



Bulk chemical trends at arc volcanoes are not liquid lines of descent

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Abstract

Although the magmatic output of arc volcanoes defines clear chemical trends, the compositional sequence of eruption is often chaotic. At best, support for the concept of progressive evolution by fractionation/assimilation in a central chamber is ambiguous. Rather than defining a liquid line of descent, we suggest that chemical trends in arc systems are populated by discrete magma batches. These have been derived from a limited set of sources and undergone a common family of processes, but in different proportions. Crystal–liquid fractionation may occur at the mafic end of the spectrum in classical fashion by gravitational crystal extraction from melt-rich magma. However, at the silicic end where melt viscosity is high, fractionation may occur by forceful expulsion of interstitial melt from near-solidus arc plutons and adjacent crust. The andesite/dacite middle ground is commonly occupied by relatively crystal-rich magma batches with complex mixing and mingling histories. Such intermediate magmas appear to be stable residents of shallow arc crust—mechanically because of their neutral buoyancy, thermally because they are frequently recharged with basalt, and chemically because they resist large shifts in composition. Stored magma that “regresses” in an evolutionary sense by mixing or mingling during recharge is preferentially erupted and forward evolution during quiescence is inhibited by high viscosity of melt. The surface expression is the building of a stratocone or dome complex. In contrast, silicic magmas, once they have coalesced from interstitial melt in their plutonic sources through coupled porous flow and diking, tend to be erupted promptly, flushing out resident andesite/dacite magma batches in their path. The surface expression is caldera formation, which commonly terminates one episode of cone building and serves as a prelude to another.

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1. Introduction

Arc volcanic centers exhibit well-defined trends in composition space (e.g., Hildreth et al., 2004). Although basalt (e.g., Fuji, Japan; Shishaldin, Alaska), andesite–dacite (e.g., Redoubt, Alaska; Unzen, Japan), or even rhyolite (e.g., Taupo Volcanic Zone,

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New Zealand) may be dominant, the overall chemical spectrum extends from basalt to rhyolite.

What do these trends mean? The fact that bulk compositions are not random but that chemical components correlate with each other requires that some finite, common set of chemical and mechanical processes are at work to constrain magma chemistry. That the trends “aim” (in the sense of increasing SiO₂ content) toward the temperature minimum of the liquidus in SiO₂–NaAlSi₃O₈–KAlSi₃O₈–H₂O suggests the control of crystal–liquid equilibria (Tuttle and Bowen, 1958). Bowen (1928), with the Darwinian title of his famous book, clearly had a uniformitarian view in mind: infinitesimal fractional crystallization steps integrated over time. Absent inefficiencies in crystal extraction, whole-rock chemical variation would then represent a liquid line of descent. Each point in composition space would represent the product of crystal/liquid fractionation (sometimes including assimilation) that is derived from the adjacent point of lower SiO₂ content (or other index of differentiation) and in turn gives rise to the next point on the trend. Thus SiO₂ could be viewed as a proxy for time, though perhaps time is an exponential function of it, with an increment in weight percent of SiO₂ requiring much greater time at high SiO₂ than at low SiO₂ (e.g., Hawkesworth et al., 2000; Reagan et al., 2003).

The purpose of this paper is to present the case for these trends instead representing discrete magma batches that have very different histories, though sharing the same sources and processes in different proportions. Mixing or at least contact between batches is common and often triggers eruption. The mechanism of crystal/liquid fractionation changes from crystal extraction from a melt-dominant regime at the mafic end of the chemical spectrum to melt extraction from a crystal-dominant regime at the silicic end.

Readers may find this discussion selective and somewhat oversimplified, and undoubtedly it is. Our intent, however, is to show through some clear examples that chemical whole-rock trends are neither a one-way path nor are they necessarily conducted in incremental steps. Indeed, the textbook concept of protracted in situ fractionation in a central subvolcanic magma chamber is an oversimplification that has received insufficient critical examination and insufficient consideration of alternatives.

2. Background

A primary goal of igneous petrology is to understand how magmas evolve and give rise to the observed assemblages of igneous rocks. The verb “evolve” and the adjective applied to more silicic rocks as “evolved” themselves carry a connotation of slow inexorable change. Evolution is formalized in models of fractionation with variable assimilation (e.g., Spera and Bohrson, 2001; Bohrson and Spera, 2001) and bulk chemical trends are named “tholeiitic”, “calcalkaline”, and so on with the implication that they represent liquid lines of descent.

That bulk compositions describe a liquid line of descent requires a considerable leap of faith. The rocks being analyzed often contain abundant crystals of different origins and glasses that greatly diverge from bulk composition. If the crystalline components come from different sources, it is likely that the melt phases and even the volatile components do as well. Many volcanic and plutonic rocks were probably never all-liquid magmas.

Nevertheless, a continuous liquid line of descent from basalt to rhyolite does exist. It is most clearly seen in the crust of lava lakes (Helz and Thornber, 1987). In coring into these lakes, interstitial rhyolitic melt is first encountered when the low-pressure anhydrous solidus temperature of 1000 °C is reached. With depth, melt fraction increases while SiO₂ content of melt decreases until a high melt fraction of basaltic composition is encountered in the residual “magma” lens. Thus, the generation of a suite of melts from basalt to rhyolite is a relatively simple chemical and thermal problem. These are not magmas, however. They are only the liquid portions of a material that, if it were mobile, would still be a basaltic magma. The generation of a suite of magmas from basalt to rhyolite is a mechanical problem of separation of solid and liquid phases, just as the problem of tephra to lava is a mechanical problem of the separation of liquid and vapor phases.

3. Simplest scenario

The simplest way to derive a basalt to rhyolite suite is to extract crystals from a cooling magma body as soon as the crystals are formed. At each step, the

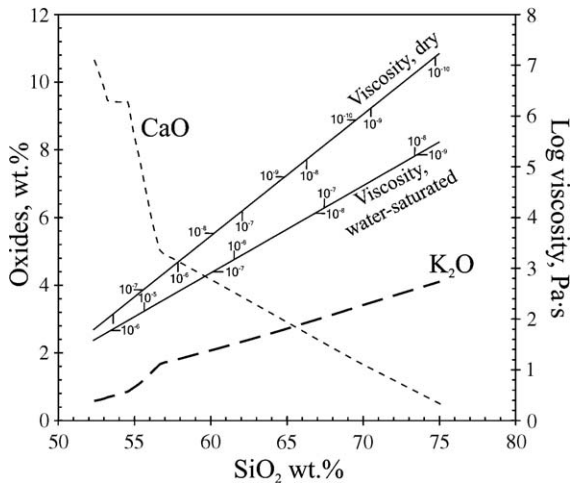


Fig. 1. Predicted simple fractional crystallization of the 1996 Academy Nauk high-alumina basalt showing behavior of K_2O and CaO with respect to silica (COMAGMAT algorithm, Ariskin, 1999) and the corresponding change in melt viscosity (CONFLOW algorithm, Mastin and Ghiorso, 2000) at 100 MPa total pressure. Numbers indicate Stokes settling velocities (m/s) for grains of 1-mm radius of plagioclase (horizontal ticks) and magnetite (vertical ticks). A velocity of 10^{-9} m/s is about 1 m in 30 years. For differentiation by crystal settling, other factors being constant, an order of magnitude increase in melt viscosity should require an order of magnitude increase in time.

system is redefined as the liquid. The bulk system trend will then be a liquid line of descent. Assimilation may occur at the chamber walls, balancing latent heat released by crystallization. As Bowen (1928) observed, assimilation generally increases the amount but not the major element composition of the end product.

Traditionally, crystal extraction is thought to occur by crystal settling (e.g., Bowen, 1928; Martin and Nokes, 1988). Plating out of crystals on the cold chamber walls with expulsion of melt as a thin ascending flow has more recently received attention (e.g., McBirney, 1980; Wolff et al., 1990), despite an absence of direct physical evidence for such a process. In either case, the separation of crystals from the melt in which they grew will be impeded by melt viscosity (Fig. 1). The dramatic increase of melt viscosity with increasing SiO_2 content and decreasing temperature leads to the expectation that the rate of fractionation should decline exponentially as the magma evolves. U–Th isotopic systematics have been interpreted as reflecting such slowing (Hawkesworth et al., 2000; Reagan et al., 2003). Long times require large magma bodies; otherwise the magma will cool faster than fractionation can progress. For

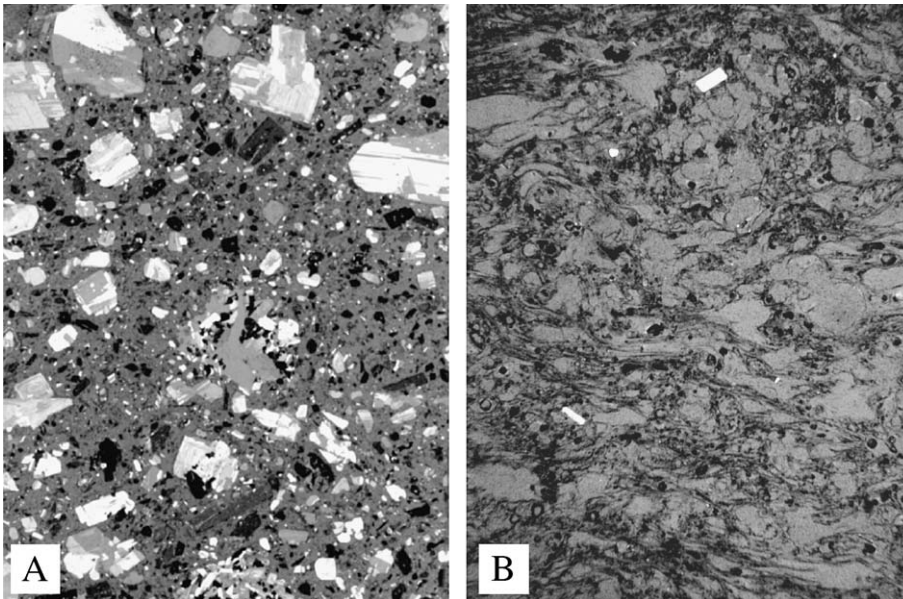


Fig. 2. Photomicrographs of (A) dacite of Unzen Volcano, Japan and (B) rhyolite of Novarupta, Alaska. The dacite is rhyolite melt plus abundant crystals and mafic enclave debris, yielding a dacitic bulk composition. The rhyolite is almost entirely rhyolite melt, yielding magma of nearly the same bulk composition. Widths of the field of view are 24 and 8 mm, respectively.

example, a generic 0.002-m-diameter crystal that is 500 kg/m³ denser than 10⁶ Pa s hydrous rhyolite melt will require 3000 years to settle 100 m, but a 10 km³ water-rich rhyolite magma chamber will be completely crystallized in 2000 years (Hawkesworth et al., 2000) to 10,000 years (Spera, 1980). This is part of the reason for proposing magma chambers that exceed erupted volume by an order of magnitude, even in the case of the largest eruptions (e.g., Smith, 1979). But geophysics has failed to reveal large mostly liquid bodies in the crust and in turn the concept of plutons as representing bodies that were at one time mostly liquid throughout is being reassessed (Glazner et al., 2004).

Melt-crystal separation efficiency might decline with increasing melt viscosity as well as separation rate, so that an increasing proportion of previously precipitated crystals might remain suspended in the magma as evolution proceeds. This would yield “dirty” rather than clean magmas at more silica-rich bulk compositions. Indeed, many andesites and dacites consist of abundant crystals suspended in rhyolitic melt (e.g., Nakada and Motomura, 1999) and these may be erupted again and again from the same center with little chemical change with time. Hawkesworth et al. (2000) attribute this to slowing of crystal settling. However, it can equally well be argued that this is because fractional crystallization has not just slowed but has stopped and that there exists no direct path from crystal-rich andesites to crystal-poor rhyolites (Fig. 2).

4. Evidence from natural systems

The examination of the chemical chronology of arc volcanoes yields little support for a simple magma evolution scenario, wherein a single large magma body undergoes internal differentiation to yield eruption products on an extended liquid line of descent. Although some volcanic fields may well display a dominance of mafic compositions early in their history (e.g., Eichelberger and Gooley, 1977), few if any known progressions of eruption product compositions with time could reflect eruptive withdrawal of samples from a gradually fractionating reservoir. The reality is quite chaotic (e.g., Nye et al., 1995; Gamble et al., 1999; Hobden et al., 1999;

Dungan et al., 2001). Moreover, “evolved” silicic magmas are sometimes seen to make a “surprise” appearance, without substantial precursors that could hint of protracted evolution (e.g., Hildreth, 1983).

An apparent exception to lack of evidence of protracted in situ fractionation of magma has been thought to be the voluminous chemically zoned ignimbrite sheets associated with caldera formation in arc systems (McBirney, 1980; Hildreth, 1981). These may span the range andesite to rhyodacite or rhyolite, sometimes including basaltic andesite compositions as well. The sequence is generally silicic to mafic: the most silicic magma is erupted first and the most mafic last and thereby stratigraphically highest.

For many years, the consensus has been that this eruptive stratigraphy represents an inversion of a stratigraphy that existed in a single source chamber (e.g., Smith, 1979). The question of how such an ordered stratigraphy could survive catastrophic eruption (e.g., Spera, 1984) has not diminished the firmness with which this view is held, perhaps because alternatives have not been seriously advanced. The further assumption is then made, often implicitly, that the stratigraphy had to develop through fractional crystallization and assimilation processes within the chamber and that the coerupted magmas have a parent/descendent relationship. Isotopic contrasts, which might otherwise point toward separate magma batches, are incorporated into this view—a small amount of contamination from isotopically very different chamber wall or roof rock instead of a large amount of magma from a subtly different source. The petrologist’s task is thus narrowly—and we suggest unnecessarily—circumscribed to determine how Magma A was derived from Magma B. Although not an arc volcano, nowhere has such a quest been more famously or inconclusively pursued than at Long Valley Caldera, where the coupling of a small contrast in major element concentrations with a large contrast in trace element concentrations has led to extensive debate about fractionation processes, some quite novel (Hildreth, 1979; Cameron, 1984; Anderson et al., 2000).

Such a quest is only meaningful if a fractionation relationship actually exists. Otherwise it becomes a comparison of “apples and oranges”. In fact, a se-

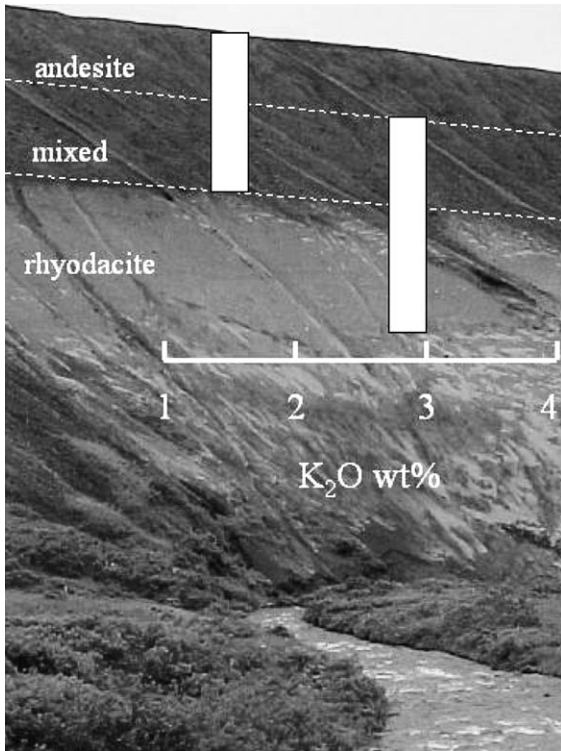


Fig. 3. Vertical profile through Aniakchak ignimbrite sheet showing K_2O concentrations in pumices with height through the Aniakchak Caldera ignimbrite sheet (Dreher, 2002).

quence of eruption characterized by changes in composition with time does not uniquely require a horizontally layered source chamber. A common order of eruption of magma types may simply reflect relationships of relative density and mobility or explosivity among magma batches to composition. For example, the normal “first out” position of rhyolite may be because it is least dense and able to rise through other magma batches prior to eruption, not that it formed atop those magmas. Nor does the pre-eruption arrangement of magma types have to be vertical. For the case of Long Valley Caldera, perhaps the most influential example in the development of the paradigm of a huge unitary chamber zoned by fractionation in situ, it now appears that the early highly evolved Bishop Tuff magma erupted mostly at the southern margin of the caldera, whereas later less evolved crystal-rich Bishop Tuff magma erupted from marginal vents some 15 km to the north (Wilson and Hildreth, 1997).

5. The problem of coincident storage

Where a discontinuity exists in melt composition and temperature (and the two must accompany each other), prolonged storage as a single body is thermodynamically impossible. Aniakchak Caldera, formed at 3400 a on the Alaska Peninsula, illustrates this situation well (Dreher, 2002). What might appear in the field to be an ignimbrite sheet gradationally zoned from a rhyodacite base to an andesitic top is in fact an ignimbrite sheet with a single-magma base, a central zone with two magma types, and a single-magma top (Fig. 3). There is a large gap in both bulk and melt compositions between the two component magmas of the sheet (Fig. 4).

The erupted assemblage cannot itself possibly be a liquid line of descent because of the jump in liquid composition. Fractional crystallization yields continuity in the compositions of generated liquids. One solution to this problem has been to postulate sidewall fractionation. Plating out of crystals on the cold chamber wall could release a silicic melt fraction, which might then ascend along the wall because of buoyancy, counter to thermally driven convective flow, and accumulate as a silicic cap (Shaw, 1974). The cap, it is thought, might not be mixed back into the parent mafic volume because of the contrast in density and viscosity. McBirney (1980) demonstrated the principle with a tank experiment in which a salt solution crystallized against a cold hemispherical cover, producing a

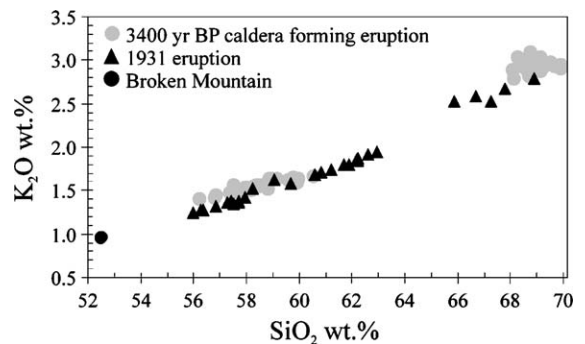


Fig. 4. K_2O vs. SiO_2 for the Aniakchak Caldera forming eruption at 3400 a (Dreher, 2002) and for an intra-caldera eruption in 1931 (Nicholson, 2004). The caldera-forming eruption involved two separate magma batches, one of andesite and one of rhyodacite. Subsequent small-volume mixed eruptions have begun to fill in the chemical gap and basalt has erupted within the caldera from the Broken Mountain vent.

salt-depleted cap. However, the cap is continuously, not discontinuously, zoned; the boundary-layer flow mechanism encounters serious fluid dynamical problems when applied to viscous magma (Carrigan and Cygan, 1986); and there is an absence of any direct evidence because the boundary layer, if it exists, is neither erupted by volcanoes nor preserved in plutons.

That the two Aniakchak Caldera magmas were coerupted is required by the considerable thickness of the central layer in which pumices of both magmas are present. That they came into contact at depth, below the level of fragmentation, is indicated by the presence of banded pumices. Yet the implied step functions in temperature and composition with depth require infinite vertical heat and heat fluxes, which would be impossible to sustain. For two magmas to be in contact without generation of thermally and chemically intermediate material means that the time that passed between the onset of interaction and eruptive quenching was negligible. But how long is that?

One might argue that chemical diffusion is so slow as to make this constraint unimportant. Only mixing between layers could produce rapid chemical change and mixing might be prevented by the stable density profile. However, this objection could not apply to thermal diffusion, the effects of which should be evident from resorption of phenocrysts from heat gain in the rhyodacite (if present; in Aniakchak rhyodacite they are sparse) and crystallization in the cooled andesite. Such textures can be recognized in effusive eruptions, where slower mobilization and eruption processes permit crystalline/melt phase re-equilibration in response to heat transfer (e.g., Coombs et al., 2000).

Suppose that there had been horizontal stratification in the source chamber beneath Aniakchak Caldera as well as time for interaction. The most easily recognizable feature in eruption products would be the microphenocryst-charged, glass-poor character of material from the cooled mafic side of the interface. Based on crystallization experiments, Coombs et al. (2003) concluded that such crystal growth would occur in a matter of days and so for the current purpose crystallization can be considered to be an immediate response to cooling. If the magma chamber had an areal extent comparable to the caldera, then contact-modified material would have been forming on a surface of 80 km² in area. Taking the characteristic thickness of a cooled

layer growing by conduction as $(Kt)^{1/2}$ and using $K = 6 \times 10^{-7} \text{ m}^2/\text{s}$ (Spera, 2000), after 1 year of contact the layer is about 15 m thick, about 1 km³ in volume, 2 vol.% of the deposit, or about one out of 50 pyroclasts. Such a component ought therefore to be readily detected in the field. Actually, the interface is preserved in the form of banded pumice and there is no evidence of a crystallization response in the andesite bands, hence the time between contact and eruptive quenching was less than days.

A more straightforward interpretation of a gap in chemical composition is that these are separate batches of magma that erupted as soon as they came into contact. There was not sufficient time for heat or mass transfer between them to alter their phase assemblages or chemistry. The ephemeral nature of a sharp interface seems borne out by the post-caldera eruption history of Aniakchak. The chemical gap at Aniakchak, so prominent in the caldera-forming event, has substantially closed in the subsequent 3400 years (Fig. 4). Also, basalt has appeared for the first time within the former central vent region, erupting on the caldera floor at the Broken Mountain vent. Apparently, small batches of magma residual from the caldera event were able to partially blend and have been triggered to erupt by new basaltic inputs. But basalt can now erupt directly, rather than as an admixed component of andesite, apparently because the large magma batches have been cleared from the system.

The two (or more)-batch interpretation of caldera eruptions is often supported by a gap in isotopic composition that corresponds to the chemical gap (e.g., Reagan et al., 2003). This is traditionally incorporated into a single-chamber fractionation model by postulating coupling of minor wallrock or “roofward” contamination to sidewall fractionation. But if the assumption that coerupted magmas have coevolved within a single source chamber is not justified, then much of the speculation involving “double diffusion” and “counter-buoyant flows” is unnecessary. In that case, however, we are unfortunately left without an obvious source for the silicic magma.

6. The case of some andesite/dacite systems

Whatever the difficulties in understanding individual eruptions, chemical continuity is generally ob-

served over long times, clearly implying relationships among magmas. Do basalt magmas differentiate to produce andesites, andesite magmas to dacites, and dacite magmas to rhyolites? Periods of andesite cone growth spanning thousands or even hundreds of thousands of years are sometimes punctuated by sudden catastrophic eruptions of rhyolite or rhyodacite (e.g., Bacon and Druitt, 1988; Izbekov et al., 2004a). Is this the culmination of slow fractionation in an andesitic system or something else? And what is the relationship to basalt? There is reason to believe that basalt is what is emerging from the mantle wedge to feed arc volcanic systems (e.g., Hildreth et al., 2004; Price et al., in press). While basalt is poorly represented in some arc settings, calcic plagioclase and magnesian olivine are commonly observed in andesitic magma suggestive of basaltic parentage (e.g., Izbekov et al., 2004b). Are these introduced by admixing of new basalt magma batches or residuals from inefficient fractional crystallization?

The current eruption of Karymsky Volcano is illustrative. The event began on January 2, 1996 with essentially simultaneous eruption of andesite from Karymsky's summit and basalt from a new vent, within Academy Nauk Caldera, 6 km to the south (Fedotov, 1998). The summit eruption continues as of this writing and represents resumption of frequent activity fed from a seismically inferred shallow magma chamber (Balesta, 1991). The basaltic vent lies on the same prominent north–south fault that the Karymsky cone sits astride. A geodetic net encompassing the volcanic system recorded a maximum extension located between the two vents of 2.3 m, perpendicular to the fault and vent trend during the onset of the eruption (Maguskin et al., 1998). The extension is measurable well north and south of the eruptive vents and large fissures opened along this trend, in some cases issuing steam.

It seems unambiguous that a north–south striking dike was emplaced beneath the two eruptive vents. The Karymsky andesite exhibits a dip in silica content of the melt phase at the start of activity and contains highly calcic plagioclase matching the major and trace element composition of plagioclase in the basalt erupted at Academy Nauk, as well as traces of magnesian olivine (Izbekov et al., 2004b; Izbekov et al., 2002; Eichelberger and Izbekov, 2000). Thin overgrowths on the calcic plagioclase indicate that it was

very recently introduced and olivine only survives where protected from melt by other crystal phases. The implication is that the dike was basaltic and erupted both directly to the surface and into the shallow Karymsky andesite magma chamber, triggering the eruption from Karymsky's summit (Fig. 5).

The interception of a magma pod by a dike may not be a coincidence because magma chambers may act as dike attractors. (We will use the word “pod” to emphasize that we envisage smallish, 1–10 km³ magma accumulations beneath andesite/dacite volcanoes, rather than much larger fractionating chambers.) An over-pressured chamber embedded in brittle upper crust will exert an extensional stress tangential to its surface. A cooling magma body will normally be over-pressured because it will be undergoing second boiling. The pod will therefore perturb the regional stress field in its vicinity, adding a component of extension tangential to its surface but of decreasing magnitude with distance from the body. A dike rising from depth will propagate normal to the least principal stress, which may be near horizontal far from the chamber. But the least principal stress direction must rotate with decreasing distance from the magma pod, as the growing extensional perturbation is added (Fig. 6). This would cause the dike to also turn and to propagate towards the pod. The effect is to increase the size of the “target” for the dike. Once the dike encounters the free solid-to-liquid surface of the pod wall, it may behave much like a dike encountering the earth's surface: magma flow across the surface into a fluid encounters less resistance than magma flow that would continue to open the dike fracture within solid rock and therefore is favored. The flow becomes channelized, in this case into the inflating magma pod (Fig. 6).

Such a process also yields chemical variation with a compositional gap in the eruption products. As with Aniakchak, the gap arises because these are separate batches of magma. In the Karymsky case, mixing and adjustment of phase relations occur in the andesite, but the amount of introduced basalt in the erupted mixture is sufficiently small not to shift the bulk composition of the andesite significantly toward basalt. There is no reason to believe, and indeed there is much reason to doubt, that the basalt and andesite came from a single zoned chamber. The gap in composition means that the andesite was not derived by simple fractionation from the basalt batch with which

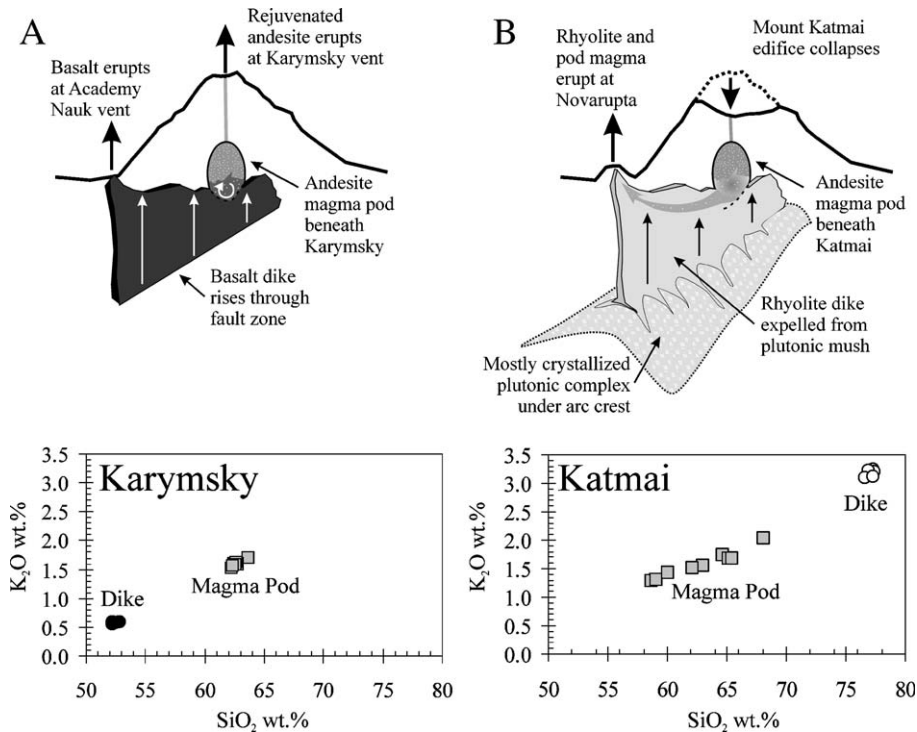


Fig. 5. Interpretive dike and magma pod relationships for the 1996 Karymsky and 1912 Katmai eruptions and resulting chemical patterns of eruption products (schematic; not to scale). A. Rejuvenation of an andesite magma pod beneath Karymsky by a basalt dike that also erupts directly to the surface 6 km to the south at the Academy Nauk vent. The eruptives are Karymsky andesite with mild contamination by the recent basaltic component and uncontaminated basalt, separated by a large chemical gap. B. A rhyolite dike simultaneously erupts on the floor of the Valley of Ten Thousand Smokes and ruptures the crystal-rich andesite–dacite magma pod beneath Mount Katmai, draining it and causing Mount Katmai to collapse. The andesite–dacite pod or pod complex exhibits internal mixing but was in contact with rhyolite so briefly before eruption that no intermediate compositions were generated. It is therefore separated from the 1912 rhyolite, which is nearly pure recently extracted melt, by a large chemical gap.

it erupted. This does not mean, however, that the andesite could not have been derived from a similar basalt (or more likely many similar basalts and other ingredients) some time in the past.

Karymsky has been erupting similar crystal-rich andesite for about 5000 years (Braitseva and Melekestsev, 1991). Plagioclase phenocrysts exhibit calcic zoning spikes, large compared to what could be expected from anything but shifts in nutrient melt composition (Izbekov et al., 2002). This suggests that the pod magma has undergone numerous mafic recharge events during its history, similar to the 1996 event. As such, the pod magma, while sustained thermally by these inputs, has a complex origin, with no single parent. Because it is crystal-rich with crystals from multiple sources, its bulk composition was never a liquid composition.

At Karymsky, the simultaneous chamber injection and direct eruption of a rising dike, while not without precedent, is fortuitous because it provides us with an unambiguous sample of the dike magma. The relative rarity of such events is consistent with the presumption that once magma ruptures a free surface, whether earth's surface or a chamber wall, subsequent flow is likely to be channelized through the rupture rather than contributing to further dike propagation. Thus, a basalt eruption or a basalt-triggered andesite eruption is much more common than both happening simultaneously.

In many andesitic to dacitic eruptions it is possible to “see” a newly introduced basaltic component in the form of mafic enclaves and therefore recognize that the Karymsky-type event is a special example of a general case of basalt injection triggering eruption

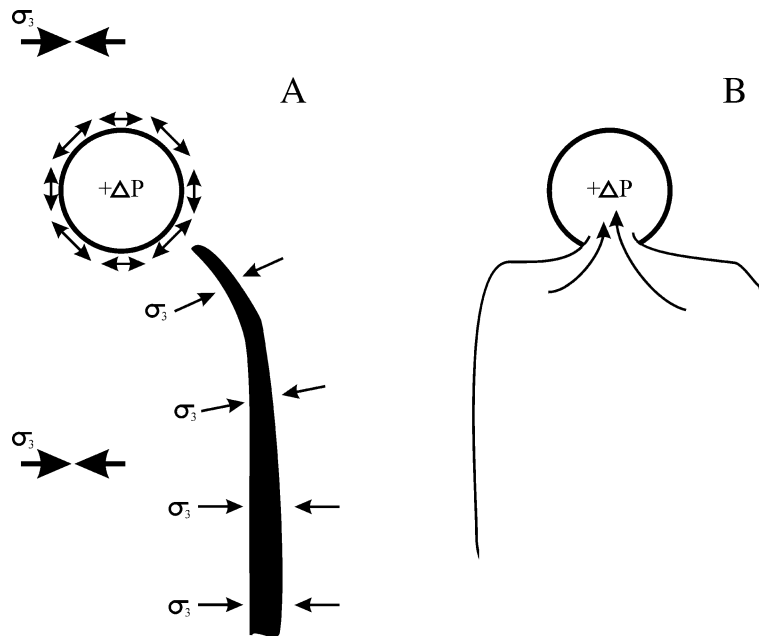


Fig. 6. Attraction of rising dike magma by a magma pod embedded in the crust. A. A cooling pod will be over-pressured due to second boiling. This exerts a tangential extensional stress at its surface, decreasing radially away from its surface. The least principal stress in the far field is everywhere horizontal, and dikes rise on planes perpendicular to this direction. However, in the near field the extensional stress field surrounding the pod must be added to the regional field, rotating the direction of least principal stress toward tangential to the pod surface. The result will be to turn the propagating dike tip toward the pod. The overall effect is to increase the size of the target for the rising dike. B. The dike propagates upward by fracturing solid rock. However, once a pathway to the pod is opened, dike magma can more easily inflate the pod than continue to drive fracturing of rock. Thus dike flow will be channeled into the pod.

(Fig. 7). Examples include: Unzen, Japan (Nakamura, 1995); Pinatubo, Philippines (Pallister et al., 1996); Soufriere Hills, Montserrat (Murphy et al., 2000); and Trident Volcanoes, Alaska (Coombs et al., 2000). Such clotting or coagulation, sometimes referred to as “mingling”, occurs when a large temperature contrast exists between intruding basaltic magma and resident pod magma, inducing rapid crystallization of the mafic magma and derivative mixtures and concomitant heating of the andesite or dacite magma with which it is in contact.

The effusive eruption products of these interactions exhibit two dominant textural types: enclaves and host lavas. Enclaves, dominantly plagioclase and amphibole or plagioclase and pyroxene, are fine-grained and exhibit crystal forms indicative of rapid crystallization in an initially melt-rich system (e.g., Eichelberger, 1978). They contain interstitial glass broadly similar to the glass of the host lava and vesicles and some contain reacted and/or resorbed phenocrysts similar to crystals present in the host. The host lava is typically

porphyritic but with a disequilibrium phenocryst assemblage that includes both reacted and/or resorbed phenocrysts appropriate to relatively silicic magma (significantly, quartz, a relatively rare phenocryst phase in arc eruptives), crystals appropriate for precipitation from basalt, and crystals that do not exhibit strong disequilibrium effects. Of note is the side-by-side occurrence in the same thin section of plagioclase with resorbed sodic cores and thin, strongly normally zoned rims and similar sodic plagioclase without the resorption and overgrowth features. These features can be readily interpreted as resulting from interaction between resident chamber magma and injected basaltic magma.

Whole-rock chemical variations produced by basalt injection into andesite/dacite magma pods exhibit similar general patterns (Fig. 8). Enclaves range from basalt or basaltic andesite to andesite. Host lavas range from silicic andesite to dacite. There is normally a gap between enclave and bulk lava composition.



Fig. 7. An enclave-rich dacite of Mutnovsky Volcano, Kamchatka, Russia. Inset shows close-up of enclaves in dacite of Unzen Volcano, Japan (width of field of view is approximately 30 cm). Basaltic recharge exhibits this clotting behavior if the temperature contrast between new magma and resident magma is large, thereby forcing crystallization of the injected magma before mixing can achieve homogeneity.

The development of the petrologic features and the bulk compositions indicative of basaltic injection can be divided into two stages: prompt response and gradual restoration of equilibrium (Fig. 9). These stages were described by Heiken and Eichelberger (1980) and further explored by Tepley et al. (1999) for the Chaos Crags dacite domes of Lassen National Park. Additional discussions are found in Eichelberger (1980) and Sparks and Marshall (1986). The basaltic magma is at first highly fluid upon injection. It readily incorporates (given the contrast in viscosities, the resident magma behaves almost like a solid) the resident magma in proportions that approach 1:1 resident magma to basalt. The liquids of the respective magmas mix completely, but phenocrysts from the resident magma often survive to become the reacted and resorbed phenocrysts of the enclaves. This process of direct mixing, a sort of bulk assimilation of andesitic or dacitic magma by injected basalt, is limited by the crystallization that is induced in two ways: (1) advective cooling by the mixing itself (this will be important if liquidus temperatures are not linear with composition—i.e., if the andesite bulk composition liquidus is comparable to, not lower than, the basalt composition liquidus, e.g. Piwinski, 1968) and (2) rapid conductive cooling because heat flows out of the dispersing basalt-rich enclaves into the “cold” basalt-

free host magma centimeters away. Crystallization within the enclaves, which at first lags temperature changes, shuts off the fluid mixing process but sets the stage because the enclaves are now mush rather than completely solid, for protracted mechanical dispersal of the crystalline debris of this interaction. Resorbed host-derived phenocrysts in enclaves may be returned to the host (Tepley et al., 1999). Thus, crystals with apparently completely different histories may never have been more than centimeters apart. Their differences reflect the extreme thermal and chemical disequilibrium that existed very briefly during the recharge event. The puzzling occurrence of “cooked” plagioclase in both the enclaves and host, which would seem to imply some sort of exchange, is really the result of the two sequential processes of mixing, the first fluid and the next mechanical. The first is true mixing in that it produces a new melt composition through stirring and a relatively homogeneous material on the macro-scale. The second is “mingling” in the sense that the products may remain heterogeneous on the macro-scale and changes in melt composition occur through thermal equilibration rather than direct interaction. However, it should be noted that mechanical dispersal could be so thorough as to produce a homogeneous product at the macro-scale, so that the term “mingling” becomes misleading.

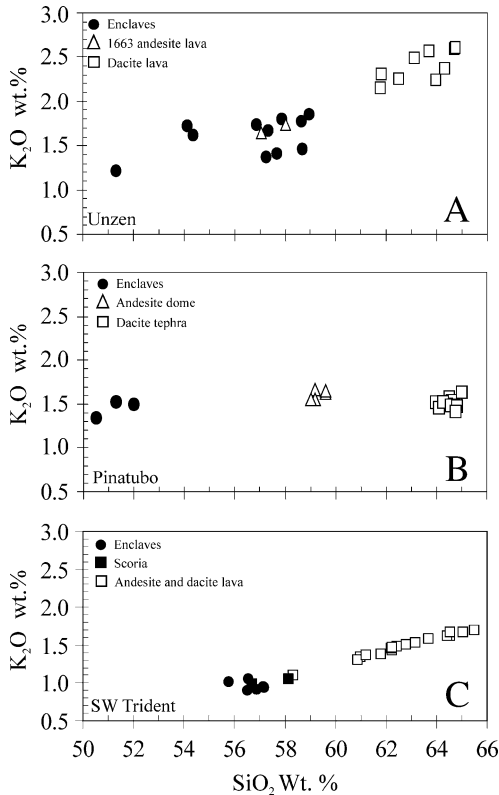


Fig. 8. Chemical patterns produced by basalt injection into andesite/dacite magma chambers. A. Chemical variation at Unzen (Browne et al., 2003). Variation from basalt to dacite is generated by contrasting mechanisms of mixing that approach each other from the extremes of introduced basaltic magma and resident reservoir magma. B. Chemical variation at Pinatubo (Pallister et al., 1996). What might have been simply an Unzen-like event characterized by extrusion of enclave-laden lava was interrupted by wholesale dumping of the largely uncontaminated reservoir magma. C. Trident variation diagram (Coombs et al., 2000). Similar to Unzen but some mafic input has erupted directly as scoria, perhaps because the reservoir is small enough to permit passage of forcefully injected mafic magma without dispersal as enclaves.

Prompt fluidal mixing produces the range of enclave compositions, pulling their initially basaltic composition towards the host magma composition, whereas the subsequent mechanical dispersion pulls host chemistry towards the basalt. But a chemical gap remains between enclaves, where the component of assimilated resident magma ranges from close to 0 wt.% up to 50 wt.%, and host lava, where the component of entrained enclave material is usually below 25 wt.%. At arc volcanoes that normally erupt pod magma containing enclaves, however, there are rare cases of

more mafic eruption products that contain no enclaves yet overlap enclaves in bulk composition. These are essentially enclave magma that erupted without going through the chamber dispersal, thermal equilibration step that causes crystallization. For example, several years of effusion of enclave-bearing andesite and dacite lavas at Trident Volcano (4 km south of Novarupta vent in Katmai National Park) were interrupted by mafic scoria that matches the enclaves in bulk composition but lacks their crystallinity (Coombs et al., 2000). This scoria apparently represents basaltic magma injected so forcefully into the andesite/dacite pod that, although it incorporated some resident magma during its passage, it erupted immediately without dispersal into the chamber. Similarly, 1663 andesite lava at Unzen resembles the most silicic enclaves in Unzen dacite lava and apparently represents prompt eruption from the contact zone of mixing between resident and intruding magma (Fig. 10),

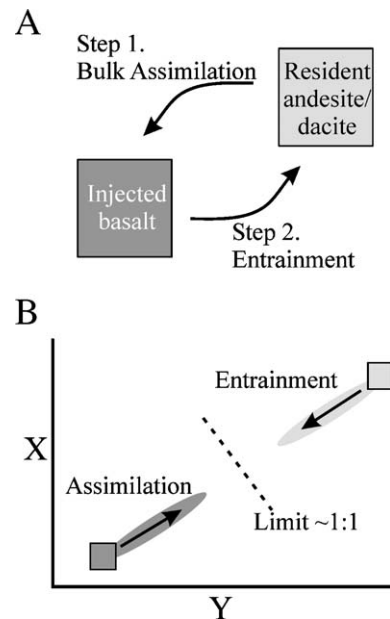


Fig. 9. Steps in mixing between reservoir magma and injected basalt. The first step involves both magmas in a liquid state, though the reservoir magma is so much more viscous it may be thought of almost as a solid that is assimilated. This process may be over in a matter of hours, as it is limited by crystallization of the mafic magma that contact with the silicic magma induces. Once thermal equilibration and resulting crystallization are achieved, the viscosity relations are reversed and it is the mafic domains (now variably contaminated) that are relatively solid. This material is dispersed mechanically as enclaves.

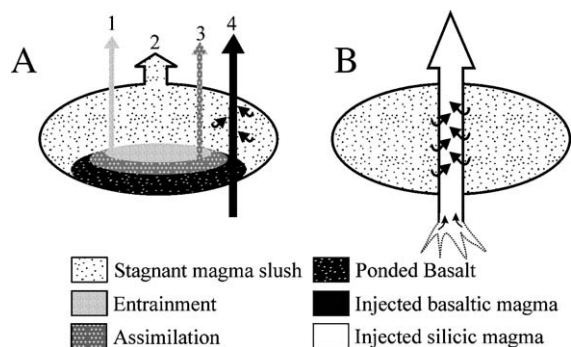


Fig. 10. Schematic sketches of inferred behavior of andesite/dacite magma pods during replenishment, adapted from models of Pallister et al. (1996) and Koyaguchi and Keneko (2000). A. Basaltic magma ponds at base of magma pod. Eruption product is normally (1) lava consisting of heated and mobilized reservoir magma with entrained crystalline mafic debris from basalt. However, with very large basaltic input, as at Pinatubo, uncontaminated portions of the reservoir may also be expelled as dacite tephra (2). Also less commonly, the zone of assimilation is erupted directly without the enclave entrainment and crystallization step (3) as in Unzen 1663 andesite lava. Basalt can also penetrate through the pod with variable mixing and erupt explosively as scoria (4) as at Trident. B. Newly generated silicic magma passes rapidly through an andesite/dacite pod with little interaction, erupting promptly and flushing some resident andesite/dacite magma with it, creating a zoned ignimbrite. It should be emphasized that the zoning depicted in these schematic sketches is highly ephemeral, formed immediately in response to replenishment rather than representing a long-existing stratigraphy.

through the “tunneling” mechanism proposed by Bergantz and Breidenthal (2001).

The existence of a sharp chemical boundary between host and enclave in enclave-bearing lavas might seem to contradict what was said earlier about the impossibility of sustaining a chemical discontinuity. However, chemical equilibrium is almost achieved across the enclave border (crystalline phases are jacketed with zones of similar composition to comparable phases in the host and the melt phase is the same composition); the assemblage of host and enclaves has simply become an equilibrated but heterogeneous mixture. Even here the boundary is ephemeral, however, because the enclaves are evidently subject to mechanical disintegration.

Wiebe (1994) and Wiebe et al. (2002) have shown through plutonic examples that basaltic magma injected into resident granitoid magma can behave like a lava flow on the chamber floor or directly disperse into an enclave swarm. In either case, the zone of

interaction is considerably smaller than the resident magma body, which may have been more a stagnant slush than a proper magma when the injection event took place (Fig. 10). Koyaguchi and Keneko (2000) have modeled this situation, arguing that the eruption product is a selective sample of the chamber magma that comes into close contact with introduced basalt. The far-field stored magma receives only the time-integrated heat flux from repeated recharge events. It would also receive no direct chemical influence from these events.

This concept helps to explain why frequent recharge events often do not show a lasting effect on pod composition. The Unzen system has been predominantly erupting enclave-bearing dacites for some 500,000 years (Hoshizumi et al., 1999). The 1991 eruption of Pinatubo began in very much the same way as a typical Unzen eruption, with the extrusion of enclave-bearing dome lava, representing selective eruption from near the basalt/pod interface of directly affected pod magma Pallister et al. (1996). This was followed quickly by the massive tephra eruption of dacitic pod magma that was chemically unaffected by the recharge event (Figs. 8 and 10). The Pinatubo eruption illustrates the volumetric dominance of chemically unaffected magma in the source pod, whose eventual evisceration was probably due to exceptional and continuing vigor of the basaltic recharge event: more new magma entered the pod than could be accommodated through dome extrusion.

These processes of interaction produce interesting patterns in chemical variation. Although mixing might be expected to generate a continuous range of compositions, in fact it can produce quite discrete populations. In the case of enclave composition as compared to Unzen 1663 andesite or Trident 1950s scoria (Fig. 8), different histories can lead to the same bulk composition. Variation in bulk composition is close to linear, meaning that there is not selective incorporation of melt or crystal components of one magma in another. The most obvious exception to linearity is phosphorus, which may be subject to vapor transport and therefore be an indicator of protracted chamber residence (Dreher, 2002). This issue will be dealt with in a subsequent paper.

Why do we see this rich development of features in response to basalt–andesite/dacite encounters but not

in response to the andesite/dacite–rhyodacite/rhyolite encounters cited earlier? One reason may be that the former takes place at higher temperature and so reactions and crystal growth are faster. More importantly, if basalt is a new input to the system, its greater density will hold it in contact with the andesite/dacite pod magma (Fig. 10). In contrast, a silicic input will be able to rise rapidly through the pod and erupt, minimizing time available for interaction. Another factor to consider, regardless of the geometry of the encounter, is that heating of silicic magma makes it more likely to erupt, whereas cooling of more mafic magma makes it less likely to do so.

Where are the Unzen and Pinatubo systems headed in an evolutionary sense? The answer seems to be nowhere. The dominant magmatic volume beneath Unzen and Pinatubo and similar arc systems is a crystal-rich slush of andesitic to dacitic bulk composition. The melt phase is rhyolitic in composition and sufficiently evolved to be saturated with quartz. The system has thus “arrived” at its petrologic destination, the end of its liquid line of descent, in terms of generating a eutectic-like liquid. It has not however produced a rhyolitic magma—a mobile, eruptible volume of rhyolitic bulk composition. The crucial step of mechanical separation has not occurred, nor does it show any sign of occurring since at Unzen andesitic/dacitic eruptions have spanned half a million years (Hoshizumi et al., 1999). The viscosity of the rhyolitic melt is so high as to preclude significant separation of phases. Rhyolitic melt is common as the liquid fraction of arc eruptions, but rhyolite magma is comparatively rare.

7. Katmai: a silicic complement to Karymsky

The long (millennia to hundreds of millennia) stretches of modest central vent andesitic effusions at arc volcanoes are sometimes punctuated by paroxysmal eruptions of rhyolite or rhyodacite spanning hours to perhaps weeks. As in the Aniakchak case cited previously, these almost entirely explosive eruptions in volcanic arcs are commonly accompanied by an andesitic to dacitic component (e.g., Hildreth, 1981; Wilson, 2001).

The Katmai rhyolite-dominant eruption of 1912 may be, like Karymsky in 1996, a case of simulta-

neous eruption of new dike magma and intercepted pod magma. Like Karymsky, activity was centered at two vents several kilometers apart. Like Karymsky, typical andesite/dacite magma associated with a central cone was erupted. Like Karymsky, the other coerupted magma was at the extreme of the chemical variation trend and separated from the cone magma by a large chemical gap. But in the Katmai case, the chemical outlier was rhyolite, virtually all magma exited the distal vent, and the central cone collapsed. We have therefore suggested that 1912 Katmai is the rhyolite complement to 1996 Karymsky: the interaction of a rhyolitic rather than basaltic dike with an andesite/dacite magma body beneath Mount Katmai (Eichelberger and Izbekov, 2000). Because the magma in the dike that ruptured the floor of the Katmai pod was less dense than the Katmai magma and the dike was already breaching the surface to the west, flow reversed and the denser Katmai magma drained into the dike, emerging at Novarupta (Fig. 5). Although caldera collapse at Katmai without eruption at the site of collapse has been thought odd, it is not without precedent. Analogs include a basalt dike draining a basalt pod of different composition under Tolbachik Volcano, Kamchatka (Fedotov et al., 1983), in 1975 and a basalt dike draining an andesite pod under Miyakejima Caldera, Japan, in 2000 (Murse and Kimata, 2004).

For silicic compositions, the generally accepted interpretation of chemical heterogeneity is to postulate that the source is a layered magma chamber that fractionated in situ. Hildreth and Fierstein (2000) have presented the 1912 Katmai eruption as a case of a zoned chamber in which the rhyolite cap was cleanly decanted through a sill to erupt 10 km distant at Novarupta. They cite the chemical similarity of the rhyolite magma to the melt fraction (i.e., whole rock minus crystals) of the coerupted dacite as evidence that the former was derived by crystal extraction from the latter. Yet there is a gap in U–Th isotopic composition that corresponds to the gap in bulk composition and which can be interpreted as indicating that the rhyolite is in some sense older (Reagan et al., 2003), precluding a simple fractional crystallization relationship. The rhyolite plots on the equiline suggest that U/Th fractionation last occurred several half-lives or hundreds of thousands of years ago.

Rather than suggesting a separate source for the rhyolite, Reagan et al. (2003) argue that the rhyolite accumulated slowly above a fractionating, open-system (recharging) andesite/dacite zone, whose $^{230}\text{Th}/^{232}\text{Th}$ vs. $^{238}\text{U}/^{232}\text{Th}$ correlation corresponds to an “age” of 40 ka. This correlation may be a mixing line, with dacite—not the rhyolite—and unseen basalt as end members, broadly reflecting an average pod residence time. An identical andesite to dacite range of bulk chemical compositions at neighboring Trident volcano has been shown by Coombs et al. (2000) to arise from mixing. Nothing like the 1912 rhyolite had been erupted from the system before, so it seems unlikely that it had been accumulating as magma at the roof of the Katmai chamber for hundreds of thousands of years.

But while the 1912 dacite could not be the source of the 1912 rhyolite despite the dacite’s chemically equivalent matrix glass, the chemical and isotopic relationships demonstrate that extracting melt from an older dacitic body chemically like the 1912 dacite but isotopically like the 1912 rhyolite could produce the 1912 rhyolite. This is because no change in U/Th occurred in the dacite to rhyolite step, as can be readily seen in the whole rock data. If the dacitic source body, presumably a near-solidus granodiorite pluton, was old enough to be on the equiline, the extracted rhyolitic melt would be on the equiline as well.

We will return to this point later, but a remark should be made about the source depth of the rhyolite. Experimental work has suggested that, assuming vapor saturation in a water-only volatile system and applying measured rates of crystal growth to the sparse, small crystals of the 1912 rhyolite, the depth of last equilibration was shallow (~4 km) for at least months before eruption (Coombs and Gardner, 2001). This could argue for a sill rather than a dike source, though it is not clear that a dike source could not also be shallow. Conversely, shallowness of derivation does not necessarily favor the zoned magma chamber hypothesis because the same phase equilibria approach indicates the same or even shallower depth of last equilibration for the dacite and andesite (Hammer et al., 2002). Recently, magnetic evidence for the hypothesized dike connecting Katmai and Novarupta has been found (Hill, 2004) and will be discussed in a subsequent paper.

8. The source of some silicic magmas

The eruption of andesites as suspensions of crystals in rhyolite melt (Fig. 2) over long time spans suggests that separation of crystals from rhyolitic melt is a difficult process, as one might expect from crystal settling times in rhyolite (Fig. 1). What is surprising, however, is that when crystal–rhyolitic melt separation does occur, it takes place with remarkable efficiency (Fig. 2) and, judging from the sometimes absence of precursors, with equally remarkable rapidity. We do not see increasingly inefficient removal of crystals from earlier stages of fractionation and eruptive evidence of buildup to evolved compositions over a protracted period. Rather, we see an apparent halt to crystal–liquid separation in the intermediate bulk composition range and then re-emergence of clean fractionation at the rhyolite end of the spectrum.

A key characteristic of rhyolites in volcanic arcs is that they are often crystal-poor or even aphyric yet are multiply saturated. This is true for the 1912 Katmai rhyolite (Hildreth, 1983; Coombs and Gardner, 2001). The near-liquidus nature of such rhyolites has been attributed to maintenance of high temperatures through contact with hotter mafic magma. Yet evidence presented here suggests that the consequence of prolonged contact is mixing and these rhyolites contain no admixed or entrained mafic component (an exception is last-erupted Novarupta Dome, but clearly contact with mafic magma had by then occurred). Even if heating could occur without mixing, it would likely be irregular in effect, resulting in erratic variations in phenocryst assemblage and even in superheated (crystal-undersaturated) magma. To believe that a rhyolite owes its paucity of crystals to heating by mafic magma, one is left with the paradoxical appeal to perfect mixing within the rhyolite magma to produce its homogeneity simultaneous with a perfect absence of mixing with the hotter magmatic heat source with which it is in contact.

If a special heating mechanism is not an option, then the aphyric nature of some rhyolites should be attributed to recent extraction from their sources. Multiple saturation is then evidence that the source is poly-mineralic. Thus it seems likely that these rhyolites were interstitial melts in nearly crystallized plutonic bodies (and/or in crust beginning to melt) until shortly before eruptive quenching. The neces-

sary characteristics of such a source would seem to be that it:

1. Contains sufficient melt fraction so as to be permeable, in order that the melt can flow out of the system.
2. Contains sufficient crystal fraction to form a network so that:
 - a) The crystals are locked in place and left behind by the escaping melt.
 - b) The network has sufficient rigidity to support anisotropic stress and to sustain fractures, so that diking and rapid transport of melt is possible.

Such a scenario is depicted in Fig. 11. Consider the situation of a mostly crystallized pluton, which has gained sufficient strength to support the external anisotropic stress field. The continuous melt phase is in pressure equilibrium with the least principal stress. A deviatoric stress is applied and this deforms the material, decreasing the pore space and thereby causing an increase in melt (pore) pressure. The melt pressure now exceeds the least principal stress and, when the strength of the crystal framework is exceeded as well, will propagate a fracture as a melt dike in a plane perpendicular to the least principal stress (Rubin, 1995a,b). In the normal faulting regime commonly associated with silicic calderas, this direction will be vertical. A family of such fractures will form and, since they are unlikely to be absolutely parallel, will

intersect each other and support extensive, rapid melt extraction. If the melt, now magma, dike can rupture a free surface then a substantial quantity of pure melt can be expelled before collapse of the crystalline framework in the source eliminates permeability.

Imposition of the additional stress on the hyper-solidus body needed to trigger melt extraction could be by tectonic loading during some portion of the earthquake cycle. It could also be from the buoyant force applied by arrival of new basaltic magma from depth. Alternatively, the rise in pore pressure might be generated internally through exsolution of gas by second boiling as the melt crystallizes anhydrous phases. Sisson and Bacon (1999) and Eichelberger (1980) have explored this mechanism for mafic magma.

Perfit et al. (1980) noted the chemical resemblance of Aleutian rhyolites to the aplites contained within Aleutian granodiorites. Such features represent the extraction of late stage melt from the plutons (Fig. 12) (Miller and Mittlefehldt, 1984). Marsh (1995) considered extraction of melt from within a chamber mush layer. Bachmann and Bergantz (2003, 2004) have taken this “filter pressing” process out of a central chamber and considered it to occur in a compacting batholith. They concluded it is effective though quite slow, as long as porous flow is the mechanism of transport. We suggest that if short-distance porous flow is combined with long-distance dike flow, as in aplite dikes, the extraction process could be very rapid. Perhaps it could be rapid enough to actually sustain a large eruption or perhaps there would still need to be a brief stage of accumulation of newly separated rhyolite magma above the source body.

For the aplite dike case, the state of the pluton is one in which both the crystalline network and the melt phase are continuous. This is analogous to conditions under which gas is hypothesized to separate from silicic magma (Eichelberger et al., 1986): continuous melt and gas phases comprise a permeable foam and gas escapes from a melt network formed of the walls of partially coalesced bubbles. When pathways are open to a lower pressure regime, the gas in the case of ascending magma and the melt in the case of the pluton are able to flow into it, whereas the melt network in the case of ascending magma and the crystalline network in the case of the pluton are not. A large amount of gas can be extracted from a silicic

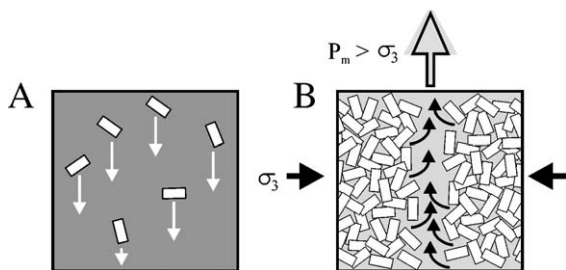


Fig. 11. Postulated mechanisms of fractionation, dependent on melt viscosity. A. Mafic (fluid) melt: extraction of crystals from melt-dominant system, conventional fractional crystallization. B. Silicic (viscous) melt: extraction of melt from crystal-dominant system. This is forcible expulsion of rhyolitic melt from a mostly crystallized pluton. A rise in pore (melt) pressure due to vapor exsolution from crystallization of melt or reduction in porosity from deformation of the crystalline framework produces diking perpendicular to the least principal stress and rapid extraction of melt, with crystals locked behind in the crystal framework.

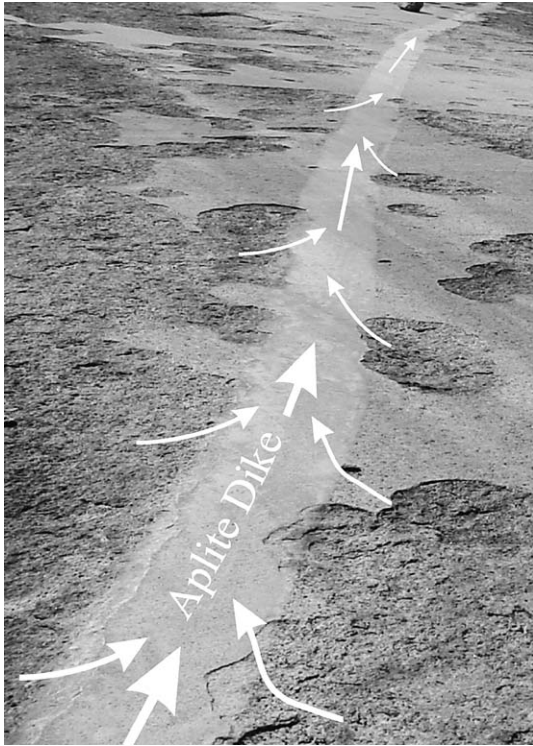


Fig. 12. An aplite dike in granodiorite of the Sierra Nevada batholith. Melt traveled by porous flow through the crystal network (small arrows) to join sheet flow within the dike fracture (large arrow), where it can be properly considered rhyolite magma. It is suggested that such a process could feed rhyolite eruptions, perhaps even during an eruption without an intermediate step of collection in a magma chamber.

magma before the foam collapses and becomes impermeable and perhaps this is the case for extraction of melt from a crystalline framework as well.

For a pluton or magma pod, if the melt fraction is greater than a critical value the material will behave as a suspension, like the Unzen and Pinatubo dacites, and crystals and melt will not separate but will flow together. With decreasing melt fraction, the material becomes uneruptible in bulk. *Marsh (1981)* speculated that the critical melt fraction for eruptibility is around 50 vol.%. But we argue here that the melt fraction still remains eruptible, down to some yet-to-be determined second critical porosity, at which point the crystal network becomes impermeable to melt flow and the melt cannot escape.

How fast could melt be extracted? The key is that the opening of multiple fracture pathways by excess

pore pressure in the source volume would create a huge surface area across which even a slow melt velocity would create a huge flux. Consider the case of 20 vol.% interstitial rhyolitic melt in a $1 \times 1 \times 1$ km granodiorite pluton. If the volume becomes laced with 100 aplite dikes, an average of one dike per 10 m, and we allow porous flow of melt at 10^{-3} m/s into the dikes, the entire volume can be emptied of melt in 2 h, a rate sufficient to sustain a Plinian eruption. Decades ago, *Jahns and Tuttle (1963)* hypothesized that the small grain size of aplite dikes as compared to their hosts is due to a “pressure quench” and the easiest way to produce such a quench would be through a massive eruption.

The puzzling paucity of quartz phenocrysts in silicic arc eruptives is consistent with this fractionation mechanism. Quartz would not be precipitated in the dominantly intermediate composition arc plutons until the crystal framework begins to connect and so is locked behind and prevented from extraction and eruption. But quartz does appear in basalt-activated dacite lavas because these are mobilized slushes that were on the verge of connectedness at an advanced state of crystallization. Seen in this way, moderately crystallized arc magma pods can yield andesitic to dacitic lavas upon basalt injection, but with greater crystallization only melt can be produced from them, yielding rhyodacitic to rhyolitic eruptions. The crystal-poor nature of Aleutian rhyolites and rhyodacites combined with the apparent rarity of Aleutian plutons of comparable composition points toward a conclusion that once these low-density magmas are formed, they are not stored but are rapidly, perhaps immediately, erupted.

9. Summary

The mechanism of crystal–liquid fractionation may be fundamentally different for systems containing viscous silicic melts from those containing fluid mafic melts. The result of these contrasting mechanisms is the Daly gap. “New” magmas, that is, magmas newly released from a mostly solid source, tend to be either basaltic from the mantle or rhyodacitic or rhyolite from the crust. Basaltic magmas may undergo classical fractional crystallization due to gravitational crystal settling, combined with assimilation at the crust–mantle

boundary or within the crust, but no farther than intermediate (andesite/dacite) bulk composition. This is because viscosity of melt rises with fractionation, so that when intermediate bulk composition is reached magmas consist of crystals effectively suspended in rhyolitic melt. Magma pods of intermediate composition can be maintained in arc crust for long times in a dynamic equilibrium. Their high crystal content reduces buoyancy forces to within what the cold upper crust can sustain. Repeated basaltic recharge events offset thermal losses by conduction and eruption. The pods also resist chemical change. Backward evolved magmas, that is, hybrids formed during recharge, are promptly erupted and forward evolution is inhibited by high melt viscosity. However, if crystallization advances, these pods become subject to a second form of fractionation: stress-induced expulsion of melt from the near-solidus system through coupled porous and fracture flow. Resultant low-density magmas are not retained within the dense crust of immature arcs but are promptly erupted.

This is a scenario appropriate for crust where little fertile material is available for production of silicic melts other than comagmatic arc plutons (Vogel et al., 2004). However, the physics of melt extraction from hot crust may be quite similar for both the oceanic and continental cases. (Critical porosities for porous flow may differ, however, because of textural differences between melting crust and crystallizing plutons.) Hence, in a more continental environment, extraction of melts from old crust occurs in addition to or in place of melts from comagmatic plutons (Huysken et al., 2001). There will also be greater retention of rhyolitic melts as magma bodies as opposed to immediate eruption, so that more crystal-rich rhyolites will be seen. Indeed, in a truly continental environment, the stable long-term magma pod may more often be crystal-rich rhyolite rather than andesite/dacite and rhyolite–rhyolite interactions may be the norm (e.g., Vogel et al., 1989; Smith et al., 2004). Isotopic contrasts between mafic and silicic magmas then become apparent in long half-life systems such as Rb–Sr as well as in short half-life systems like the U–Th decay chain. The greater relative volume of silicic magmas and greater isotopic contrast simply represent greater involvement of old crust as opposed to new comagmatic crust as the source of silicic magmas. The state of stress in the

crust may also play a critical role in determining whether rhyolitic melts are extracted from hypersolidus volumes, a normal faulting regime with the least principal stress horizontal being preferred.

Acknowledgments

We have been greatly influenced by the work of RA Wiebe, specifically by his concepts that: (1) crustal magma chambers undergo silicic as well as mafic recharge; (2) the details of magma interaction during mafic recharge can be seen through composite mafic–silicic intrusive complexes (MASLI) as exposed along the Maine coast; and (3) most granitoid plutonic rocks have lost melt after emplacement. We have likewise benefited from TA Vogel's insights into arc volcanism and discussions of stress in the crust with Jeff Freymueller. The manuscript was made understandable (we hope) through the comments of Jon Davidson, John Gamble, and Ben Edwards. This work was supported through the NSF Volcanology and Petrology Program grant EAR0310406 and the USGS Volcano Hazards Program under a cooperative agreement for the Alaska Volcano Observatory.

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