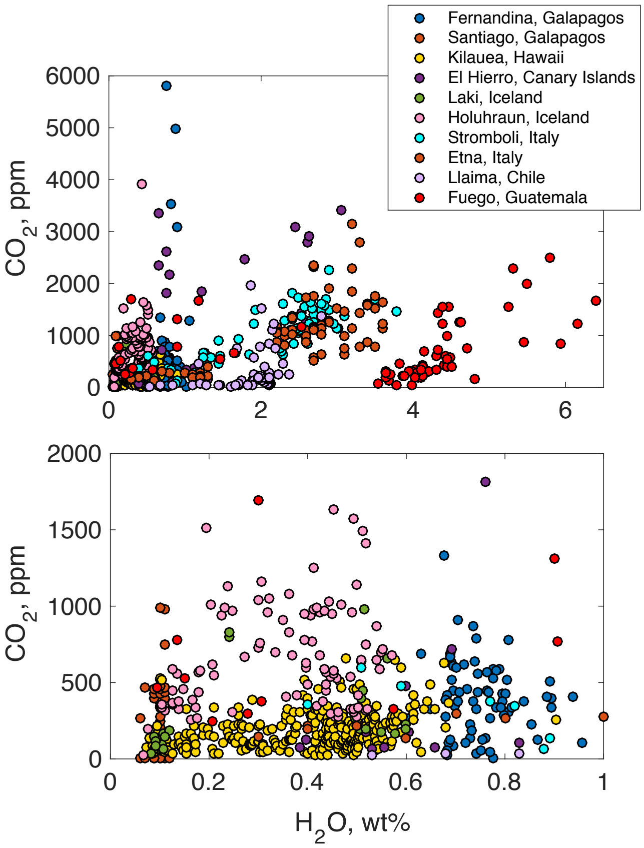

Figure 6: The differences in water contents between arc and hotspot melts may lead to fundamental differences in the form and extent of the magma reservoirs in these settings. Magmatic systems in arc settings will generate larger fractions of exsolved volatiles; and will undergo degassing-induced crystallisation at greater depths in the crust (144), which promotes the generation of crystal mushes. Olivine melt inclusion concentrations of H_2O and CO_2, for a range of ocean island volcanic products. The bottom plot is a zoomed in portion of the top plot. Isobars, labelled with pressure in MPa, are shown, appropriate for basalt (198). Data for Fernandina and Santiago from (199); for Kīlauea from (54), for El Hierro from (200) for Laki (89) , for Holuhraun (87), for Stromboli (201), for Etna (202) for Llaima (203, 204) and for Fuego (205-208).

Figure 6 shows a compilation of basaltic melt inclusion water contents. The melt inclusions trapped melt prior to (most of) the degassing of water. The data show that pre-eruptive melt water concentrations may vary between 0.1-0.3 wt% (for mid-ocean ridges and relatively dry ocean island melts) up to 3-6 wt% for arc basalts and some wet ocean island melts (e.g. Canaries). The water content of melt, which varies with tectonic setting (**figure 6)**, probably exerts an important control on on the architecture of magmatic systems (Cashman and Edmonds, this issue). In relatively dry magmatic systems (<0.5-1.0 wt% H_2O), stacked sills with incompressible melts form in subsolidus country rocks; melt accumulates near the seismic Moho when magma input rates are low, and in the upper crust when magma input rates are high. In contrast, water-rich arc magmatic systems are typically vertically extensive and dominated by mushy super-solidus conditions except at shallow levels, where compressible gas-rich caps and lenses are common. An important but poorly constrained region in all magmatic systems is the magmatic-hydrothermal interface, which is an important location of heat and fluid transfer.

Modeling magma reservoir processes

How is a magma reservoir built?

Magma reservoirs may be built, over long timescales, by repeated injections of magma (33, 209-211). Modeling shows that emplacement of basaltic sills into the (amphibolitic) lower crust generates zones of partial melting, with mixing of these anatectic melts with the evolved melts derived from the basalt intrusion. The relative importance of these two components depends on magma supply rate, but melt differentiation is fundamentally driven by melt migration upward through the partially molten matrix by reactive flow (210). To build a long-lived zone of partial melt in the crust, it is necessary to supply more heat by advection than is lost by conduction (2, 33). Numerical models show that it is easier to form persistent regions of melt-bearing crust (reservoirs) in the (hotter) lower crust than in the cooler upper crust (33). The magma supply rates should not, however, be sufficiently high to trigger eruptions and thereby prevent magma accumulation. The development of extensive mush regions in the lower crust in the models is consistent with prevailing views of lower crustal fractionation, mixing and assimilation based on magma geochemistry (212). The inevitable involvement of crustal partial melts in melt differentiation also explains the pervasive crustal signatures observed in the geocehmical characteristics of granites and eruptions from large rhyolite bodies (213, 214).

It has been suggested that large, upper crustal magma reservoirs may form only if a sufficiently large lower crustal mushy reservoir exists (215). Such a lower crustal reservoir is likely to be active over timescales of 10^5-10^6 years and may be similar to that imaged beneath Yellowstone caldera (64). The presence of this lower crustal magma reservoir elevates the upper crustal geotherm sufficiently that an upper crustal magma reservoir can form with magma fluxes that are 1-2 orders of magnitude lower than for a cold upper crust. Under these conditions, the upper crustal mush can remain stable for 10^5-10^6 years (215), which is in line with the radiometric ages of crystals erupted from large continental magma bodies (162, 164). To generate large eruptible volumes (similar to those erupted by 'supervolcanoes' (216)) requires not only high magma flux, but also protracted magmatic activity (215). These concepts are likely to prove useful in modeling the formation of hydrothermal ore deposits in crustal environments.

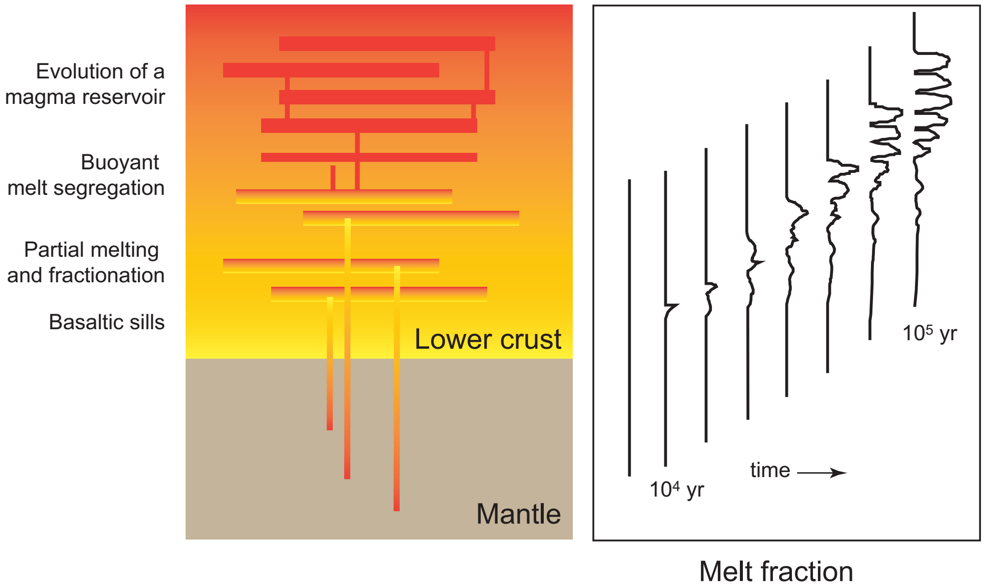

Figure 7: Models have been developed to simulate emplacement of successive sills in the lower amphibolite crust (33, 209, 210). The sills crystallise, and heat the surrounding crust, eventually generating melting. The mixed melts (residual melts from the basalts and partial melts of the crust) segregate by buoyant flow along grain boundaries. Over time, repeated injections of basalt leads to upward-migrating permanent melt lenses in the lower crust (210). There are important implications of such a process: melts differentiate by reactive flow during upward percolation; and there is substantial assimilation of crustal melts.

Modeling mush recharge, mixing and disaggregation

The evidence discussed in earlier sections suggests that crystal-rich magmas are often disrupted by intruding melts (and perhaps erupted) on timescales of years to decades. How may a locked crystal mush be remobilised on such short timescales? What are the material properties of the mush? How does it deform in response to stress? Numerical models have been developed to answer some of these questions (27, 217), based on particle-particle-fluid coupling and interactions at the crystal scale. The results indicate that the mush may become fluidised, with crystals entrained by the intruding melt and efficiently mixed. This and other mixing processes that can occur during ascent and shallower storage (e.g. laminar flow ina shear zone, perhaps in the conduit) may explain the co-existence of antecrysts and phenocrysts that have experienced very different petrogenetic histories (29, 134, 154).

Magma intrusion into pre-existing, partially solidified sills, can create complex density profiles, as reviewed by Woods and Stock in this issue. For example, density-stratified cumulate layers may be interspered with more homogeneous layers. Crystal-rich suspensions also have complex rheologies. Crystal mush with a touching framework of crystals can support a stress. Intrusion of a liquid–rich magma into the crystal mush, however, may dilate it sufficiently to mobilise the suspension, thereby allowing new melt to intrude (218). The heterogeneity in crustal architecture produced by successive intrusions then controls the loci of successive intrusions, and the ability of magmas to ascend through the crust (218). Other factors that may affect the local formation and ascent of magmas include compaction waves and buoyancy layers in the mush (which may be melts or exsolved volatiles) which may be unstable and overturn, reviewed by Sparks *et al* in this issue (17). On longer timescales, magmatic systems may form stably stratified layers as felsic melts ascend to shallow depths, thus creating the structure observed in exposed crustal sections (100).

Applications to volcano monitoring

What implications does our new understanding of the depth and form of magma reservoirs have for monitoring (and forecasting) eruptions? The previous sections summarise geophysical and petrological evidence for multiple magma storage areas in the crust, and deep storage regions that are vertically and/or laterally extensive. Magmas fractionate and mix when injected into mushy regions, where they may pick up crystals from a transiently mobilised mush. Magma may be compositionally zoned within a single melt lens as well as throughout the crust, and may comprise mush-dominated, or even sub-solidus, over long times. Near-solidus magma reservoirs are remobilised by intruding (more mafic) magmas that may entrain individual crystals, glomerocrysts or cognate xenoliths from the mush, which may then be erupted. Importantly, intrusions of more primitive magma do not always produce eruptions, even when they are volatile-rich. Instead, exsolved volatiles may accumulate in the upper melt-rich parts of the magma systems, particularly in intermediate and evolved magmatic systems found in subduction zone and continental settings. Migrating volatiles may then trigger instabilities, magma mixing and bursts of outgassing, with or without eruptions. Alternatively, if volatiles are not outgassed, they may contribute to the formation of hydrothermal ore deposits.

How does this heuristic model of magmatic system behaviour help to interpret observations of volcanic behaviour and the signals measured at the surface during periods of unrest and volcanic eruptions? Pritchard *et al*. (this issue) explore the implications of a *trans-crustal magmatic system* (9) for evaluating signals – seismic, deformation, outgassing – associated with non-magmatic unrest, and those which may be precursors to volcanic eruptions. Generalised models for pre-eruptive seismicity (219, 220) include deep (> 5km) Volcano Tectonic (VT) earthquake swarms followed by shallower Long Period (LP) events and tremor leading up to eruption; the latter are thought to be associated with the flow of gas-rich magma. These trends are genrally consistent with upwards magma transport from the mid-crust prior to eruptions, perhaps triggered within a magma body by pressurization induced by second boiling (26) or by intrusion of deep magma. In detail, however, stress changes can also propagate downwards after the onset of an eruption (221), and demonstrate the importance of pressure changes caused by both conduit development (Scandone et al., 2007) and emptying of shallow magma storage regions; this may happen sequentially when the magmatic system comprises stacked sills (221). These kinds of trends were also observed after the onset of the 1959 Kīlauea Iki eruption, where deep VT swarms (> 30 km) occurred after the onset of the eruption (222). Critically, a considerable quantity of melt be stored simultaneously in lenses at different depths in the crust, and perhaps also in the uppermost mantle; this suggests that very large eruptions may not require magma accumulation into a single very large magma chamber, but instead may result from tapping multiple vertically or laterally oriented magma bodies (Cooper et al., 2012; Cashman and Giordano, 2014; Swallow et al., 2018).

Ground displacements measured by InSAR and well-populated GPS networks are also variable, with inversions varying from simple geometric sources to multiple, complex sources (49, 223), often at a range of depths. Additionally, ground displacements are frequently muted, and not as large as expected for the volume of erupted magma. This can be explained by magma compressibility, whereby the presence of an exsolved gas phase allows the magma to contract as pressure increases, diminishing the displacment signal observed at the surface (75, 178, 179, 224). This compressibility effect has recently been linked directly to gas emissions during explosive eruptions, an analysis which shows that shallower and more water-rich magmas typically show this compressibility signature, and these eruptions are often linked with high gas emissions (180). Improving interpretations of ground deformation data thus requires a better understanding of the role of the exsolved gas phase in modifying the physical properties of the magma reservoir, reviewed by Segall in this issue.

Post-eruptive subsidence can be linked to a viscoelastic response associated with a ductile region of melt-bearing crust (225-228). The deformation consequences of a viscoelastic aureole surrounding a spherical magma chamber can be modelled (Segall, this issue). This scenario might correspond to the case where magma is intruded and heats the surrounding country rock, or when magma intrudes (or leaves, during explosive eruption) a mushy region of the crust. Interestingly, the post-eruption inflation may occur (without recharge) if the magma is volatile-free or volatile-poor (i.e. incompressible); this result holds for both oblate and prolate geometries. If the magma is sufficiently compressible, post-eruptive deformation will be deflation. These models raise the possibility of reconciling petrological inferences related to pressure changes with observations of deformation. In fact, the timescales derived from diffusion studies described in section 5 have been linked directly to the occurrence of earthquake swarms and deformation prior to, and during, eruptions (162, 165, 229, 230). These correlations shows that the monitoring signals are linked to significant pressure and/or temperature perturbations, such as those expected for magma movement and/or mixing in shallow crustal reservoirs.

Gas emissions during and between eruptions also hold clue to processes occurring in crustal magma reservoirs. Some volcanoes exhibit strong correlations between the mass of erupted magma and the mass flux of SO_2 (e.g. Kīlauea; (231)), suggesting that degassing occurs in a shallow reservoir and that gas is advected by the magma to the surface, making the SO_2 flux a good proxy for eruption rate. In other settings, however, the patterns of gas emission are rather different. At many volcanoes, an elevated gas flux is maintained between eruptions, e.g. at Bagana, Etna, Stromboli, Colima, Popocatepetl volcanoes (232, 233). At open-vent basaltic volcanoes, convection of magma in a conduit provides a sustained and steady supply of gas to the atmosphere e.g. at Masaya, Stromboli, Villarica, Nyiragongo, Erta Ale, Ambrym volcanoes (234, 235). Superimposed on this steady supply is the arrival of large CO_2-rich gas slugs and bubbles derived from deeper in the system); this is typical of volcanoes exhibiting strombolian activity such as Stromboli and Yasur volcanoes (236). Steady state convection requires deep magma storage bodies for degassed magmas to accumulate, thereby growing the crust endogenously (237). The formation of gas slugs further requires re-supply of CO_2-rich magma at depth, most likely sourced from gas segregation in sills (238).

At intermediate volcanoes, where magmas undergo significant decompressional degassing and crystallization and where the magmas are more evolved and crystal-rich, gas may accumulate throughout the plumbing system. Gas migration in these systems varies with suspension crystallinity and gas volume; importantly gas migration through crystal-rich mush can be slow, under conditions where bubbles are trapped within the mush (239) or fast, where gas migrates through open, quasi-brittle fractures (184, 185). Sustained, open pathways that allow the passage of deep-derived magmatic gases to the surface have been invoked to explain the high and continuous emission of SO_2 gas at Soufriere Hills Volcano, where it is known that sulfur partitions into an exsolved vapor phase at <5-7 km depth (193, 240, 241). These pathways may also exist at other volcanoes exhibiting sustained degassing with little eruptive flux.

Conclusions and future challenges

Summary and conclusions

- In this volume we bring together a range of disciplines and types of observations and models to define the state of our understanding of magma reservoirs, and future directions of research. The results of our discussion meeting have highlighted a range of possible directions. It is clear that geophysical methods, combined with petrological barometry, offer a powerful way to delineate the depth of magma bodies, but the information obtained from these methods is not strightforward to reconcile, given the differences in temporal and spatial scales.
- The evidence for magma mixing is ubiquitous in all tectonic settings, and includes mixed and mingled crystals and melts, as well as entrainment of mush components in intruding melts, which may then be erupted at the surface. There are still fundamental questions about whether and when ascending melts may intersect ‘foreign’ mushes which are not products of their own liquid line of descent. In most cases it appears that entrained antecrysts are related genetically to the carrier liquids.
- There remains considerable heterogeneity in reservoir architecture from place to place, which ranges from extensive mush-dominated magmatic systems (in the arc crust), to multiple dispersed sills in the country rock (or sub-solidus parts of the reservoir) that may contain magma prior to and between eruptions (Iceland).

- The water content of melt exerts an important control on the architecture of magmatic systems. Relatively dry systems appear to be dominated by stacked sills with incompressible melts in subsolidus country rocks; water-rich systems, in contrast, develop vertically extensive mushy super-solidus reservoirs with compressible gas-rich caps and upper crustal lenses.
- Timescales of magmatic processes point to prolonged storage (103-106 years) in the crust, followed by rapid remobilisation and ascent. This remobilisation may occur from shallow storage regions (in evolved systems) or deep sills, perhaps even close to the Moho (in ocean island settings).
- Exsolved volatiles are an important part of the magmatic system, particularly in arc settings. Here, gas-rich regions may form where gas is 'held up' in liquid lenses, but may also be present in the mush and thus may affect its rheological properties. The gas may also generate instabilities in the magmatic system, which may cause local mixing, or even eruption.
- Models describing magma reservoir construction show that important parameters include the magma supply rate (coupled with the accommodation space available) and the thermal state of the crust. Preconditioned upper crust may be most favourable for the formation of large magma reservoirs required to feed supervolcanic eruptions.
- Models of magma emplacement into viscous mush show that complex layering and instabilities may form on crustal scale, owing to the thermal and mechanical effects of intrusion, partial melting, cooling, convection, mixing, degassing and crystal settling. On a particle scale, high input rates of magma recharge into a mush may trigger chaotic mixing and entrainment of crytal cargo.
- Our new understanding of trans-crustal magmatic systems will prove useful for interpretation of volcano monitoring data. Improved seismic networks may show that downward propagating seismic swarms after the onset of eruptions are common. Such events may even be detected during periods of unrest, which will tell us something about intrusion and instabilities in the magmatic system, particularly when combined with other observations of the volcanic system. Increasingly sophisticated modeling of viscoelasticity means that we can now attempt to explain some of the time-dependent signals that occur after eruptions. Volcanic degassing fluxes (typically SO_2) may be used to infer deep gas accumulation and release processes, particularly for intermediate volcanoes, but also for mafic open vent volcanoes that emit a lot of gas but little magma, requiring gas-melt segregation and the accumulation of large volumes of mushy cumulate magma at depth.

Future Challenges

This issue contains new observational constraints on micro-scale textures and compositions of magma components, on thermal conditions of storage and timescales of mobilization from dating and diffusion modeling, and results from a new generation of physical and numerical models which examine processes of heat and mass transfer in crystal mush at the grain scale. These studies offer the potential to make rapid progress in understanding the dynamics of magma reservoirs and the eruptions that they produce.

Outstanding questions relate to the mechanisms and timescales of liquid extraction from mush and are summarised below.

- An outstanding question is 'does compaction operate in crustal systems, and if not, what are alternative processes?' How is the physical process of melt extraction modified by reactive flow, production of latent heat and the presence of an MVP?
- How do we detect structures and processes, some at quite small scale, in ductile magmatic environments? What laboratory experiments and advances in theory are needed to constrain the key physical properties and dynamics of magma reservoirs?
- What models need to be developed to understand trans-crustal mush system dynamics, including the magma fluxes, timescales, volumes produced during episodic volcanism? Can these models be coupled to conduit and surface flow models to provide the basis for forecasts? To what extent do we need to re-examine basic assumptions in view of the emerging idea of dynamic mush-magma systems?
- There is substantial work required to better interpret volcano monitoring data. Mass must be conserved during magma transport in the crust, so when we see shallow inflation (as measured by surface ground displacements), we should see deeper deflation. As yet, however, our measurements are not precise enough. How can we image deeper magma reservoirs? A step change in technology is required.
- The inherent limitations of individual techniques requires integration of datasets to better understand magma reservoirs, which calls for sophisticated modelling approaches (242). Additionally, the growing importance of satellite-based sensors for global monitoring provides exciting opportunities, but also new tools for rapid data processing and integration with models are required.

Additional Information

Authors' Contribution
All authors contributed equally to this manuscript.

Competing Interests
The authors declare that they have no competing interests.

Acknowledgments
The authors acknowledge the Royal Society for the funding provided to hold the Theo Murphy Royal Society Discussion Meeting at the Kavli Centre in November 2017. NERC COMET are acknowledged for providing support for the meeting.

Figure and table captions

Figure 1: Schematic cartoon of the magma reservoir end member types that may exist in A: continental and arc settings (38); and B: in Iceland (discussed in text) (39).

Figure 2: Last equilibration pressures of magma from clinopyroxene-melt barometry (95, 96) for a range of ocean island volcanoes (89-95).

Figure 3: Schematic sketches of exposed arc sections of Talkeetna (Alaska, USA) and Kohistan (Pakistan) (100). Brown represents volcanic rocks, orange are intermediate to felsic plutonic rocks (53–72 wt% SiO_2), red is

mafic lithologies (gabbro cumulates) (100). The Moho is defined as the transition between ultramafic rock (plagioclase absent) and gabbroic rock (plagioclase present). The righthand four sections are a summary of seismic velocities for a range of arcs (100): Aleutian (106), Kurile (107), Cascade (108) and Izu-Bonin (109).

Figure 4: Petrological tools to decode microstructure. Left: plagioclase aspect ratios may be used to estimate cooling rates in volcanic and plutonic rocks (129). Dihedral angles of ~ 60° are characteristic of textures arising from igneous processes ('impingement'), arising from crystal settling, which may equilibrate with time in the mush pile (130, 131).

Figure 5: Time- and lengthscales associated with different magmatic processes in the crust; discussed in the text.

Figure 6: The differences in water contents between arc and hotspot melts may lead to fundamental differences in the form and extent of the magma reservoirs in these settings. Magmatic systems in arc settings will generate larger fractions of exsolved volatiles; and will undergo degassing-induced crystallisation at greater depths in the crust (144), which promotes the generation of crystal mushes. Olivine melt inclusion concentrations of H_2O and CO_2, for a range of ocean island volcanic products. The bottom plot is a zoomed in portion of the top plot. Isobars, labelled with pressure in MPa, are shown, appropriate for basalt (198). Data for Fernandina and Santiago from (199); for Kīlauea from (54), for El Hierro from (200) for Laki (89) , for Holuhraun (87), for Stromboli (201), for Etna (202) for Llaima (203, 204) and for Fuego (205-208).

Figure 7: Models have been developed to simulate emplcement of successive sills in the lower amphibolite crust (33, 209, 210). The sills crystallise, and heat the surrounding crust, eventually generating melting. The mixed melts (residual melts from the basalts and partial melts of the crust) segregate by buoyant flow along grain boundaries. Over time, repeated injections of basalt leads to upward-migrating permanent melt lenses in the lower crust (210). There are important implications of such a process: melts differentiate by reactive flow during upward percolation; and there is substantial assimilation of crustal melts.

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