Crystal and volatile controls on the mixing and mingling of magmas

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Abstract

The mixing and mingling of magmas of different compositions are important geological processes. They produce various distinctive textures and geochemical signals in both plutonic and volcanic rocks and have implications for eruption triggering. Both processes are widely studied, with prior work focusing on field and textural observations, geochemical analysis of samples, theoretical and numerical modelling, and experiments. However, despite the vast amount of existing literature, there remain numerous unresolved questions. In particular, how does the presence of crystals and exsolved volatiles control the dynamics of mixing and mingling? Furthermore, to what extent can this dependence be parameterised through the effect of crystallinity and vesicularity on bulk magma properties such as viscosity and density? In this contribution, we review the state of the art for models of mixing and mingling processes and how they have been informed by field, analytical, experimental and numerical investigations. We then show how analytical observations of mixed and mingled lavas from four volcanoes (Chaos Crags, Lassen Peak, Mt. Unzen and Soufrière Hills) have been used to infer a conceptual model for mixing and mingling dynamics in magma storage regions. Finally, we review recent advances in incorporating multi-phase effects in numerical modelling of mixing and mingling, and highlight the challenges associated with bringing together empirical conceptual models and theoretically-based numerical simulations.

1 Introduction: Magma mixing and mingling and volcanic plumbing systems

It is now widely accepted that magmas of different compositions can mix and mingle together (Blake et al., 1965; Eichelberger, 1980; Sparks & Marshall, 1986, Wiebe, 1987; Snyder, 1997; Wilcox, 1999; Perugini & Poli, 2012; Morgavi et al., 2019). Textural consequences of mingling have long been observed (Phillips, 1880; Judd, 1893) although the earliest observations were not necessarily interpreted correctly (Wilcox, 1999), with heterogeneities interpreted as originating from metasomatism (Fenner, 1926) or solid-state diffusion (Nockolds, 1933). Advancements in geochemical analysis combined with an understanding of phase equilibria led to acknowledgement of mixing and mingling as key processes, alongside crystal fractionation, in producing the compositional diversity of igneous rocks (Vogel et al., 2008). Additionally, interaction between magmas became recognised as a potential trigger for volcanic eruptions (Sparks et al., 1977). Evidently, understanding mixing and mingling processes is crucial for deciphering the evolution of igneous rocks and the eruptive dynamics of volcanoes.

Previous work has sometimes been flexible with regards to precise definitions of the terms 'mixing' and 'mingling'. We here define mixing to be chemical interaction between two magmas that produces a composition intermediate between the original end-members (Bunsen, 1851). Chemical mixing proceeds by chemical diffusion (Watson, 1982; Lesher, 1994) and, if allowed to complete, leads to hybridisation and homogeneous products (Humphreys et al., 2010). By contrast, mingling is the physical interaction of the two magmas, such as through convective stirring (e.g., Oldenburg et al., 1989) or chaotic advection (e.g., Perugini & Poli, 2004; Morgavi et al., 2013), and creates compositional heterogeneities. Mixing and mingling often occur together, with mixing acting to 'smooth–out' compositional heterogeneities produced by mingling. However, mixing and mingling can be inhibited by large contrasts in magma viscosity (Sparks & Marshall, 1986; Frost & Mahood, 1987; Sato & Sato, 2009) and density (Blake & Fink, 1987; Koyaguchi & Blake, 1989; Grasset & Albarade, 1994). If homogenisation is sufficiently slow then cooling and/or degassing of the system can lead to crystallisation and preservation of a variety of textural and chemical signatures (D'Lemos, 1987; Morgavi et al., 2016) reflecting the temperatures, compositions, crystallinities and relative proportions of the initial magmas (Eichelberger, 1980; Bacon, 1986; Sparks & Marshall, 1986).

Mixing and mingling models typically assume injection of a hotter, mafic magma into a cooler, more felsic host (Campbell & Turner, 1989; Clynne, 1999). This can be followed by later intrusion (or back-injection) of veins and pipes of remobilised felsic material into the mafic component (Elwell et al., 1960; 1962; Wiebe, 1992; 1994; 1996; Wiebe & Collins, 1998; Wiebe et al, 2002; Wiebe & Hawkins, 2015). Such injections have been modelled experimentally (Huppert et al., 1984, 1986; Campbell & Turner, 1986; Snyder & Tait, 1995; Perugini & Poli, 2005), theoretically (Sparks & Marshall, 1986) and numerically (Andrews & Manga, 2014; Montagna et al., 2015). Additionally, heat and volatile transfer from the mafic to the felsic end-member induces physico-chemical responses in both magmas. The mafic component undergoes crystallisation and degassing due to undercooling (Eichelberger, 1980; Cashman & Blundy, 2000; Coombs et al., 2002; Petrelli et al., 2018), leading to an increase in bulk viscosity (Caricchi et al., 2007; Mader et al., 2013) and potentially a decrease in density (if bubbles of the exsolved gas phase remains trapped), whereas the felsic magma partially melts due to super-heating (Pistone et al., 2017). This can create a temporal window where the bulk viscosities of the two magmas become closer thereby facilitating mingling and mixing before continued crystallisation of the mafic magma increases its viscosity. Another scenario is mixing and mingling between partially-molten silicic rocks and a hot, rhyolitic injection (Bindeman and Simakin, 2014), which is important for the formation of large, eruptible magma bodies containing crystals mixed from different portions of the same magma storage system (antecrysts; Francalanci et al., 2011; Ubide et al., 2014; Stelten et al., 2015; Bindeman & Melnik, 2016; Seitz et al., 2018). In all cases, the physico-chemical changes and their associated timescales govern the style of mixing, the resultant textures and the eruptive potential.

1.1 Chemical mixing

Chemical mixing occurs through the diffusion of different components along spatial gradients in chemical potential (Adkins, 1983) to create homogeneous products. If all components have equal diffusivities, the mixing of two chemically-distinct magmas gives rise to linear trends on Harker-type variation diagrams (Harker, 1909) that can be used to constrain the end-member compositions. Non-linear mixing trends produced by variable diffusivities amongst melt components, including trace elements (Nakamura & Kushiro, 1998; Perugini et al., 2008; Perugini et al., 2013), are also common and have been identified in various localities (Reid et al., 1983; Bacon & Metz, 1984; Cantagrel et al., 1984; Gourgaud & Maury, 1984; Bateman, 1995; Bacon, 1986; Coombs et al., 2000; Troll & Schmicke, 2002; Perugini et al., 2003; Choe & Jwa, 2004; Janoušek et al., 2004; Prelević et al., 2004; Kumar & Rino, 2006; Ruprecht et al., 2012; Kim et al., 2014; Weidendorfer et al., 2014). Further complexity arises from uphill diffusion in some species (e.g. Sr, Nd, Al), since diffusion is governed by gradients in chemical potential rather than concentration, and the temporal dependence of diffusivities in mixing events caused by changes in temperature and bulk composition (Lesher, 1994; Bindeman & Davis, 1999).

Evidence of mixing is preserved primarily at the microscale since the relatively slow rate of diffusion alone (Morgan et al., 2008; Acosta-Vigil et al., 2012) cannot redistribute chemical components over large spatial scales (Bindeman & Davis, 1999). Crystals, in particular, can preserve chemical records of changing storage conditions that can be associated with mixing. For instance, resorption zones and reverse zoning in plagioclase might indicate changes to more mafic melt compositions, possibly due to multiple mixing events (Hibbard, 1981; Tsuchiyama, 1985; Lipman et al., 1997). The mixing history can be determined by combining these observations with methodologies such as major-element (Rossi et al., 2019), trace-element (Humphreys et al., 2009) and isotopic analyses (Davidson et al., 2007), along with measurements from the bulk rock or other minerals. This can include timescales of mixing (Chamberlain et al., 2014; Rossi et al., 2019) and ascent (Humphreys et al., 2010), temperatures and pressures of mixing (Samaniego et al., 2011), and the relative contribution of processes such as fractional crystallisation (Foley et al., 2012; Ruprecht et al., 2012; Scott et al., 2013).

1.2 Physical mingling

Mingling results from fluid flow, either directly due to shear between two magmas during injection, or as a consequence of buoyancy-driven convection. Although mingling cannot occur in the complete absence of mixing, if convection timescales are shorter than diffusive timescales, mingling dominates the interaction and produces heterogeneities that are preserved as mingling textures if the magma cools and consolidates before homogenisation is complete. Examples include composite dykes and sills (Wiebe, 1973), intermingled layered intrusions of alternating composition (Wiebe, 1993; Wiebe, 1998), banded pumice (Clynne, 1999), and mafic enclaves (Eichelberger, 1980). Enclaves are perhaps the most widely-reported mingling texture and are widespread in both plutonic (D'Lemos, 1986; 1996; Topley et al., 1990; Blundy & Sparks, 1992; Williams & Tobisch, 1994; Baxter & Feeley, 2002) and volcanic (Bacon, 1986; Martin et al., 2006; Browne et al., 2006a,b; Perugini et al., 2007; Fomin & Plechov, 2012) settings. Enclaves are produced by disaggregation of intrusions into host magmas (Eichelberger, 1980; Thomas, et al., 1993; Tepley et al., 1999; Perugini & Poli, 2005;

Hodge, et al., 2012a; 2012b; Caricchi et al., 2012; Andrews & Manga, 2014; Vetere et al., 2015); Figure 1 illustrates the sizes, shapes and crystallinities that can result. Enclaves often contain crystals mechanically transferred from the surrounding host magma (xenocrysts), which can be interrogated to infer conditions (e.g., temperature, crystallinity, melt or bulk rock composition) at the time of mixing (Reid et al., 1983; Cantagrel et al., 1984; Wiebe, 1993; Coombs et al., 2000; Humphreys et al., 2009; Borisova et al., 2014; Ubide et al., 2014).

The multi-phase nature of magma is important for mingling dynamics. Experiments have demonstrated that the presence of phenocrysts can enhance mixing (Kouchi & Sunagawa, 1983; 1985) although a crystal framework can also inhibit efficient mingling (Laumonier et al., 2014; 2015). Crystallisation-induced degassing (Cashman & Blundy, 2000) of the mafic end-member due to heat and water loss to the felsic component (Pistone et al., 2017) causes the exsolution of buoyant volatile phases that can enhance mingling (Eichelberger, 1980; Wiesmaier et al., 2015). There is also growing recognition that magma storage systems are dominated by mushy regions with melt concentrated in isolated, possibly transient, lenses (Hildreth, 1981; 2004; Bachmann & Huber, 2016; Cashman et al., 2017; Sparks et al., 2019). Despite this, many studies continue to model mingling as taking place between two crystal-free fluids in a vat (Montagna et al., 2015). Such a picture is hard to reconcile with evidence from petrological analysis (Turner and Costa, 2007; Druitt et al., 2012; Cooper, 2017) and the lack of geophysical evidence for large extended bodies of melt (Sinton and Detrick, 1992; Miller and Smith, 1999; Farrell et al., 2014; Pritchard et al., 2018). It is therefore clear that the presence of crystals and volatiles, and their effect on magma rheology (Caricchi et al., 2007; Mueller et al., 2010; Pistone et al., 2012; Mader et al., 2013), must be accounted for when modelling mingling (Hodge et al., 2012; Andrews & Manga, 2014; Laumonier et al., 2014).

2) Controls on magma mingling: Observations, experiments, and numerical models

2.1 Field observations

Mingling textures preserved in the field record the varying extents to which magma mingling can occur. At one extreme, mafic sheets in granite plutons (Bishop & French, 1982; Topley et al., 1990; Wiebe, 1993, 1998) provide an example of individual intrusions that remain intact following injection. Multiple injected sheets can create layered intrusions that remain hot for an extended period of time, although such layers could also result from porosity waves within a mush (Jackson & Cheadle, 1998; Solano et al., 2012). When buoyant (silicic and volatile-rich) layers underlie mafic sheets, irregular protrusions or pipes of felsic magma are generated by gravitational instabilities and can penetrate the overlying mafic sheets (Fig. 2; Elwell et al., 1960; 1962; d'Ars & Davy, 1991; Snyder & Tait, 1995; Caroff et al., 2011). By contrast, examples of intimate mingling include compositionally-banded pumice (Clynne, 1999; Andrews & Manga, 2014), which might have hybridised fully had eruption not interrupted the mixing process. Enclaves represent an intermediate outcome between layered intrusions and banded/hybridised products and are the preserved fragments of a disaggregated mafic intrusion into a more felsic body. Some, but not all, show fine-grained quenched margins and coarse, vesicular cores suggesting slower cooling towards the centre of the enclave (Eichelberger 1980; Bacon & Metz, 1984; Bacon, 1986; Blundy & Sparks, 1992; Browne et al., 2006a).

In mingled rocks, it is common to find crystals derived from one mixing end-member residing in the other (Fig. 3). Such xenocrysts have been found in composite dykes (Judd, 1893; King 1964, Prelević et al., 2004; Litvinovsky et al., 2012; Ubide et al. 2014), mafic sheets (Wiebe, 1993; Wiebe & Collins, 1998; Bishop & French, 1982; Topley et al., 1990) and basaltic lava flows (Iddings, 1890; Diller, 1891; Hiess et al., 2007) but are most commonly described in mafic enclaves hosted in both volcanic (Fenner, 1926; Bacon & Metz, 1984; Bacon 1986; Stimac & Pearce, 1992; Coombs et al., 2000; Murphy et al., 2000; Leonard, 2002; Browne et al., 2006a; Martin et al., 2006; Humphreys et al., 2009; Ruprecht et al., 2012; Borisova et al., 2014) and plutonic (Reid et al., 1983; D'Lemos, 1986, 1996; Frost & Mahood, 1987; Larsen & Smith, 1990; Pin et al., 1990; Vernon, 1990; Blundy & Sparks, 1992; Wiebe, 1994; Bateman, 1995; Wiebe et al., 1997; Akal & Helvaci, 1999; Silva et al., 2000; Baxter & Feeley, 2002; Kim et al., 2002; Choe & Jwa, 2004; Janoušek et al., 2004; Wada et al., 2004; Kumar & Rino, 2006; Şahin, 2008; Xiong et al., 2012; Kim et al., 2014) rocks. Textures within these minerals, such as reaction rims on olivine xenocrysts in andesites, can be used to estimate magma ascent timescales (Reagan et al., 1987; Matthews et al., 1992, 1994; Dirksen et al., 2014; Zhang et al., 2015). Transfer of different minerals can also influence the mixing signature on Harker diagrams (Ubide et al., 2014). Crystal transfer is likely to be accompanied by entrainment of its original melt (Cantagrel et al., 1984; Gourgaud & Maury, 1984; Coombs et al., 2000; Wright et al., 2011; Perugini & Poli, 2012; Ubide et al., 2014). However, direct observation of entrained melt is rare in natural volcanic examples (Wright et al., 2011) and is not evident in plutons where melts hybridise and crystallise. One example (Fig. 3b) shows an olivine xenocryst in an andesitic lava flow (White Island, New Zealand), where the crystal is surrounded by a film of basaltic glass (light grey), that is clearly distinct from the bulk of the lava (dark grey) and is the original melt from which the olivine crystallised. Such entrainment provides a mechanism by which the crystal's original magma can 'dilute' the intrusion (Perugini & Poli, 2012; Ubide et al., 2014) and enhance mingling. However, outstanding questions concerning the role of crystal shape on entrainment remain.

In addition to xenocryst capture, evidence for crystal transfer from the enclave back to the host is provided by disequilibrium phenocryst textures indicative of interaction with a more mafic magma (Cantagrel et al., 1984; Stimac & Pearce, 1992; Clynne, 1999; Nakada & Motomura, 1999; Tepley et al., 1999; Coombs et al., 2000; Troll & Schmincke, 2002; Ruprecht & Wörner, 2007; Humphreys et al., 2009; Ruprecht et al., 2012). This can occur through disaggregation of xenocrystic enclaves which disperse their load into the host (Tepley et al., 1999; Humphreys et al., 2009; Fomin & Plechov, 2012).

2.2 Analogue experiments

Early analogue experiments used non-magmatic fluids and particles to model magma mingling by injecting one viscous fluid into another (Huppert et al., 1984, 1986; Campbell & Turner, 1986). These studies considered magmas as pure melts and demonstrated that large viscosity contrasts prohibit efficient mingling. Field observations that some mafic magmas became vesiculated in response to undercooling by the host magma (Eichelberger, 1980; Bacon & Metz, 1984; Bacon, 1986) motivated experiments focussed on bubble transfer from one viscous layer into another, and demonstrated that the rise of bubble plumes could cause mingling (Thomas et al., 1993; Phillips & Woods, 2001; 2002). Recent experiments have examined the effect of crystals on intrusion break-up. For example, Hodge et al. (2012) injected a particle-rich corn syrup (high density and viscosity) into a large, horizontally sheared body of particle-free corn syrup (low density and viscosity) to model the injection of cooling (partially crystallised) mafic magma into a convecting magma chamber. They found that low particle concentrations caused the injection to fragment and form 'enclaves', whereas at high particle concentrations it remained intact and formed a coherent layer. Although no analogue experiments have considered liquid injection into variably crystalline suspensions, experiments with gas injection into particle-liquid suspensions show a strong control of particle concentration and injection style, with a threshold between ductile and brittle behaviour at random close packing (Oppenheimer et al., 2015; Spina et al., 2016).

2.3. High-temperature and/or high-pressure experiments

Investigations of magma interactions in high-temperature and/or high-pressure experiments can be broadly divided into two categories. Static experiments consider the juxtaposition of heated magmas and study mixing resulting from the diffusion of different melt components (Watson & Jurewicz, 1984; Carroll & Wyllie, 1989; Wylie et al., 1989; Van der Laan & Wyllie, 1993). Fluid motion can still occur in these static experiments, as variable diffusion rates between elements can create density gradients that drive compositional convection (Bindeman & Davis, 1999). Additionally, since water diffuses much more rapidly than other components (Ni & Zhang, 2008), transfer of water from hydrous mafic magmas to silicic bodies lowers the liquidus temperature of the latter, leading to undercooling and the production of quenched margins in the mafic member, even without a temperature contrast (Pistone et al., 2016a). Bubbles that exsolve in a lower, mafic layer can also rise buoyantly into the upper layer, entraining a filament of mafic melt behind them (Wiesmaier et al., 2015). Such bubble-induced mingling can be highly efficient and has also been documented in natural samples (Wiesmaier et al., 2015). It has been proposed that a similar style of mingling can occur through crystal settling (Renggli et al., 2016; Jarvis et al., 2019).

Dynamic experiments apply shear across the interface between two magmas and reproduce mingling behaviour. The shear can be applied in various ways, with a rotating parallel plate geometry (Kouchi & Sungawa, 1982; 1985, Laumonier et al., 2014; 2015), a Taylor-Couette configuration (De Campos et al., 2004, 2008; Zimanowski et al., 2004; Perugini et al., 2008), a Journal Bearing System (De Campos et al., 2011; Perugini et al., 2012) or by using a centrifuge (Perugini et al., 2015). These experiments have produced a variety of textures from homogenous mixed zones to banding. When pure melts are used, the combination of diffusional fractionation and chaotic advection can produce phenomena such as double-diffusive convection (De Campos et al., 2008) and reproduce non-linear mixing trends for various major and trace elements (Perugini et al., 2008; De Campos et al., 2011). Experimental results also suggest new quantities to describe the completeness of mixing, such as the concentration variance (Perugini et al., 2012) and the Shannon entropy (Perugini et al., 2015). Where crystals are considered, the presence of phenocrysts can enhance mingling by creating local velocity gradients and disturbing the melt interface (Kouchi & Sunagawa, 1982; 1985; De Campos et al., 2004). In contrast, other studies (Laumonier et al., 2014: 2015) have shown that the presence of a crystal framework in the mafic member prevents mingling, whilst the presence of water can enhance mingling by lowering the liquidus temperature, and thus the crystallinity, of the magma (Laumonier et al., 2015).

2.4 Numerical models

Sparks and Marshall (1986) developed the first simple model to describe viscosity changes caused by thermal equilibration of a hot mafic magma and a cooler silicic magma, and the resulting (limited) time-window in which mingling/mixing can occur. More sophisticated models have simulated mingling between melts driven by double-diffusive convection (Oldenburg, 1989), compositional melting (Kerr, 1994; Cardoso & Woods, 1996) and the Rayleigh-Taylor instability (Semenov & Polyansky, 2017). Another group of studies has used single-phase models to simulate elemental diffusion and advection in a chaotic flow field (Perugini & Poli, 2004; Petrelli et al., 2006). These models reproduce naturally-observed geochemical mixing relationships, including linear-mixing trends between elements with similar diffusion coefficients and large degrees of scatter when diffusion coefficients differ (Perugini & Poli, 2004; Nakamura & Kushiro, 1998). Interestingly, the simulations produce both regular and chaotic regions, which are unmixed and well-mixed, respectively, and have been interpreted to correspond to enclaves and host rock (Petrelli et al., 2006). This framework has been extended to account for a solid crystal-phase (Petrelli et al., 2016) by including a Hershel-Buckley shape-dependent rheology (Mader et al., 2013) and a parameterisation of the relationship between temperature and crystallinity (Nandedekar et al., 2014). This body of work has demonstrated that chaotic advection can speed-up homogenisation.

Models of mixing and mingling that consider two-phase magmas containing either solid crystals or exsolved volatiles often assume coupling between the phases. In this way, the solid or volatile phase can be represented as a continuous scalar field and the resultant effect on rheology is accounted for through a constitutive relationship. For example, Thomas and Tait (1997) used such a framework to show that volatile exsolution in an underplating mafic magma could create a foam at the interface with an overlying silicic magma. Depending on the exsolved gas volume fraction and melt viscosity ratio, mixing and mingling could then proceed through foam destabilisation, enclave formation, or a total overturn of the system. Folch and Martí (1998) showed analytically that such exsolution could lead to overpressures capable of causing volcanic eruptions. Recent finite-element models show that injection of a volatile-rich mafic magma into a silicic host can cause intimate mingling when viscosities and viscosity contrasts are low (Montangna et al., 2015; Morgavi et al., 2019). The combination of reduced density in the chamber and the compressibility of volatiles can (non-intuitively) lead to depressurisation in the chamber (Papale et al., 2017), which is important for interpretation of ground deformation signals (McCormick Kilbride et al., 2016).

The effect of crystals on mixing and mingling has also been modelled by treating the crystals as a continuous scalar field. Examples include simulations of mixing across a vertical interface between a crystal suspension (30% volume fraction) and a lighter, crystal-free magma (Bergantz, 2000), and injection of a mafic magma into a silicic host with associated melting and crystallisation (Schubert et al., 2013). The role of crystal frameworks in both the intruding and host magma is addressed by Andrews and Manga (2014), who model the role of thermal convection in the host, and associated shear stress on the intruding dyke. If convection occurs whilst the dyke is still ductile, then mingling will produce banding. Otherwise, the dyke will fracture to form enclaves. Woods and Stock (2019) have also coupled thermodynamic and fluid modelling to simulate injection, melting and crystallisation in a sill-like geometry.

Finally, isothermal computational fluid dynamic simulations have been used to examine the case of aphyric magma injecting into a basaltic mush. For sufficiently slow injection rates, the new melt percolates through the porous mush framework, whereas for faster injections, fault-like surfaces delimit a “mixing bowl” within which the crystals fluidise and energetic mixing takes place (Bergantz et al., 2015; 2017; McIntire et al., 2019; Schleicher & Bergantz, 2017; Schleicher et al., 2016). By explicitly modelling the particles with a Lagrangian scheme it is possible to account for particle-scale effects, including lubrication forces (Carrara et al., 2019), that are neglected when using constitutive relations from suspension rheology. These simulations suggest that mushes with $\leq$ 60% crystals can be mobilised by injection, but neglect welded crystals or recrystallisation of crystal contacts. Furthermore, geophysical observations suggest that mushes spend the majority of their lifetimes with much higher crystallinities (80-90%; Sinton and Detrick, 1992; Farrell et al., 2014; Pritchard et al., 2018).

3) Petrologic constraints on mingling conditions: Petrographic interpretations

Here, through the use of examples, we show how the texture and chemistry of enclaves and xenocrysts have been interrogated to interpret information on mixing and mingling processes. Although many studies have examined mixed and mingled rocks from both plutonic and volcanic realms, here we review work on examples from four volcanoes (Chaos Crags and Lassen Peak, both USA; Mt. Unzen, Japan; and Soufrière Hills, Montserrat) which have erupted intermediate composition lavas containing mafic enclaves (Fig. 4). We use common features, as recorded in the literature and augmented by an additional sample of Mt. Unzen lava from the 1792 dome collapse deposit, to develop a conceptual model that describes how volatile and crystals contents control mixing and mingling in magma storage regions. We analyse the latter using back-scatter electron images (BSE) obtained using both a Hitachi S-3500N (University of Bristol) and a Tescan Mira II (University of Lausanne) scanning electron microscope (SEM). Plagioclase compositions were measured on a Cameca SX100 (University of Bristol) with an accelerating voltage of 20 kV, emission current of 10 nA and a spot size of 3 μm.

3.1. Volcanic systems

3.1.1 Chaos Crags

Chaos Crags comprises of a series of enclave-bearing rhyodacite lava domes that erupted between 1125 and 1060 years ago (Clynne, 1990). The host lavas are crystal-rich, containing phenocrysts of plagioclase, hornblende, biotite and quartz, whilst the enclaves are basaltic andesite to andesite with occasional olivine, clinopyroxene and plagioclase phenocrysts in a groundmass of amphibole and plagioclase microphenocrysts (Heiken & Eichelberger, 1980). Many, but not all, enclaves have fine-grained and crenulated margins and all contain resorbed phenocrysts captured from the host (Fig. 4a). Some phenocrysts in the host also show resorption textures (Tepley et al., 1999).

3.1.2 Lassen Peak

Lassen Peak erupted in 1915, producing a dacite dome and lava flow followed by a sub-Plinian eruption that deposited two types of pumice: homogeneous dacite and banded dacite/andesite. The dome and flow are porphyritic with phenocrysts of plagioclase, biotite, hornblende and quartz in a glassy, vesicular groundmass containing microphenocrysts of plagioclase, pyroxenes and Fe-Ti oxides. The dacite dome and lava flow also contain xenocryst-bearing andesitic enclaves with equigranular texture and a lack of crenulated margins (Fig.

4b; Clynne, 1999). The enclaves have olivine phenocrysts (which occasionally appear as xenocrysts in the host) with plagioclase and pyroxene microphenocrysts.

3.1.3 Mt. Unzen

Mt. Unzen has erupted lavas and domes since 300-200 ka (Hoshizumi et al., 1999), most recently during the 1991-1995 eruption. With the exception of an andesitic lava flow from 1663, Mt. Unzen lavas are consistently dacitic, containing basaltic to andesitic enclaves (Hoshizumi et al., 1999; Browne et al., 2006a). Dacite erupted in the 1991-1995 eruption is porphyritic with about 20% phenocrysts of plagioclase, hornblende, biotite and quartz, with plagioclase, pargasite, pyroxenes, apatite and Fe-Ti oxides occurring as microlites in a highly crystalline groundmass (Nakada & Fuji, 1993; Nakada et al., 1999; Nakada & Motomura, 1999; Venezky & Rutherford, 1999; Cordonnier et al., 2009; Hornby et al., 2015). Two types of enclaves are observed: porphyritic and equigranular. Porphyritic enclaves contain large crystals of plagioclase, hornblende and rare quartz within a finer groundmass of plagioclase and hornblende microphenocrysts, minor amounts of clinopyroxene and olivine, and interstitial glass (Fig. 4c). The overall crystallinity is 70-90%. Equigranular enclaves contain equant microphenocrysts of plagioclase with smaller quantities of hornblende and orthopyroxene (Browne et al., 2006a).

3.1.4 Soufrière Hills

The 1995-2010 Soufrière Hills eruption produced a series of andesitic lava domes containing enclaves of basaltic to basaltic-andesitic composition (Wadge et al., 2014; Plail et al., 2014). The andesite contains approximately 40% phenocrysts (plagioclase, hornblende, orthopyroxene, Fe-Ti oxides and minor quartz) in a much finer-grained groundmass with up to 25% glass (Murphy et al. 2000; Humphreys et al., 2009; Edmonds et al., 2016). The enclaves have a diktytaxitic groundmass of plagioclase, pyroxenes, amphibole and Fe-Ti oxides with larger xenocrysts inherited from the andesite (Fig. 4d). Some enclaves have crenulated and fine-grained margins, whereas others are more equigranular and of a less mafic composition (Murphy et al., 2000; Plail et al., 2014; 2018).

3.2 Phenocryst, xenocryst and groundmass textures and chemistries

3.2.1 Enclave groundmass textures

The enclaves from all four volcanoes show both similar and contrasting textural features. At Chaos Crags, most enclaves have fine-grained and crenulate margins (Fig. 4a; Tepley et al., 1999), although those erupted in later domes are more angular and lack fine-grained margins. Enclaves in Lassen Peak samples are subrounded to subangular with an equigranular texture (Fig. 4b; Clynne, 1999). Many enclaves from the 1991-1995 eruption at Mt. Unzen have crenulate and fine-grained margins (Browne et al., 2006), although some have angular edges and a uniform crystal size (Fig. 4c; Fomin & Plechov, 2012). Similar features are observed at Soufrière Hills, with many inclusions being ellipsoidal (Fig. 4d) and some angular; most, but not all, have fine-grained, crenulate margins (Murphy et al., 2000). Both the size and volume fraction of enclaves increased during the eruption (Barclay et al., 2010; Plail et al., 2014; 2018).

In all localities, fine-grained margins and crenulate contacts are attributed to undercooling of the mafic magma due to juxtaposition against the much cooler felsic host (Eichelberger, 1980) and associated rapid crystallisation of the mafic melt near the contact with the felsic host. These crystalline rims have a greater rigidity than the lower-crystallinity enclave interiors so that as the enclave continues to cool and contract, the rims deform to a crenulate shape that preserves the original surface area (Blundy & Sparks, 1992). Enclaves not exhibiting such quench textures are also found at all localities.

3.2.2 Plagioclase

The composition and texture of plagioclase crystals are extremely good recorders of magmatic processes because 1) their stability field in pressure-temperature-composition (P-T-X) space is very large in volcanic systems, and 2) compositional zoning modulated by changes in the P-T-X space is well preserved due to the relatively slow diffusion in the coupled substitution between Na-Si and Ca-Al (Grove et al., 1984; Morse, 1984; Berlo et al., 2007).

Texturally, plagioclase phenocrysts in the host lavas at all four localities comprise a population of unreacted, oscillatory zoned crystals with a lesser amount of reacted crystals that have sieved cores and/or resorption rims (Fig 5a; Tepley et al., 1999; Clynne 1999; Murphy et al., 2000; Browne et al., 2006b). Associated enclaves contain plagioclase xenocrysts incorporated from the host with sieved-texture resorption zones that consist of patchy anorthite-rich plagioclase and inclusions of glass (quenched melt). These reacted zones can penetrate to the cores of smaller crystals (Fig. 5b,c), but in larger xenocrysts appear as a resorption mantle surrounding a preserved oscillatory-zoned core (Fig. 5d). All xenocrysts are surrounded by a clean rim that is of the same composition as the plagioclase microphenocrysts in the enclave groundmass.

The relationship between anorthite (An) and FeO content of plagioclase crystals can also provide insight into magma mixing and mingling. Plagioclase crystals erupted from Soufrière Hills volcano between 2001 and 2007 show a shallow, linear trend between FeO and An contents in oscillatory-zoned regions of plagioclase phenocrysts in the host (Humphreys et al., 2009); sieved zones in the same phenocrysts form a curved trend at higher FeO (Fig. 6d). In enclave-hosted xenocrysts, oscillatory-zoned cores plot on the same linear trend as oscillatory-zoned phenocrysts, whereas the clean rims overlap with the curved trend of the phenocryst sieved zones (Fig. 6f). The same curved trend is found for enclave microphenocrysts (Fig. 6e; Humpreys et al., 2009). We observe similar characteristics in our sample of Mt. Unzen dome lava (Figs. 6a-c).

3.2.3 Quartz

Quartz crystals in mingled lavas can also show distinctive features. Host phenocrysts are rounded and embayed (Fig. 7a; Clynne, 1999; Tepley et al., 1999; Murphy et al., 2000; Browne et al., 2006a; Christopher et al., 2014) and can also be fractured (Clynne, 1999). In the enclaves, quartz xenocrysts are surrounded by reaction rims of clinopyroxene and hornblende microphenocrysts and glass (Fig 7b; Clynne, 1999; Tepley et al., 1999; Murphy et al., 2000; Browne et al., 2006).

3.3 Interpretation of textures and chemistries

The common textural and chemical features of these volcanic systems suggest commonalities in the mixing and mingling processes. First, since enclaves from all volcanoes contain xenocrysts that originated in the host magmas, the mafic component must have been sufficiently ductile to incorporate these crystals during mixing. Plagioclase xenocrysts contain rounded, patchy zones with a sieved texture showing that both partial and simple dissolution occurred (Tsuchiyama, 1985; Nakamura & Shimakita, 1998; Cashman & Blundy, 2013), suggesting that the enclave magmas were undersaturated in plagioclase at the time of incorporation. Since up to 70% of the enclave groundmass consists of plagioclase microphenocrysts, this implies the mafic magmas were crystal-poor at the time of xenocryst incorporation.

Compositional variations of FeO and An in the plagioclase crystals provide further information on the relative compositions of the host and enclave melt at Soufrière Hills (Humphreys et al., 2009) and Mt. Unzen (Fig. 6). Most analyses from host phenocrysts show a shallow, increasing linear trend between An and FeO content (Fig. 6a,d); the few points with FeO enrichment correspond to resorbed zones. Unresorbed cores of xenocrysts have similar compositions, suggesting that both crystal core populations derive from the same host dacite magma. Enclave microphenocrysts, however, show greater FeO enrichment (Figs. 6b,e) and overlap with xenocrysts rim compositions. Similar results are reported for plagioclase in andesite lavas erupted from El Misti, Peru, which underwent resorption in response to mafic recharge (Ruprecht & Wörner, 2007). At Mt. Unzen, enclave microphenocrysts and xenocrysts rims show a strong positive correlation for the whole An range, whilst these phases at Soufrière Hills show a negative correlation for An > 75 mol% (Fig. 6). This difference is interpreted to reflect the absence of Fe-Ti oxide as an early crystallising phase in the Soufrière Hills mafic end-member, which would cause FeO to increase in the residual melt as other phases precipitated until the point of oxide saturation (Humprheys et al., 2009). The lack of this inflection in the Mt. Unzen sample suggests that Fe-Ti oxides were present in the mafic magma prior to mixing, as suggested for the 1991-1995 eruption (Holtz et al., 2005; Botcharnikov et al., 2008).

Whereas the observed enrichment in FeO in enclave microphenocrysts, sieved zones in phenocrysts and xenocrysts, and xenocrysts rims is likely due to crystallisation from a more mafic melt, it is also possible that growth of these regions may be sufficiently fast for kinetic effects to play a role; if growth is faster than diffusion of FeO in the melt then an FeO-rich boundary layer may develop around crystals (Bottinga et al., 1966; Bacon, 1989; Mollo et al., 2011) that could also explain the enrichment. However, such a process would generate a negative correlation between FeO and An (Neill et al., 2015), not the positive correlation observed at Unzen and Soufrière Hills.

The contrasting textures of quartz in the host and enclaves also provide insight into the mingling/mixing process. Rounding of quartz xenocrysts, together with glass-filled embayments, suggests dissolution of quartz in the host. Conversely, quartz reaction rims comprising hornblende microphenocrysts, glass and vesicles in the enclaves (Figures 3d, 7b) suggest that the dissolution-induced increase in the silica content (and H_2O solubility) of the surrounding melt caused diffusion of H_2O towards the quartz (Pistone et al., 2016a).

Whereas the presence of resorbed xenocrysts in enclaves suggests that there was time for crystals to be incorporated, and to react, before the enclave started to crystallise, the presence of fine-grained rims on some enclaves (Tepley et al., 1999; Murphy et al., 2000; Browne et al, 2006a; Barclay et al., 2010; Plail et al., 2014) implies rapid cooling and crystallisation (chilling) of the mafic magma against the cooler silicic host (Bacon, 1986). Xenocrysts must therefore have been incorporated prior to the formation of the chilled margin, providing a limited temporal window for crystal transfer. A comparison of the thickness of xenocryst re-sorption zones at Mt. Unzen (Brown et al. 2006a) with those produced experimentally (Tsuchiyama & Takahasi, 1983; Tshuchiyama, 1985; Nakamura & Shimakita, 1998) suggests resorption on a timescale of days; this contrasts with thermal modelling (Carslaw & Jaeger, 1959) suggesting enclaves should thermally equilibrate on a timescale of hours. Again, this requires incorporation of xenocrysts prior to intrusion disaggregation and enclave formation (Browne et al., 2006a). As the considered volcanic lavas contain similarly resorbed plagio-clase xenocrysts within enclaves of comparable sizes, it seems likely that this temporal con-straint on the sequence of crystal transfer prior to enclave formation is generally true for the systems presented here.

Importantly, all locations also contain enclaves with unquenched margins (Tepley et al., 1999; Plail et al., 2014) and equigranular textures (Heiken & Eichelberger, 1980; Browne et al., 2006a). Equigranular enclaves at Mt. Unzen have been interpreted as originating from disaggregation of the interior of the intruding magma, which cooled more slowly than the in-trusion margin where porphyritic enclaves (xenocrysts-bearing, chilled margin) formed. Sim-ilarly, at Soufrière Hills, the quenched enclaves may form from an injected plume of mafic magma, whereas unquenched and more hybridised enclaves form from disturbance of a hy-brid layer at the felsic-mafic interface (Plail et al., 2014). Angular enclaves with unquenched margins may record the break-up of larger enclaves (Clynne, 1999; Murphy et al, 2000; Fomin & Plechov, 2012; Plail et al., 2014), which can return resorbed host-derived crystals to the host; this explains the presence of resorption zones in crystals in the host lavas (Fig. 5b), and chemical signatures (Fig. 6a) of crystallisation from mafic magma. Further support for enclave fragmentation comes from microlites that are chemically indistinguishable from en-clave phases at Soufrière Hills (Humphreys et al., 2009). A possible method to determine whether equigranular enclaves form from a hybrid layer or disaggregation of larger enclaves is to examine the mineralogy of the crystals in the enclave. The two different mechanisms will produce different degrees of undercooling within the enclave magma, which, in the hybrid-layer model, will depend on the relative proportions of the end-member magmas, and thus can produce different crystal assemblages/textures (Humphreys et al., 2006).

3.4. Conceptual model of magma mixing and mingling

Common features of eruptive products described above suggest common aspects of mixing and mingling. Xenocrystic mafic enclaves with chilled margins, in particular, require that magma injection is accompanied by crystal incorporation from the host magma, as also suggested by a comparison of thermal timescales with the times needed to generate the observed thicknesses of resorption zones (Browne et al., 2006a). These constraints on the sequence of mixing processes have led to a similar conceptual model of mixing and mingling (Fig. 8; Clynne, 1999; Tepley et al., 1999; Murphy et al., 2000; Browne et al., 2006; Plail et al., 2014) in which the mafic magma is injected as a fountain (Clynne, 1999) or collapsing plume (Plail et al., 2014) before ponding at the base of the silicic host (Fig. 8a). Shear caused by the injection disrupts the interface between the two magmas, leading to the formation of blobs of hybridised magma with incorporated host crystals that then rapidly chill against the silicic host, preventing further hybridisation (Tepley et al., 1999; Plail et al, 2014). Heating of the host, in turn, causes partial melting, reducing the crystallinity and causing convective motions that disperse the enclaves. Meanwhile, at the mafic-silicic contact, a hybrid interface layer forms (Fig. 8b). As this layer crystallises, second boiling drives fluid saturation; exsolved buoyant fluids produce a low-density, gravitationally-unstable, interface layer that breaks-up to form further enclaves (Fig. 8c; Clynne, 1999; Browne et al., 2006a). As cooling propagates downwards through the mafic body, enclaves can come from deeper portions resulting in more equigranular enclaves that lack chilled margins or xenocrysts (Brown et al., 2006a; Plail at al., 2014).

Enclaves, once formed, can disaggregate. Disaggregation is shown by the presence of broken enclaves (Clynne, 1999; Tepley et al., 1999; Fomin & Plechov, 2012), host phenocrysts with resorption zones and Fe enrichment caused by previous engulfment in mafic magma (Clynne, 1999; Tepley et al., 1999; Browne et al., 2006b; Humphreys et al., 2009), and small clusters of enclave-derived microlite material within the host lavas (Humphreys at al., 2009). Disaggregation allows for subsequent mixing of a type precluded during initial enclave formation, but the timing of disaggregation is poorly constrained. It could occur during high-shear conditions in the conduit (Humphreys et al., 2009); alternatively, disaggregation may be part of a continuous cycle of injection, enclave formation and fragmentation (Fig. 8d) that gives rise to a continuously convecting magma storage region, which is sometimes sampled during a volcanic eruption (Browne et al., 2006a). Regardless, the dispersion of mafic groundmass into the host has implications for interpreting end-member compositions from petrologic studies (Martel et al., 2006; Humphreys et al., 2009). Importantly, neglecting such transfer can lead to an under-estimate of the initial silica content of the felsic member.

4 Quantitative modelling of crystal and volatile controls on mixing and mingling

Many conceptual models of magma mixing (e.g. Fig. 8) have been produced based on petrologic evidence. However, quantitative models of magma mixing are limited. As described in Section 2.3, Sparks and Marshall (1986) first developed a simple model describing how thermal equilibration of a juxtaposed mafic and silicic magma led to rapid viscosity changes that inhibited mixing after a short time. Since then, models developed to account for the role of either crystals or exsolved volatiles have produced significant insights into mingling and mixing dynamics, but have failed to incorporate petrological data within quantitative frameworks. Here, we examine three models: Andrews and Manga (2014), who use continuum modelling and suspension rheology to model mingling resulting from dyke injection into a silicic host; Bergantz et al. (2015), who model the injection of melt into a basaltic mush, resolving both fluid and granular behaviour; and Montagna et al. (2015), who simulate the effect of exsolved volatiles on mafic injection. We compare the model assumptions and results, as well as their implications for interpreting petrological data.

4.1 The model of Andrews and Manga (2014)

The model considers the instantaneous injection of a mafic dyke into a silicic host, with a prescribed initial composition and temperature, and numerically solves the 1D heat equation. Changes in the crystallinity and bulk viscosity of magmas with time are calculated using MELTS simulations (Ghiorso & Sack, 1995; Asimow & Ghiorso, 1998) and viscosity models for melt (Giordano et al., 2008) and crystal-bearing suspensions (Einstein, 1906; Roscoe, 1952). If the viscosity of the host immediately juxtaposed with the dyke decreases sufficiently, then the host starts to convect (as determined by a Rayleigh number criterion), which exerts a shear stress on the dyke. If this shear stress exceeds the yield stress of the dyke (which depends on its crystal content), the dyke deforms in a ductile fashion and the model predicts banded products. Alternatively, if the yield stress exceeds the shear stress, then the dyke fractures in a brittle fashion and enclaves form.

In this model context the principal control on mingling dynamics is the development of crystal frameworks within the dyke. Dyke crystallisation, in turn, is controlled by composition and temperature contrasts. For example, injection of hot, large and wet dykes causes the silicic host to convect before a crystal framework forms in the dyke. The resultant shear causes ductile disruption of the dyke and intimate mingling of the two magmas, producing banding and, with time, homogenisation. Small and dry dykes, in contrast, experience extensive crystallisation before the host starts to convect and thus fracture to form enclaves. The precise initial conditions (temperature, dyke size and water content) that determine mingling style are sensitive to the parameterisations used (e.g. critical Rayleigh number for convection) but the qualitative results are useful.

The principal limitation of the model of Andrews and Manga (2014) is that it assumes an instantaneous injection of the mafic dyke and therefore neglects any mixing/mingling that occurs during injection itself. Instead, the dyke is disrupted only by shear due to convection in the host. Indeed, the relative importance of shear due to injection versus shear due to convection remains a considerable unknown. The assumption that brittle fragmentation of the dyke produces enclaves is supported by three-dimensional tomographic observations of enclaves from Chaos Crags, which have crystal frameworks that are lacking in banded pumices from Lassen Peak (Andrews & Manga, 2014). The inference is that these crystal frameworks created a yield stress such that the enclaves formed by solid-like fracturing and banded pumice by ductile deformation. However, this is in direct contradiction with the conceptual model presented above (Fig. 8), which is based on field and petrographic observations that suggest enclaves form from fluid-like deformation of the mafic magma. This contradiction highlights the extent to which conditions of enclave formation are unknown.

4.2 The model of Bergantz et al., (2015)

The discrete-element model, which resolves both fluid and granular physics, considers the injection of a crystal-free magma into the base of a crystal mush at random loose packing (approximately 60% crystallinity). The response of the mush is governed by stress chains formed by crystal-crystal contacts. For sufficiently slow injections, the new melt permeates through the mush, which behaves as a porous medium. Once the injection speed is large enough to disrupt the stress chains, however, part of the mush can become fluidised to form a mixing cavity, which is an isolated region where the host melt, crystals and new melt undergo over-turning. The new melt then escapes from the cavity through porous flow into the rest of the mush. For still faster flow speeds, the stress chains orientate to create two fault-like surfaces at angles of about 60° to the horizontal that bound a fluidised region of the mush, within which extensive circulation occurs.

Whilst this model captures granular and fluid dynamics on the crystal scale and demonstrates the impact of varying the injection velocity, there are numerous outstanding questions. Firstly, varying the crystallinity of the mush has not been addressed and will presumably affect the values of injection velocity at which transitions between mingling styles occur. Furthermore, temporal and spatial variations in temperature (due to heat transfer or latent heat release), and therefore in viscosity and crystallinity, have not been considered. Cooling and crystallisation of the new melt should control the dynamics of the system, as will associated latent heat release. Finally, the geometry of the modelled magma reservoir (laterally homogenous layers) will affect the specifics of the mixing process, such as the orientation of the bounding faults, and it is not yet clear if the model scales to natural systems.

4.3. The model of Montagna et al. (2015)

The two-dimensional finite-element model considers two vertically-separated magma chambers that are superliquidus and connected by a narrow conduit. The upper chamber initially contains a felsic phonolite, whilst the lower chamber and conduit are filled with a mafic shoshonite, compositions chosen to represent eruptions from Campi Flegrei. H_2O and CO_2 exsolve as functions of temperature and pressure (Papale et al., 2006), whilst the transport of exsolved volatiles is modelled as a continuum scalar field satisfying a transport equation. Bubbles are assumed to be sufficiently small that they are undeformable and an empirical law is used to parameterise their effect on bulk viscosity (Ishii & Zuber, 1979). The shoshonite initially contains exsolved volatiles and so is lighter than the phonolite, creating an unstable density interface at the inlet to the upper chamber.

Upon initiation, a Rayleigh-Taylor instability develops at the inlet to the upper chamber and a plume of light material rises into the chamber whilst the conduit is filled with a mixed, hybrid magma. Intimate mingling within the chamber is reminiscent of that created by chaotic advection (Perugini & Poli, 2004). The magma entering the upper chamber is a partial hybrid, and the pure parent shoshonite never enters the upper conduit. Intensive mingling occurs on a timescale of hours, promoted by a large initial density contrast and horizontally-elongated chambers. Importantly, the reduction in density of the upper chamber can cause depressurisation, which has implications for interpreting ground deformation signals (Papale et al., 2017).

Whilst an obvious limitation of the model is the two-dimensional domain, it seems reasonable that the results can be extrapolated to three-dimensional systems. A greater limitation is the restricted range of compositions and temperatures for which the model is valid. The end-member compositions are similar and superliquidus, so that both the absolute bulk viscosities (< 3500 Pa s) and their contrast (factor of 7) are relatively low. This allows rapid mingling and ignores entirely the effect of crystals on the flow dynamics.

4.4. Comparison and common limitations

Both Andrews and Manga (2014) and Bergantz et al. (2015) focused on the effect of crystals, but a key difference in the two models is the initial condition. Andrews and Manga (2014) assume the instantaneous injection of a dyke into an initial rheologically-locked host, whereas Bergantz et al. (2015) simulate the flow of new melt into a melt-crystal mixture; they show that new melt flows permeably through a rheologically-locked mush. The conditions that spatially constrain a mafic injection (e.g. as a dyke) have not been defined. The two models also simulate the role of crystals differently. Andrews and Manga (2014) calculate the crystallinity of a magma at a given temperature and assume the presence of a crystal framework (and yield stress) above a threshold value. Bergantz et al. (2015) allow the crystals to form force chains through which stresses are transmitted (Bergantz et al., 2017), but they consider the system to be isothermal such that no crystallisation occurs, a key feature of Andrews and Manga (2014).

Both models are limited in addressing the role of volatiles. Diffusion of volatiles from the mafic to felsic member can strongly influence the crystal composition and textures of the silicic member (Pistone et al., 2016a), while exsolution of volatiles leads to a reduction in bulk density that can drive convective motions in the mixing dynamics (Eichelberger, 1980; Thomas et al., 1993; Phillips & Woods, 2001; 2002; Montagna et al., 2015; Wiesmaier et al., 2015). The presence of exsolved volatiles also affects the magma rheology and requires the use of three-phase rheological models (Mader at al., 2013; Pistone et al., 2016b). One strategy is to treat the exsolved phase as a continuum scalar field and use a suspension model for bulk rheology (Montagna et al., 2015). However, as has been shown for solid phases (Carrara et al., 2019), small scale effects can be overlooked by this approach and explicit modelling of such phases may be needed to accurately constrain mixing/mingling processes.

Additional complications arise in the number of parameters required for a given model. For example, the Andrews and Manga (2014) model requires values for a maximum crystal packing fraction and a critical Rayleigh number for convection in the host. Constraining these parameters will require extensive experimental efforts involving both high-temperature/high-pressure and analogue experiments.

5. Conclusions and outlooks for future research

We have reviewed progress in understanding magma mixing and mingling, focusing on volatile and crystal controls on mingling processes. Whilst field and petrologic observations of mixed and mingled products are numerous, models of these processes do not yet include the full range of observed complexities. In particular, conceptual models derived from observations (Clynne, 1999; Tepley et al., 1999; Browne et al., 2006; Plail et al., 2014) suggest very different dynamics to those from numerical models (Andrews & Manga, 2014; Bergantz et al., 2015; Montagna et al., 2015). To resolve this discrepancy, several key questions need to be addressed:

1) How do mixing and mingling occur within the framework of crystal mushes, and how does the volume fraction of crystals control the interaction dynamics?

2) How do volatiles, both exsolved and dissolved, affect mixing and mingling? What is the relative importance of chemical quenching (due to volatile diffusion) vs. thermal quenching (due to heat diffusion)?

3) How much mingling/mixing takes place during intrusion of the mafic magma compared to that driven by later processes such as convection in the host or the buoyant rise of vesicular mafic/hybrid magma?

4) How does latent heat, released from crystallisation of the mafic component and absorbed by melting of the felsic component, affect the mixing and mingling process? Latent heat

release may have a strong local effect in a dynamic system but is not evaluated, for example, in isothermal experiments.

5) To what extent are mafic injections spatially limited, e.g. dykes, and under what conditions might they affect the entire intrusion?

6) If magma storage regions undergo repeated replenishments with occasional eruptions, what factors determine if a particular injection leads to an eruption?

Only by combining field and analytical observations with experimental (analogue and natural materials) and numerical modelling can we start to address these challenges.

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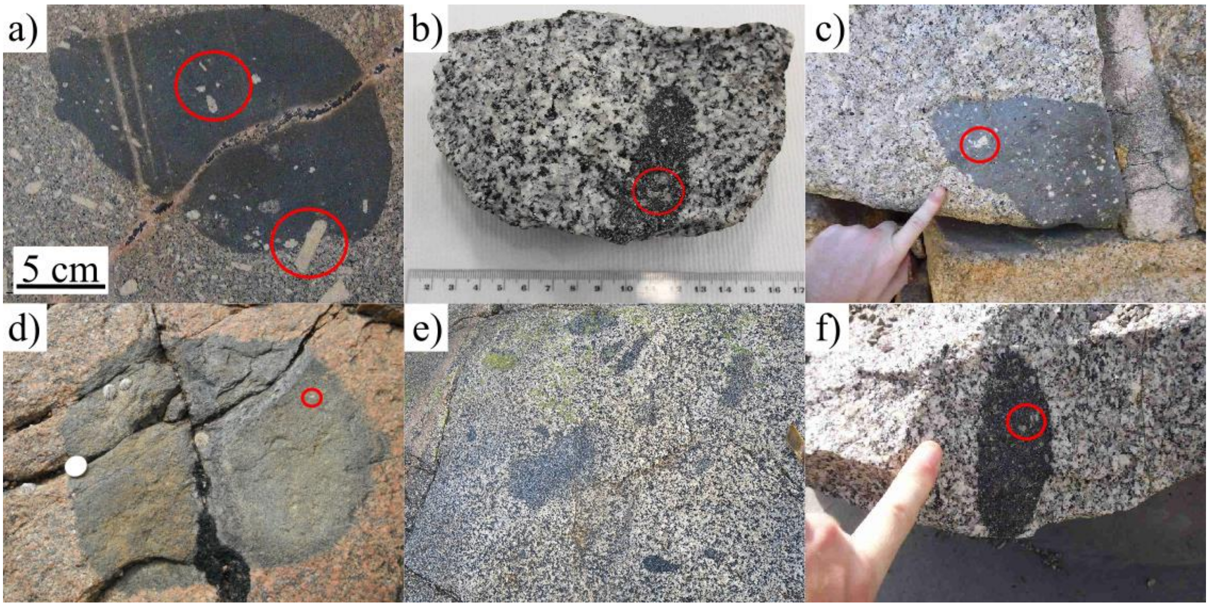

Figure 1: Examples of mafic enclaves. All are generally finer grained than their hosts but contain occasional large crystals (circled in red), which are xenocrysts mechanically transferred from the host. a) Large (≈ 18 cm) fine-grained enclave hosted in alkali-feldspar granite from Blackenstone Quarry, Dartmoor, England. b) High aspect ratio enclave from the Adamello Massif, Italy. c) Mafic enclave in granite of stone wall at Hiroshima Castle, Japan. d) Mafic enclave within the Cobo Granite, Guernsey. e) Numerous enclaves in an outcrop of the Northern Igneous Complex, Guernsey. The outcrop shown is about 1 m^2. f) Mafic enclave in a granite statue in Alexander Garden, Moscow.

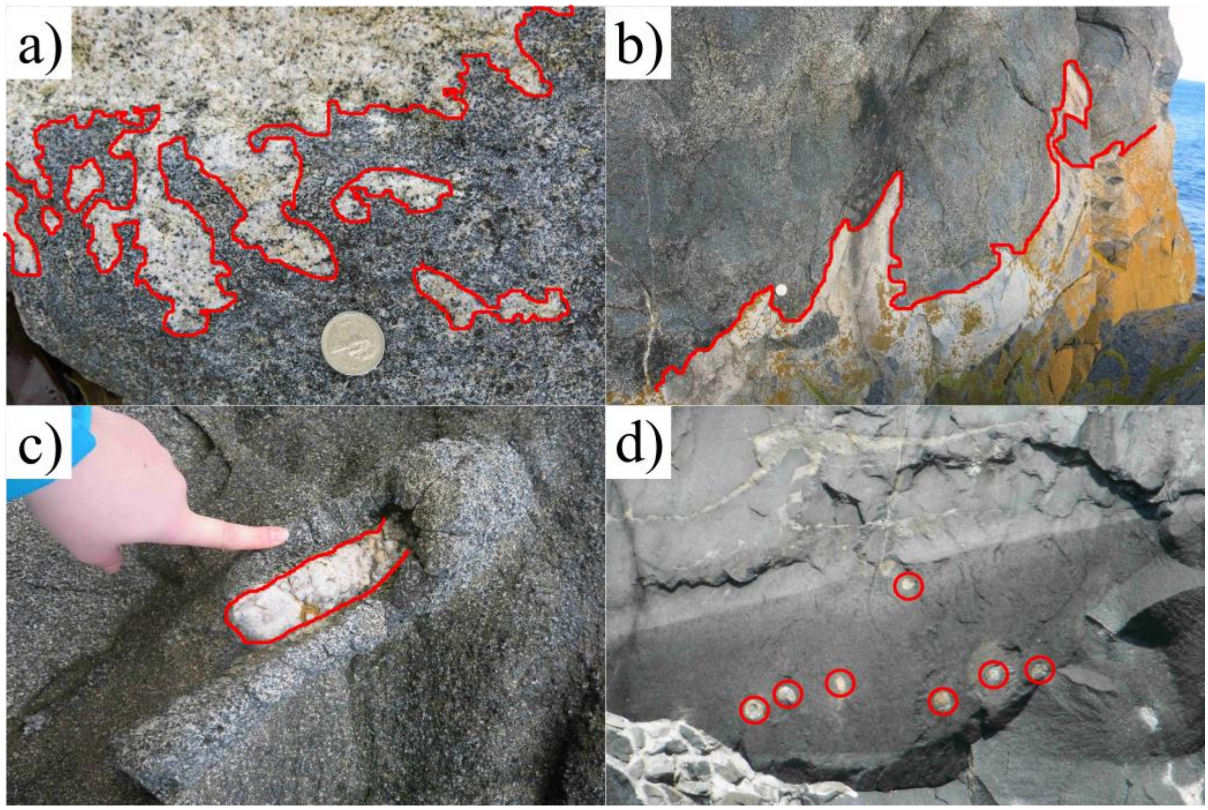

Figure 2: Examples of mingling textures from layered intrusions of the Northern Igneous Complex, Guernsey. Noted textures are outlined in red. a) Loose block showing intimate mingling between a felsic and a mafic magma. b) Diapir-like structures of felsic material rising into a mafic layer. c) Pipe of felsic material penetrating a mafic layer. d) Cross section through pipes similar to that seen in c).

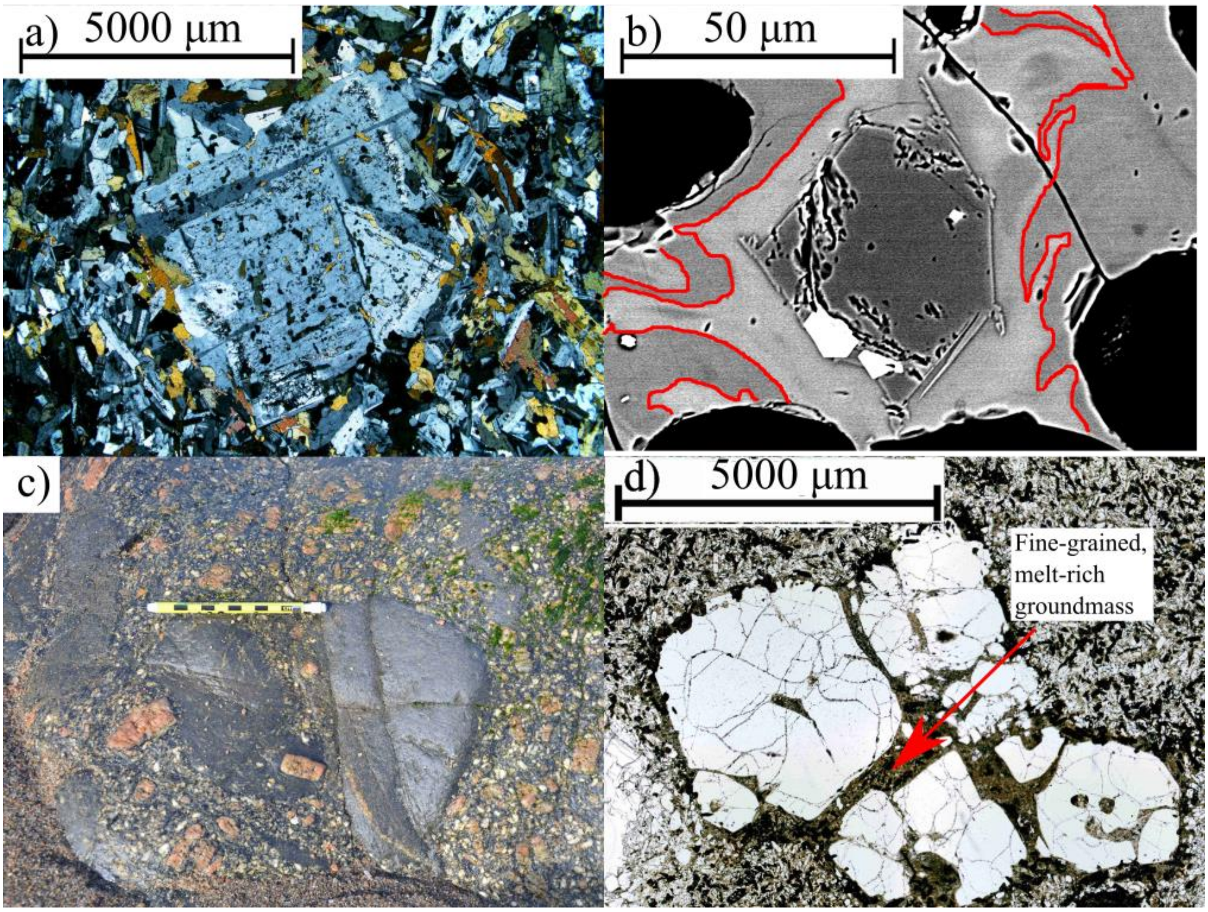

Figure 3: Examples of magmatic xenocrysts. a) Plagioclase xenocryst in a mafic enclave from the Adamello Massif, Italy, showing a sieved core with many inclusions of hornblende. b) Back-scatter electron (BSE) image of an olivine xenocryst in an andesitic lava flow from White Island volcano, New Zealand. The crystal is surrounded by an irregular film of basaltic glass (bounded by red contour). Image courtesy of Geoff Kilgour. c) Alkali feldspar xenocrysts, up to 3 cm, within mafic rocks on Shetland, Scotland. The relation of the mafic rocks to the felsic rocks from which the feldspars originated in unknown since the contact is in the subsurface. Image courtesy of Amy Gilmer. d) A cluster of rounded and highly fractured quartz xenocrysts in the Cardones ignimbrite, Chile (van Zalinge et al., 2016). The surrounding groundmass is much finer-grained and melt-rich than the rest of the material. The cluster has a rim of opaque crystals.

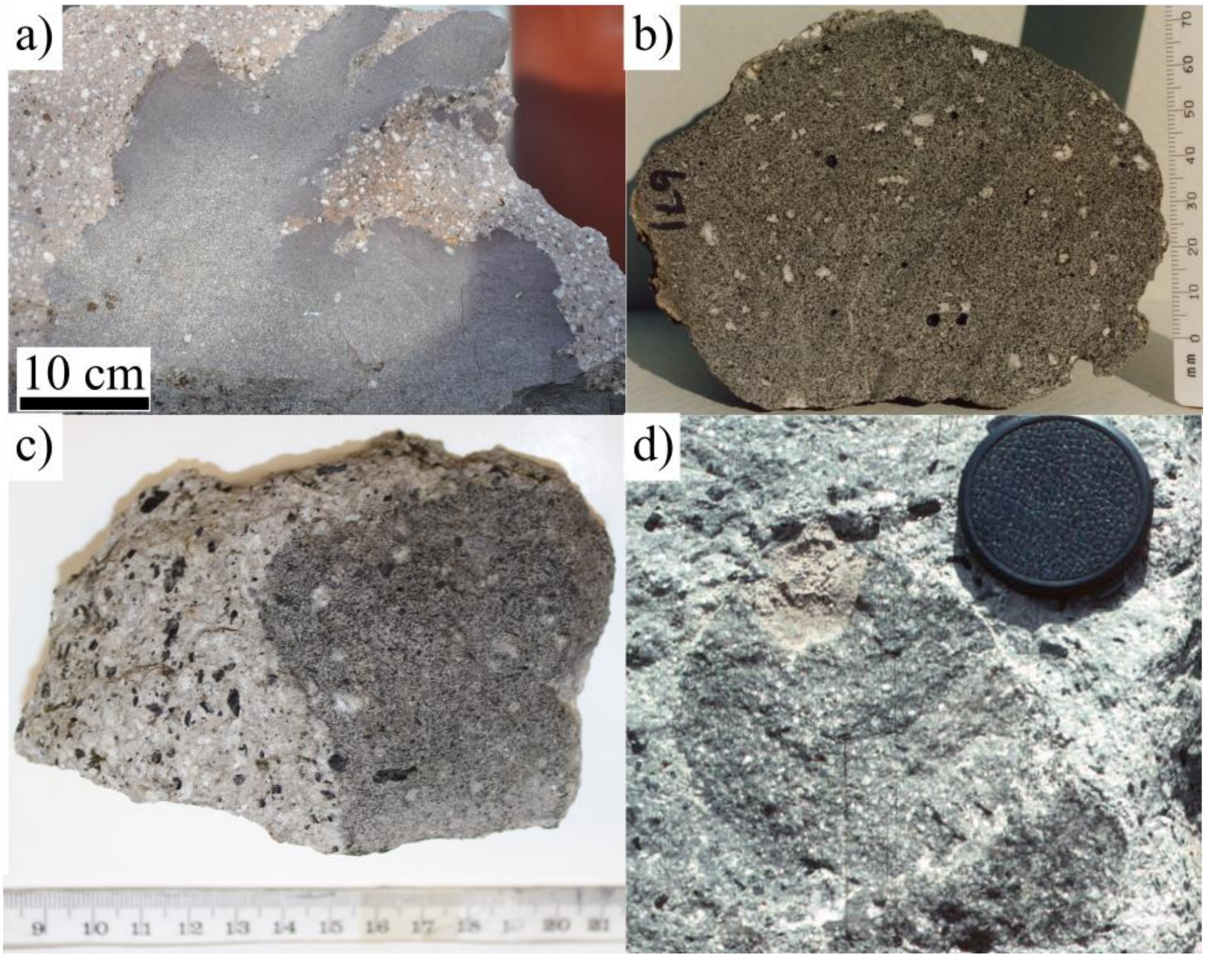

Figure 4: Examples of enclaves from four volcanic systems. a) Mafic enclave, with fine-grained, crenulate margin and numerous xenocrysts, from Chaos Crags. Image courtesy of Michael Clynne. b) Andesitic enclave from 1915 eruption of Lassen Peak, showing an equigranular texture and numerous partially reacted xenocrysts. Reproduced with permission from Clynne (1999). c) Basaltic enclave in an andesitic lava flow from the 1792 dome collapse at Mt. Unzen, Japan, at about 4 ka. There is no evidence for a fine-grained margin in the enclave. Sample collected by Julie Oppenheimer, Karen Strehlow and Emma Liu. d) Mafic enclave from 1995-2010 eruption of Soufrière Hills volcano. Image courtesy of Steve Sparks.

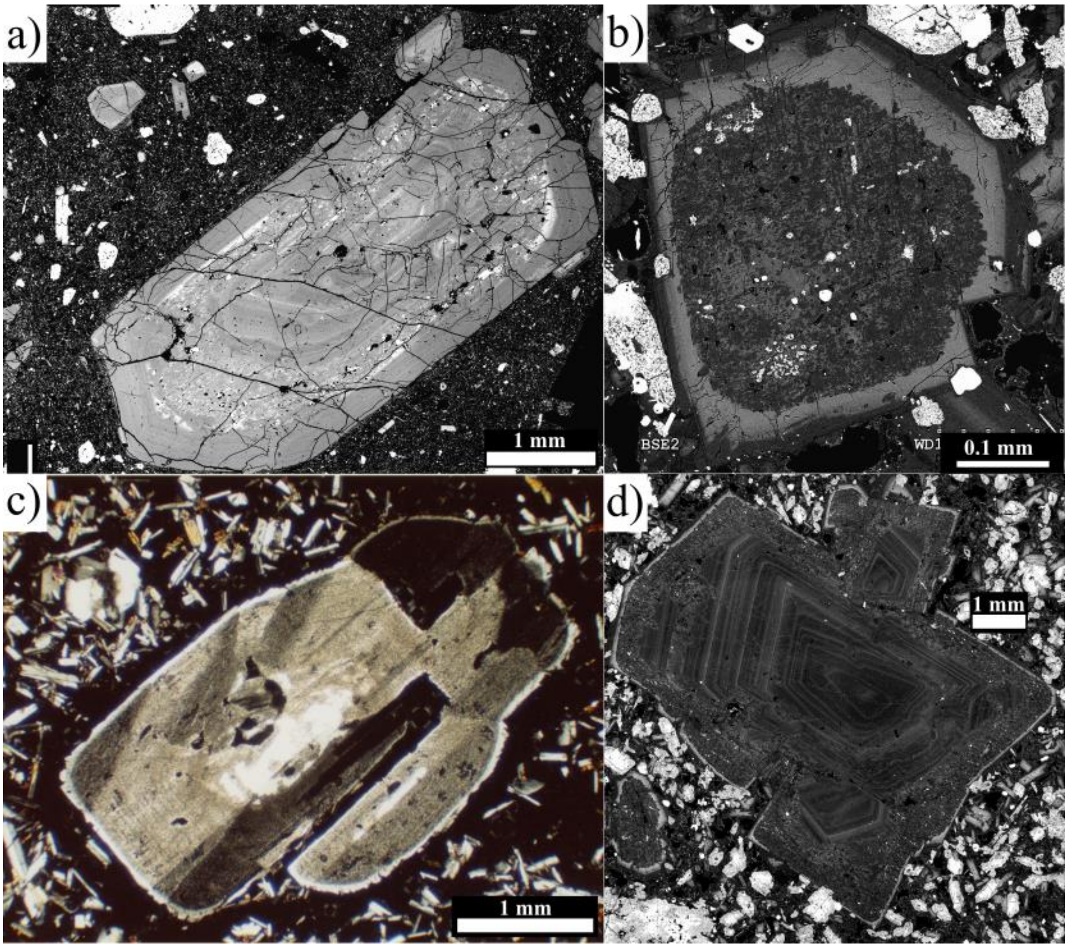

Figure 5: BSE images of plagioclase phenocrysts and xenocrysts from Mt. Unzen (a, b, d) and Lassen Peak (c). a) Host-rock plagioclase phenocryst with a wide heterogeneous zone and many mineral and glass inclusions. b) Plagioclase xenocryst in an enclave with a sieved core. c) Heavily reacted plagioclase xenocryst with a clear overgrowth rim within an andesitic enclave. Reproduced with permission from Clynne (1999). d) Plagioclase xenocryst in a mafic enclave with an oscillatory zoned core surrounded by a patchily zoned and inclusing-rich mantle bounded by a normally-zoned rim.

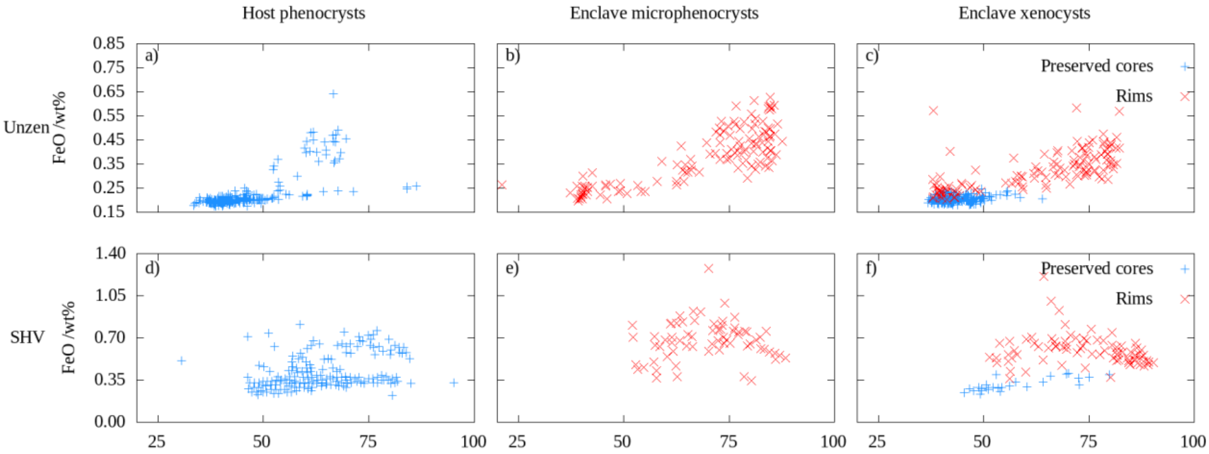

Figure 6: Plots of FeO versus An from transects across plagioclase for a,d) host phenocrysts, b,e) enclave microphenocrysts from and c,f) xenocrysts in enclaves. Data shown for Mt. Unzen sample in Fig. 4c and for rocks from Soufrière Hills (Humphreys et al., 2009). a,d) Most host phenocrysts lie on a shallow, linear trend with slight positive correlation in the range An = 33-88 mol% for Mt. Unzen and An = 45 – 80 mol% for Soufrière Hills. Some analyses show much greater FeO enrichment and correspond to resorbed zones. b,e) Enclave microphenocrysts show FeO enrichment compared to host phenocrysts, up to 0.65 wt% for Mt. Unzen and 1.3 wt% for Soufrière Hills. c,f) The preserved cores of xenocrysts plot on the same shallow trend as host phenocrysts whilst rim compositions overlap with enclave microphenocryst compositions.

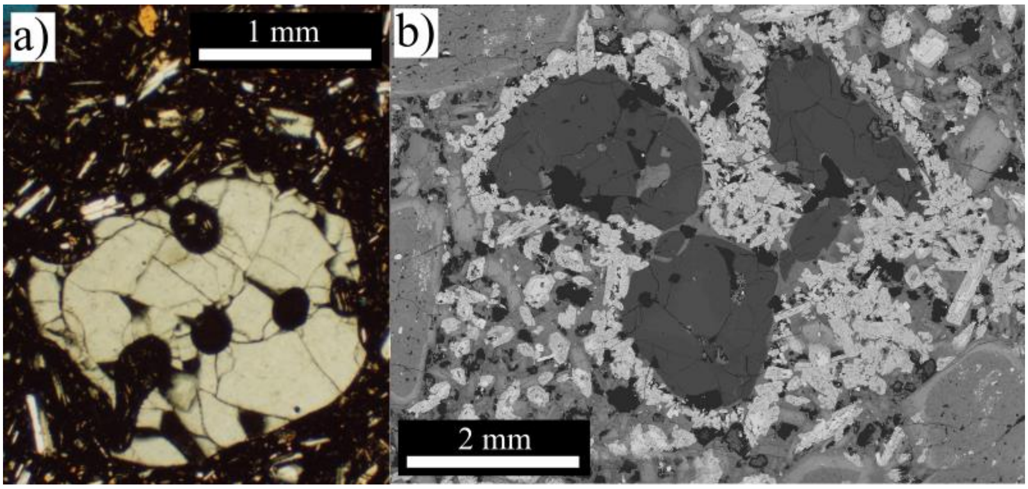

Figure 7: a) Image of a quartz phenocryst in the dacitic lava dome from the 1915 Lassen Peak eruption. It appears unreacted but has rounded edges and embayments. Image courtesy of Michael Clynne. b) BSE image of a cluster of quartz xenocrysts in an enclave from the 1991-1995 Mt. Unzen eruption. They are rounded and surrounded by an extended region of hornblende microphenocrysts (very bright), glass (light grey) and vesicle (black).

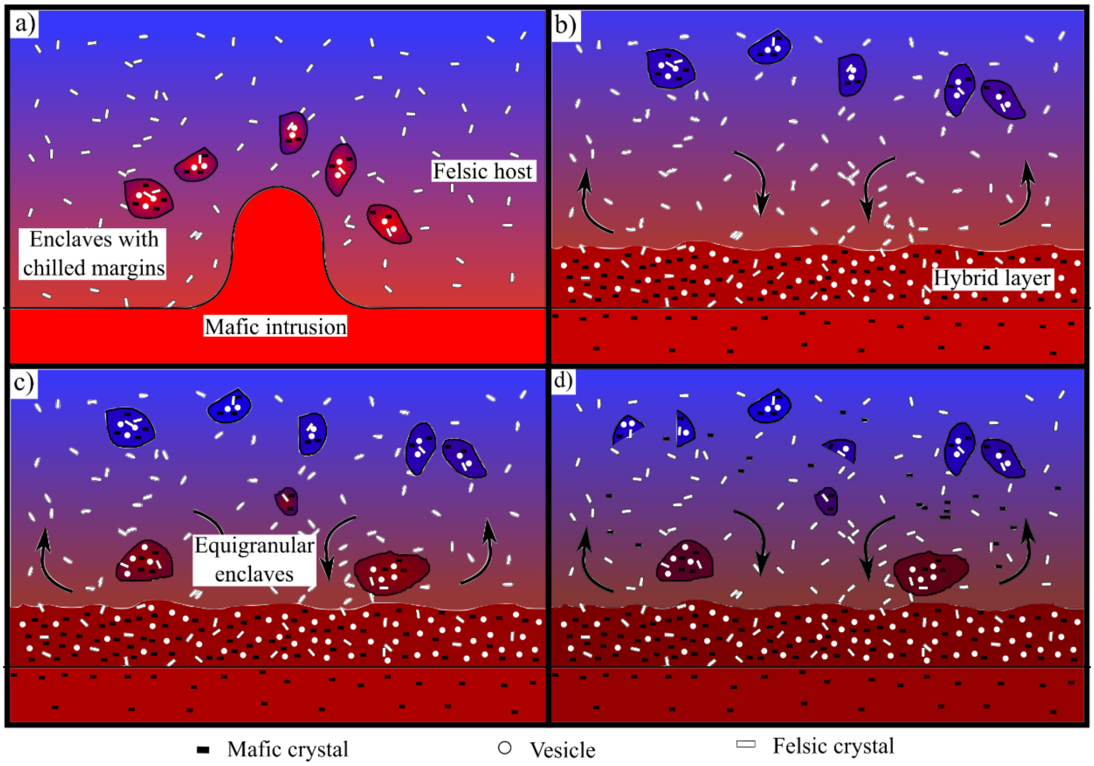

Figure 8: Conceptual model of the different stages of magma mixing and mingling following injection of a mafic magma into a partially crystallised silicic host. The processes shown follow similar diagrams from Clynne (1999), Tepley et al. (1999), Browne et al. (2006) and Plail et al. (2014). a) Mafic magma is injected into a partially crystallised host. Injected magma is initially denser and so ponds beneath the silicic host, although the momentum of the injection may produce a collapsing fountain. b) Disaggregation of the collapsing fountain produces quenched enclaves with chilled margins. These enclaves contain xenocrysts captured from the host, which became entrained during the injection. Heat transfer from the mafic to the silicic magma produces partial melting of the silicic member, reducing the crystallinity and creating convective motions that disperse the enclaves. Additionally, a hybrid layer forms at the interface between the mafic and silicic magmas. Crystallisation in this layer leads to exsolution of volatile phases. c) The presence of exsolved volatiles in the interface layer leads to a reduction in density and the hybrid layer destabilises due to a Rayleigh-Taylor instability. This leads to the formation of enclaves without chilled margins that are dispersed within the silicic host. d) Continued convective motions in the host lead to brittle disaggregation of enclaves creating angular enclave fragments, and dispersing mafic groundmass and resorbed host crystals into the host.