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Composition gaps, critical crystallinity, and fractional crystallization in orogenic (calc-alkaline) magmatic systems

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Abstract. The recognition of a three-way correlation between magmatic SiO_2 content, critical crystallinity, and the size (magnitude) of crystal fractionation-generated composition gaps in calc-alkaline magmatic systems suggests an important control of magmatic critical crystallinity on the formation of such composition gaps. To explain this correlation, it is proposed that fractionation-generated composition gaps are caused by: (1) simultaneous interior (i.e. non-substrate) crystallization and vigorous chamberwide convection which leads to progressive crystal suspension; (2) cessation of convection when the percentage of suspended crystals reaches the critical crystallinity of the magma, and; (3) eventual buoyancy-driven crystal-liquid segregation producing a discrete body of fractionated magma which is separated from the initial magma by a composition gap. This mechanism implies that many, if not most magma bodies are characterized by interior crystallization and vigorous convection, conditions which are not universally agreed upon at present. Given that such conditions characterize natural magma bodies, fractional crystallization through crystal settling in low-velocity boundary layers should be an important mechanism of fractional crystallization. In a crystallizing and convecting body of magma, composition gap formation should represent one endmember of a complete spectrum of possible evolutionary paths governed by the relative rates of crystal settling and crystal retention. As a given volcanic plumbing system matures with time, average settling/retention ratios within individual magma bodies should increase due to higher average wall-rock temperatures. It follows that, within a given volcanic center, early-stage volcanism should be more likely to display fractionation-generated composition gaps than later-stage volcanism. Such a temporal evolution has been documented at at least two Aleutian calc-alkaline volcanic centers.

Introduction

To date, numerous studies of calc-alkaline volcanic centers have documented the combined importance of fractional crystallization, crustal assimilation and/or

melting, and magma mixing in the low pressure evolution of basaltic magmas (e.g. Luhr and Carmichael 1980; Grove et al. 1982; Kay and Kay 1985). Recently, the emphasis has shifted towards understanding the physical mechanisms through which these various magmatic processes operate. Much of this work has been of a theoretical or experimental nature (e.g. Huppert et al. 1986; McBirney et al. 1985; Marsh 1988) but important evidence also lies in the rocks themselves. The primary goal of this paper is to use evidence of this sort to constrain the physical process(es) involved in low-pressure fractional crystallization. Though the analysis is specific to calc-alkaline magmatism, the results could have general application to tholeiitic and alkaline magmatism as well.

A notable feature of calc-alkaline fractional crystallization is the formation of magmas that span a continuous range of compositions. Because this phenomenon has been documented at so many volcanic centers (e.g. Perfit et al. 1980; Hildreth 1983; Luhr and Carmichael 1985; Brophy 1987, 1989, 1990; Singer et al. 1991), it would appear to be a fundamental property of calc-alkaline fractional crystallization. At the same time, however, it is also known that fractional crystallization (among other processes) can produce composition gaps that disrupt an otherwise continuous range of lava and/or pyroclastic compositions (e.g. Weaver et al. 1979; Hildreth 1983; Grove and Donnelly-Nolan 1986; Bacon and Druitt 1988). Because the formation of composition gaps through fractional crystallization appears to be the exception rather than the rule, understanding their formation may well provide an important key to understanding the fractional crystallization process as a whole.

In this paper, two important correlations that may bear on this matter are considered. The first is a relationship between the size, or magnitude of a given composition gap and the SiO_2 content of the lower boundary of the gap. The second is a close agreement between the magnitude of a composition gap and the critical crystallinity (Marsh 1981) of a magma represented, again, by the lower SiO_2 concentration boundary of the gap. Together, they may provide important insights into the formation of fractionation-generated composition gaps and, by inference, fractional crystallization in general.

Composition gaps and their origins

Since the early recognition of composition gaps by Bunsen (1851) and Daly (1925), their presence has been documented in numerous calc-alkaline, alkaline, and tholeiitic eruptive centers. Most workers today would agree that composition gaps are formed by one of two processes. Detailed field and geochemical (trace element and isotopic) studies have shown that many composition gaps form when hot, mafic magma intrudes cooler crustal rocks and partially melts them to produce compositionally distinct silicic magmas (e.g. Holmes 1931; Bailey 1974; Sigurdsson 1977; Myers and Marsh 1981; Christiansen 1984). At the same time, similar studies have sometimes shown that other composition gaps are formed during the fractional crystallization process with little or no input of crustal material (e.g. Baker et al. 1977; Weaver et al. 1979; Grove and Donnelly-Nolan 1986; MacDonald 1987; Bacon and Druitt 1988). It is this second type of composition gap that is the subject of this paper.

Several different controls on the formation of fractionation-generated composition gaps have been proposed. These include:

- (1) a physical control, wherein a density or viscosity maximum within a single liquid line of descent inhibits the ascent and eruption of a range of magma compositions (e.g. Baker et al. 1977; Jones 1979)
- (2) a phase equilibrium control, wherein the appearance of a new liquidus mineral phase causes a fractionating magma to pass through a certain composition interval more rapidly than others, thus reducing the probability that magmas within that composition range are erupted onto the surface (e.g. Weaver 1977; Clague 1978; Grove and Donnelly-Nolan 1986)
- (3) sidewall crystallization and liquid fractionation, wherein crystallization along the walls of a magma chamber yields a thin boundary layer of fractionated silicic liquid which, in alkaline and calc-alkaline systems at least, is more buoyant than the parental magma and may rise and collect at the top of the chamber (Chen and Turner 1980; McBirney et al. 1985), and
- (4) closed system fractional crystallization of quiescent mafic magma with eventual release of evolved interstitial liquid once the crystalline residue forms an interconnecting network that is subject to fracturing in response to stress (Thompson 1972).

It is not the purpose of the present study critically to evaluate each of these proposals but, rather, to suggest an alternative mechanism through which fractionation-generated composition gaps may form.

Fractionation-generated composition gaps

This section summarizes a literature-based compilation of calc-alkaline volcanic centers that display composition gaps of known or suspected fractional crystallization origin. Because fractional crystallization in the *complete* ab-

sence of other differentiation processes is unlikely, the term "fractionation-generated" is used here to refer to those systems in which fractional crystallization is believed to be the dominant differentiation process. Details of the compilation process are summarized below.

Because crustal melting and silicic magma generation is more likely to occur in volcanic arcs underlain by thick, siliceous continental crust, the first step was to restrict the literature search to volcanic arcs constructed on oceanic or young continental (transitional) crust where the likelihood of significant crustal melting should be reduced. The second step was to identify volcanic centers within these arcs that displayed composition gaps. This stage of the search posed some difficulties. First, nearly all calc-alkaline volcanic centers record very complex histories of fractional crystallization (with or without crustal melting/assimilation), episodic magma chamber replenishment and associated magma mixing (e.g. Eichelberger 1980; Luhr and Carmichael; 1980, 1990; Sakuyama 1981). Due to this complexity, evidence for existing composition gaps could be obscured and easily overlooked (e.g. Brophy 1987). More to the point, in those cases where composition gaps are not obscured, the scale at which they are preserved may range from a single lava flow or pyroclastic deposit to an entire volcanic center. In either case, detailed stratigraphic studies are required to assess accurately the presence of such composition gaps (e.g. Harris 1963) and the number of centers that have been described in such detail is small. Because of this problem, volcanic centers that contain composition gaps were compiled into two groups. Group I centers were those that displayed well-documented composition gaps based on detailed stratigraphic and geochemical studies. Group II centers were those that lacked detailed stratigraphic information but appeared to display composition gaps based on compilations of published analyses of lava and/or pyroclastic deposits (this study).

Within each of these groups, the final step was to select only those centers where the available geochemical data favored fractional crystallization as the dominant differentiation process during composition gap formation. Dissimilar isotopic signatures in rocks on either side of a composition gap would argue against a fractional crystallization origin. In addition, because fractional crystallization concentrates all incompatible trace elements at essentially the same rate, the lack of identical (or very similar) incompatible trace element ratios on both sides of a gap would also argue against a fractionation origin. Following this reasoning, only those centers that displayed similar isotopic signatures and/or incompatible trace element ratios across the gap were included in the final compilation.

Figure 1 summarizes the final list which includes 7 Group I and 19 Group II centers. The list is undoubtedly incomplete and probably includes some centers where fractional crystallization is not the dominant differentiation process. However, it is felt that the list is representative of the range of volcanic centers that display fractionation-generated composition gaps and is thus useful as a basis for further investigation.

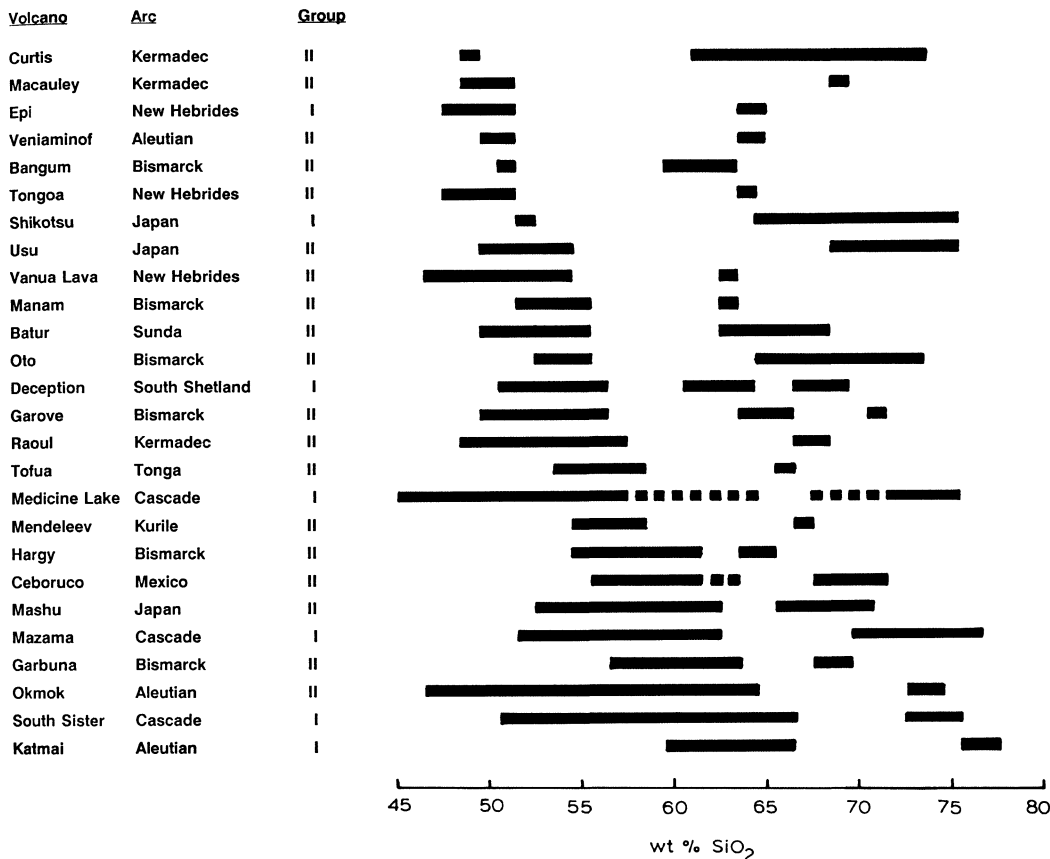


Fig. 1. A compilation of individual orogenic volcanic centers that display one or more crystal fractionation-generated composition gaps. All composition gaps are believed to be real (see text). Group I composition gaps are defined on the basis of detailed stratigraphic and geochemical studies. Group II composition gaps are defined on the basis of compilations of published lava and/or pyroclastic analyses. The solid bars indicate the reported ranges of SiO_2 contents of erupted lavas and/or pyroclastic deposits. Dashed lines indicate mixed lavas and/or pyroclastic deposits. Volcanic centers are arranged in order of increasing SiO_2 content of the lower composition gap boundary. Sources of data are: Curtis – Ewart et al. 1977; Smith et al. 1988; Macauley – Ewart et al. 1977; Epi – Gorton 1977; Crawford et al. 1987; Veniaminof – Kay et al. 1982; Bangum

– Johnson and Chappell 1979; Tongoa – Gorton 1977; Dupuy et al. 1982; Shikotsu – Katsui 1963; Katsui et al. 1978a; Usu – Katsui et al. 1978a, b; Oba 1966; Vanua Lava – Barsdell et al. 1982; Manam – Morgan 1966; Johnson et al. 1985; Batur – Wheller and Varne 1987; Oto – Blake and Ewart 1974; Deception – Weaver et al. 1979; Garove – Johnson and Arculus 1978; Raoul – Brothers and Searle 1970; Ewart et al. 1977; Tofua – Bauer 1970; Medicine Lake – Grove and Donnelly-Nolan 1986; Mendeleev – Bailey et al. 1989; Hargy – Johnson and Chappell 1979; Ceboruco – Nelson 1980; Mashu – Katsui et al. 1975; Katsui et al. 1978a; Mazama – Bacon and Druitt 1988; Garbuna – Johnson and Chappell 1979; Okmok – Byers 1961; Kay 1977; South Sister – Clark 1983; Katmai – Hildreth 1983

Composition gap magnitude and critical crystallinity

Composition gap magnitude

Fractionation-generated composition gap magnitude (composition gap magnitude for short) is defined as the percentage of crystallization necessary to span an observed composition gap assuming perfect fractional crystallization and an initial magma represented by the lower boundary of the gap. For a given composition gap, its magnitude can be estimated from the degree of enrichment of a highly incompatible trace element across the gap (Clague 1978) using the relationship

$$M = 100(1 - (C_0/C_f)) \quad (1)$$

where M is the magnitude and C_0 and C_f are the observed incompatible trace element concentrations on the low-silica and high-silica sides of the gap respectively.

With few exceptions, all of the centers in Fig. 1 are devoid of lavas or pyroclastics that contain sanidine, hornblende, biotite, or zircon, or are theoretically saturated with zircon at appropriate eruption temperatures (Watson and Harrison 1983). Thus K, Rb, Zr, U and Th were chosen as appropriate incompatible elements with which to estimate composition gap magnitude.

Using Eq. 1, Table 1 lists one or more single element-based estimates of composition gap magnitude for each Group I and Group II composition gap shown in Fig. 1. Also listed for each gap is an average estimate which, in some cases, may exclude individual estimates that differ significantly from those based on other elements. Because none of the trace elements used in these calculations can ever display perfect incompatibility (i.e. $D^{\text{sol/liq}} = 0.0$), all estimates of composition gap magnitude should be considered as minimum values.

In Fig. 2, the average estimated magnitudes of both Group I and Group II composition gaps are plotted

Table 1. Single element-based estimates of composition gap magnitude (wt% fractional crystallization needed to produce the observed gap)

Volcano	SiO ₂ *	M _K	M _{Rb}	M _{Zr}	M _U	M _{Th}	M _{Avg}
Curtis	49.62	67	70	—	—	—	69
Macauley	50.13	66	—	73 ⁺	61	65	64
Epi	50.60	70	71	70	—	—	70
Veniaminof	50.80	56	62	—	—	—	59
Bangum	51.00	67	65	62	—	—	65
Tongoa	51.17	64	68	60	67	60	64
Shikotsu	51.99	55	—	—	—	—	55
Usu	53.31	50	—	—	—	—	50
Vanua Lava	54.37	54	56	45 ⁺	—	—	55
Manam	54.2	48	—	—	—	—	48
Batur	55.0	—	—	36	—	—	36
Oto	55.41	56 ⁺	64	63	62	62	63
Deception	56.0	52 ⁺	44	44	—	—	44
Garove	56.0	59	—	51	53	56	55
Raoul	56.54	51	—	51	49	46	49
Tofua	57.16	40	—	—	—	—	40
Medicine Lake	57.75	—	—	—	—	—	59 ⁺⁺
Mendeleev	58.5	44	53	—	—	—	49
Hargy	60.42	31	28	29	—	—	29
Ceboruco	61.55	35	40	30	36	41	34
Mashu	61.0	41	—	—	—	—	41
Mazama (High-Sr)	61.6	51	50	42 ⁺	—	53	51
Garbuna	62.8	31	24	21	—	—	25
Okmok	63.9	37	—	—	—	—	37
Deception	64.0	19	24	17	—	—	20
Garove	65.5	23	—	20	18	19	20
South Sister	65.5	30	32	—	—	—	31
Katmai	66.0	46	59 ⁺	11 ⁺	41	43	43

References same as in Fig. 1

* SiO₂ content of the composition gap lower boundary

+ Excluded from the average

++ Estimate based on major element mass balance calculation

against their initial magma SiO₂ contents as deduced from the lower boundary of each gap. When portrayed in this fashion, both Group I and Group II centers display a pronounced negative correlation between initial magma SiO₂ content and composition gap magnitude, where the latter decreases systematically from around 65–70% for an initial basalt magma (50% SiO₂), to around 20–40% for an initial dacite magma (66% SiO₂). It is important to note that both the Group I and Group II centers define similar trends. This supports the notion that, despite the lack of detailed background stratigraphic studies, Group II centers represent valid examples of fractionation-generated composition gaps.

Critical crystallinity

An important rheological property of crystal-rich magma is an abrupt, several orders of magnitude change in bulk viscosity over a very narrow range of crystal contents. In a crystallizing body of magma, this transition occurs well before the magma is completely solid and marks the point at which a fluid magma takes on the rheological properties of a solid (e.g. Sakuma 1953; Arzi 1978; McBirney and Murase 1984). Critical crystallinity, as defined by Marsh (1981), is the percentage of

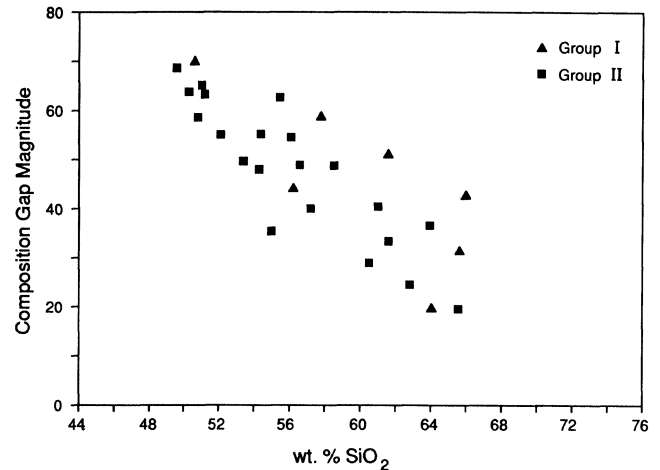


Fig. 2. Composition gap magnitude versus initial magma SiO₂ content for each of the composition gaps shown in Fig. 1. Composition gap magnitude is defined as the percentage of fractional crystallization necessary to span a composition gap. Initial magma is equivalent to the composition of the lava or pyroclastic deposit that defines the lower boundary of a composition gap. Note that both Group I and Group II composition gaps define the same trend

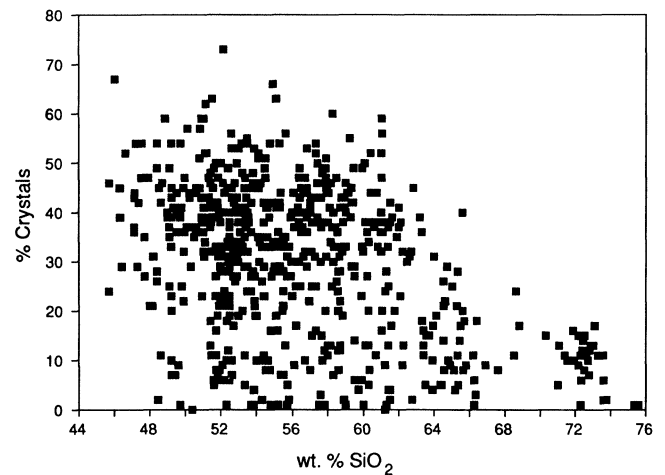


Fig. 3. Reported total phenocryst content versus bulk rock SiO₂ content for >600 orogenic lava flows and domes (ash flows are excluded). The data are taken from the literature. At a given SiO₂ content, the upper limit of phenocryst content is a measure of the critical crystallinity of a magma of corresponding SiO₂ content

crystals present in a crystallizing magma when this important transition takes place.

Of particular importance to the present investigation is the presence of a compositional (i.e. SiO₂) control on critical crystallinity values in calc-alkaline magmas. Following the lead of Marsh (1981), it is assumed that if critical crystallinity marks the point at which a magma effectively becomes a solid, magmas which have crystallized beyond this limit are too viscous to erupt as lavas (though explosive eruption and formation of pyroclastic deposits is not prohibited). In this case, the maximum observed phenocryst content in lavas of a given composition should provide a direct field measurement of the critical crystallinity of the corresponding magma composition.

This technique has been used presently to derive a general critical crystallinity-magmatic SiO_2 relationship for calc-alkaline magmas. In Fig. 3, reported phenocryst contents for over 600 calc-alkaline lava flows and domes (ash-flows are excluded) from thirteen active volcanic arcs are shown plotted against whole-rock SiO_2 content. The data display a well-defined but somewhat diffuse upper crystallinity limit which, in agreement with the findings of Marsh (1981), decreases systematically with increasing SiO_2 . The diffuse nature of the upper limit is most likely due to inter-arc and intra-arc variations in pre-eruptive magmatic conditions, particularly volatile content. From Fig. 3, it can be deduced that typical critical crystallinity values for calc-alkaline magmas decrease from around 65–70% for a basalt at 50% SiO_2 , to around 15–20% for a rhyolite at 70% SiO_2 .

Critical crystallinity and composition gap magnitude

In Fig. 4, the data from Fig. 2 showing the relationship between initial magma SiO_2 content and composition gap magnitude have been superimposed on the phenocryst content data of Fig. 3. Over the entire range of initial magma SiO_2 contents, the composition gap magnitude estimates are in close agreement with the upper boundary of the phenocryst content data. If it is recalled that this upper boundary may also be a measure of critical crystallinity, Fig. 4 indicates a correspondence between the critical crystallinity of a given magma and the amount of crystallization necessary to span a composition gap that originates from it. This correspondence is all the more remarkable when it is considered that the two data sets are completely independent of one another. This result suggests a control of magmatic critical crystallinity on the formation of composition gaps.

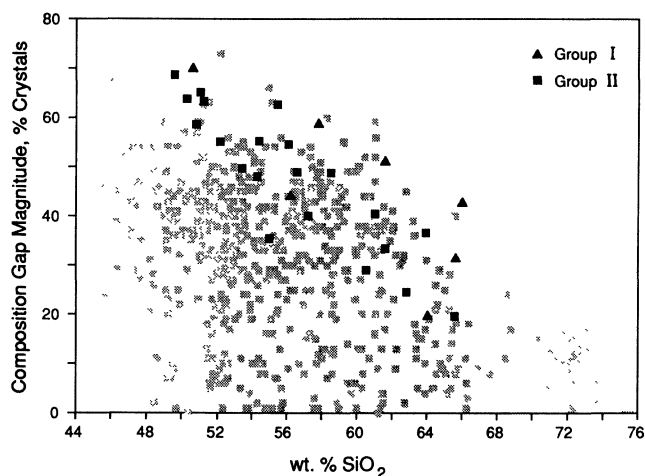


Fig. 4. Composition gap magnitude – initial magma SiO_2 content data of Fig. 2 (solid symbols) superimposed on the total phenocryst – bulk lava SiO_2 content data of Fig. 3 (light grey squares). Over the entire range of initial magma SiO_2 contents, the composition gap magnitude data coincide with the upper limit of reported phenocryst contents. Thus, for an initial magma of arbitrary SiO_2 content, the magnitude of the associated composition gap is equivalent to the critical crystallinity of the initial magma

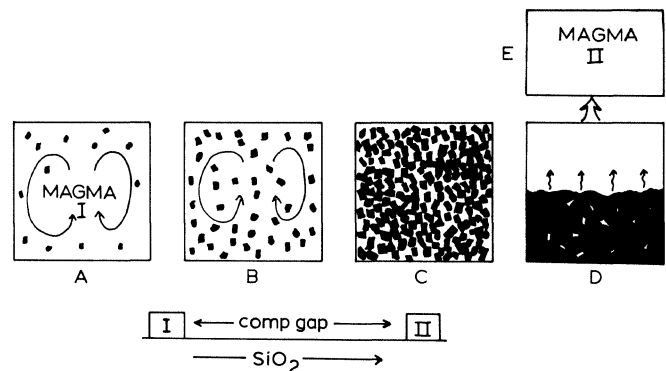


Fig. 5. Schematic diagram outlining a proposed model of crystal fractionation and composition gap formation (see text for discussion)

A possible mechanism of composition gap formation

Figure 5 diagrammatically outlines a suggested process of fractional crystallization and delayed crystal-liquid segregation which could explain the correlation between critical crystallinity and composition gap magnitude demonstrated in Fig. 4. Specific aspects of the suggested process are discussed below.

Initially, a magma of arbitrary SiO_2 content undergoes simultaneous crystallization and vigorous convection, causing all crystals to remain suspended within the moving fluid (Fig. 5A). During this time the liquid composition evolves rapidly but, because the crystals and liquid remain well-stirred, the magma composition remains the same. Over time, the percentage of suspended crystals increases towards the critical crystallinity limit for that particular magma composition (Fig. 5B). When the critical crystallinity limit is reached, the crystal-rich magma undergoes a rheological transition from a fluid to a solid and ceases to convect (Fig. 5C). Once convection has stopped, buoyancy-driven crystal-liquid segregation occurs and the fractionated liquid collects as a discrete magma body that is compositionally separated from the original magma by a composition gap (Fig. 5D). Because the magnitude of the gap is equivalent to the critical crystallinity of the original magma, the correlation between critical crystallinity and composition gap magnitude seen in Fig. 4 is explained.

For this suggestion to have any validity requires that: (1) natural magma bodies undergo vigorous convection; (2) simultaneous crystallization and convection can lead to crystal suspension; (3) convection continues until critical crystallinity is achieved; (4) crystal suspension continues until convection stops and, finally; (5) buoyancy-driven crystal-liquid segregation occurs after the cessation of convection. All of these points are considered below.

Convective Velocity and crystal retention

Estimates of the nature and velocity of magmatic convection vary widely (e.g. Sparks et al. 1984; Clark et al. 1987; Marsh 1989) but the consensus is one of unsteady,

turbulent convection at velocities well in excess of typical crystal settling velocities (e.g. Sparks et al. 1984; Martin et al. 1987). Under these conditions, Martin and Nokes (1989) have shown that crystal suspension can be widespread throughout the interior of the convecting body but that crystal settling can occur in the region immediately above the chamber floor. If the rate of crystal loss through settling is great, a steady increase in the percentage of suspended crystals is unlikely. However, if crystal nucleation and growth rates exceed the rate of crystal loss through settling, the result could be a continuous increase in the percentage of suspended crystals.

The relative rates of crystal nucleation and growth and crystal settling are largely controlled by the rate of heat loss to the surrounding country rock. This, in turn, is governed by such factors as initial wall-rock temperature, magma body size and geometry, and the presence or absence of partial wall-rock melting and/or hydrothermal circulation in the surrounding country rock. All of these factors are highly variable and poorly constrained, but the possibility exists that under certain circumstances progressive crystal suspension and eventual crystal congestion can occur. Given this possibility, the remaining requirements are now considered.

Convective Velocity and critical crystallinity

Whether or not convection continues until critical crystallinity is reached depends primarily on the effect of suspended crystals on bulk magma viscosity. Magmatic convective velocity, W , is usually estimated from a relationship of the general form

$$W = a(K_T/h)(Ra)^b \quad (2)$$

where a and b are experimentally or theoretically derived constants, h is the height of the magma chamber, K_T is thermal diffusivity, and Ra is the thermal Rayleigh number (e.g. Martin and Nokes 1989). The convective velocity of crystal-rich magma, W_c , can be approximated by

$$W_c = W_0(\eta_0/\eta)^b \quad (3)$$

(see Appendix, Eq. A4) where W_0 is the convective velocity of the crystal-free magma, η_0 and η are the bulk viscosity of the crystal-free and crystal-rich magma respectively, and b is the same constant as in Eq. 2. In essence, Eq. 3 reflects the effect of increasing suspended crystal content on Ra and, therefore, convective velocity.

Figure 6 shows the variation in convective velocity with increasing crystal content as calculated from Eqs. 2 and 3 for the case of a 1 km thick body of hydrous (1% H₂O) olivine tholeiite basalt magma with an assumed critical crystallinity value of 65% (Fig. 2). Additional assumptions are $a=0.06$ and $b=0.667$ (Martin and Nokes 1989), $K_T=8 \times 10^7$ m²/s and $Ra=10^{12}$ for the initial, crystal free basalt magma (Clark et al. 1987), and constant ΔT between the magma and wallrock. The results indicate that with increasing crystallization the convective velocity is initially only slightly effected. However, when the critical crystallinity value is approached, convective velocity drops rapidly towards zero. This sug-

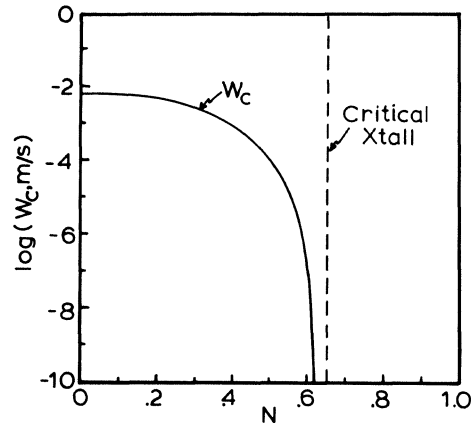


Fig. 6. Convective velocity (W_c , m/s) versus the volume fraction of suspended crystals, N , for a 1 km thick body of hydrous olivine tholeiite basalt magma. The convective velocity variation was calculated from Eqs. 2 and 3 assuming a thermal Rayleigh number, Ra , of 10^{12} for the initial crystal-free magma, and a critical crystallinity of 65% ($N_m=0.65$)

gests that vigorous convection should continue up until that time when the amount of suspended crystals approaches critical crystallinity.

Crystal suspension and critical crystallinity

For crystal suspension to continue until convection stops requires that, regardless of the suspended crystal content, convective velocities always exceed crystal settling velocities. Crystal settling velocities in crystal-rich magma, V_c , can be estimated from the Stokes settling velocity, V_s ,

$$V_s = F \left[\frac{\Delta \rho g a^2}{\eta} \right] \quad (4)$$

and equation 65 of Marsh and Maxey (1985)

$$V_c = V_s \left[\frac{(1-N)}{(1+N^{1/3}) \exp(5N/3(1-(N/N_m)))} \right] \quad (5)$$

where $\Delta \rho$ is the density contrast between crystal and liquid, g is gravitational acceleration, a is the crystal radius, η is liquid viscosity, F is a crystal shape factor coefficient, N is the volume fraction of suspended crystals, and N_m is the critical crystallinity in terms of volume fraction. For a given initial magma composition, Eqs. 4 and 5 can be used to estimate the effect of increasing suspended crystal content on crystal settling velocity provided that mineral and liquid densities and liquid viscosity are known at all stages of the crystallization history.

Grove and Baker (1984) have published a calc-alkaline liquid line of descent model which predicts both mineral and liquid compositions throughout the fractional crystallization history of a parental olivine tholeiite basalt. Assuming an initial 1 wt% H₂O in their starting tholeiite composition, their model was used as a basis for determining mineral and liquid densities and liquid viscosities throughout the entire crystallization history of a typical hydrous calc-alkaline basalt. Anhy-

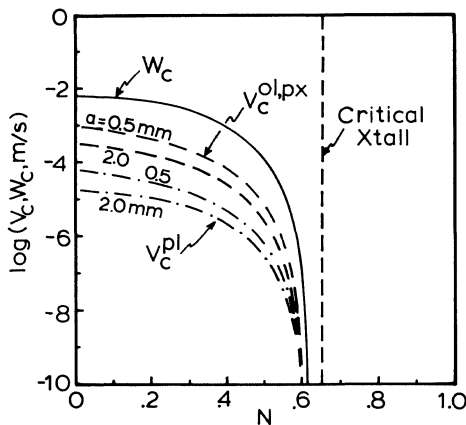


Fig. 7. Plagioclase and mafic (olivine or clinopyroxene) phenocryst settling velocity (V_c , m/s) versus the volume fraction of suspended crystals, N , for a progressively fractionating hydrous olivine tholeiite basalt magma (see text for details). The variations in settling velocities were calculated from Eqs. 4 and 5 assuming spherical crystals, crystal radii that ranged from 0.5 to 2 mm, and a magmatic critical crystallinity of 65% ($N_m = 0.65$). Also shown for comparison is the W_c versus N relationship from Fig. 6. Note that crystal settling velocities are always less than convective velocity

drous liquid densities were calculated at an assumed temperature of 1100° C from Nelson and Carmichael (1979) and then corrected for water content using the H₂O solution model of Burnham (1980) and the H₂O partial molar volume data of Burnham and Davis (1971). One-atmosphere mineral densities were estimated from Deer et al. (1966) and corrected for temperature (1100° C) using thermal expansion data from Skinner (1966). Anhydrous liquid viscosities (1100° C) were calculated from Shaw (1972) and corrected for water content through a correction factor based on the experimental data of Shaw (1963). These data were then combined with Eqs. 4 and 5 to calculate settling velocities of both mafic (olivine/pyroxene) and plagioclase phenocrysts as a function of percent crystallization assuming spherical crystals ($F = 2/9$) and a crystal radii range of 0.5 to 2 mm, the latter being typical of the observed range of maximum phenocryst radii in natural calc-alkaline basalts (Brophy 1987 and unpubl. data).

The results are shown in Fig. 7 along with the calculated variation in convective velocity previously shown in Fig. 6. A comparison of the two indicates that regardless of suspended crystal content, convective velocities always exceed settling velocities by at least one order of magnitude for the mafic minerals and several orders of magnitude for plagioclase. These results suggest that crystal suspension should continue up to the time that critical crystallinity is achieved and convection stops.

Buoyancy-driven crystal-liquid segregation

The last requirement is that crystal-liquid segregation occurs after convection has ceased. Several studies have demonstrated that crystal compaction and liquid expulsion is the likely fate of any crystal-liquid mush in which the crystals initially constitute a self-supporting frame-

work (e.g. McKenzie 1984; Richter and McKenzie 1984, Shirley 1986). Based on the experimental work of Campbell et al. (1978), the minimum percentage of solids that can sustain such a framework can be up to slightly less than 40%, a value that agrees with geochemically based estimates of initial cumulate porosity in gabbroic layered intrusions (Irvine 1980). Thus, any magma composition with a critical crystallinity value greater than 40% could, when convection stops, consist of a self-supporting framework of solids and interstitial liquid. From Fig. 4, this corresponds to magmas with less than 66% SiO₂, a composition range that encompasses the lower boundary of all composition gaps shown in Fig. 3.

Discussion

Fractional crystallization mechanisms

Interior crystallization and gravitational crystal settling has long been considered an important mechanism of fractional crystallization. However, with the recognition that magmatic convective velocities probably exceed typical crystal settling velocities (Sparks et al. 1984), the perceived role of crystal settling as an important fractionation mechanism has steadily lost ground to mechanisms involving in situ crystallization along the floors or walls of a magma chamber accompanied by buoyancy-driven convection of the fractionated liquids away from the stationary solids (i.e. convective fractionation (Rice 1981; Sparks et al. 1984)). Though convective fractionation is undoubtedly an important process, the major shift in emphasis away from interior crystallization has strongly influenced the way that most petrologists view the inner-workings of sub-volcanic magma chambers.

Martin and Nokes (1989) have recently attempted to revive crystal settling as an important fractionation mechanism by demonstrating that *if* interior crystallization and vigorous convection characterize natural magma bodies, crystal settling can occur within the low-velocity boundary layer just above the chamber floor. If the suggested mechanism of composition gap formation is correct, it implies that the required conditions of interior crystallization and vigorous convection do indeed characterize many, if not most crystallizing magma bodies.

If crystallizing bodies of magma are characterized by interior crystallization and vigorous convection then, within the framework of the general crystal settling mechanism outlined by Martin and Nokes (1989), the formation of composition gaps represents just one end-member of a complete spectrum of possible evolutionary paths governed by the relative degrees of crystal settling and crystal retention. If the settling/retention ratio is high, progressive chamberwide differentiation should occur and, over time, the evolving magma should follow a continuous compositional path. At the other end of the spectrum, if the settling/retention ratio is low, chamberwide differentiation should be minimal or absent, and progressive crystal suspension should lead to eventual crystal congestion, crystal-liquid segregation and composition gap formation.

Temporal evolution of fractionation-generated magmas

If crystal settling is an important mechanism of fractional crystallization, then the overall process suggested by Martin and Nokes (1989) and supported by the present study raises the possibility of a temporal evolution of fractionation-generated magma compositions within a given volcanic plumbing system. The settling/retention ratio in a given magma body is governed by the relative rates of crystallization and crystal settling which are controlled largely by the rate of heat loss and, in turn, the temperature contrast between the magma and the enclosing country rocks. Large temperature contrast should favor crystallization rates and, therefore, small settling/retention ratios, while smaller contrasts should have the opposite effect. Because the average temperature of crustal rocks within a long-lived magmatic plumbing system should increase with time (e.g. Singer et al. 1989), it follows that the average settling/retention ratios in magma bodies that traverse the same system should increase as the plumbing system matures over time. Thus, in the earliest stages of plumbing system development, low settling/retention ratios should minimize fractional crystallization through crystal settling and favor progressive crystal retention, crystal congestion, and eventual composition gap formation. With time, as average settling/retention ratios increase, the likelihood of composition gap formation should diminish, fractional crystallization through crystal settling should become more important, and the result should be fractionated magmas that define a continuum in composition space.

Despite the inherent complexity of natural calc-alkaline magmatic plumbing systems, at least two Aleutian calc-alkaline volcanic centers studied by the author (Cold Bay and Kanaga) appear to display the predicted temporal variations in fractionation-generated magma characteristics (Brophy 1987, 1990). Both centers are similar in that they display lava compositions that range continuously from 50 to 62% SiO₂ and contain abundant evidence for combined low-pressure fractional crystallization and magma mixing. In both cases, detailed analysis has permitted the products of fractional crystallization to be viewed in isolation without the obscuring effects of magma mixing. At both centers it was observed that fractional crystallization during the early stages of volcanic activity commonly produce derivative high-silica andesite magmas (~62–64% SiO₂) that were separated from the parental basalt (~50–53% SiO₂) by a significant composition gap. Over time, evidence for the gap diminished and was replaced by fractionation-related magmas that ranged continuously from basalt to andesite. Though only representing two examples, it is possible that similar detailed studies at other volcanic centers might reveal the same evolution in fractionation-related magma characteristics.

Conclusions

In this paper a previously unrecognized three-way correlation between magmatic SiO₂ content, magmatic critical

crystallinity, and fractionation-generated composition gap magnitude has been demonstrated. Based on this correlation, a process of composition gap formation has been proposed wherein combined interior crystallization and vigorous convection leads to crystal suspension, crystal congestion, and eventual crystal-liquid segregation. If correct, it follows that many, if not most crystallizing magma bodies are characterized by interior crystallization and vigorous convection, a conclusion which is not universally accepted at the present time. In addition, it lends support to the general mechanism of fractional crystallization through crystal settling proposed by Martin and Nokes (1989) and reaffirms the role of crystal settling as an important mechanism of fractional crystallization.

Given the general mechanism of Martin and Nokes (1989), composition gap formation would represent one endmember of a complete spectrum of possible evolutionary paths governed by the relative rates of crystal settling and crystal retention in a given body of crystallizing magma. As a magmatic plumbing system passes from immaturity to maturity, the average settling/retention ratio within individual magma bodies should increase due to a progressive rise in the average temperature of the enclosing wall-rock. It follows that the early stages of volcanism within a given volcanic center should be more likely to display fractionation-generated composition gaps than the later stages. Such a progression has been observed at two Aleutian calc-alkaline volcanic centers and could well be present in other centers which have not yet been studied in the same detail.

Appendix

Effects of crystal content on convective velocities

Estimates of convective velocity, W , are typically based on a relationship of the form:

$$W = a(K_T/h)(Ra)^b \quad (A1)$$

where a and b are constants, K_T is thermal diffusivity, and h is a characteristic length scale for the body of magma. Ra is the thermal Rayleigh number where

$$Ra = \frac{\alpha \rho g \Delta T h^3}{\eta K_T} \quad (A2)$$

and α is thermal expansion, ρ is density, ΔT is the temperature difference across the fluid, g is gravitational acceleration, and η is viscosity. If Eq. A1 remains valid for crystal-rich magma (the effects of strongly non-Newtonian properties of crystal-rich magma are unclear), then for constant h and ΔT , the convective velocity of crystal-rich magma (W_c) relative to an initial crystal-free magma (W_0) is given by

$$\frac{W_c}{W_0} = \frac{K_T}{(K_T)_0} \left[\frac{\alpha \rho (K_T)_0}{\alpha_0 \rho_0 (K_T)} \right]^b \left[\frac{\eta_0}{\eta} \right]^b \quad (A3)$$

where the non-subscripted parameters are weighted according to the relative proportions of liquid and solid. For complete solidification, the order of magnitude changes of the first two terms (both less than 10^{-2} , Clark 1966; Murase and McBirney 1973) are much less than that of the second term (10^{12} , Sakuma 1953), so Eq. A3

can be approximated by

$$\frac{W_c}{W_0} = \left[\frac{\eta_0}{\eta} \right]^b \quad (\text{A4})$$

The variation of (η_0/η) is estimated from equation 66 of Marsh and Maxey (1985):

$$\frac{\eta_0}{\eta} = \frac{1}{\exp \left[\frac{2.5N}{1 - (N/N_m)} \right]} \quad (\text{A5})$$

where N is the volume fraction of crystals present and N_m is the critical crystallinity expressed in terms of volume fraction.

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