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A practical petrological method for the determination of volume proportions of magma chamber refilling

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ABSTRACT

A method is described for determining the proportion of the volume of a magma chamber refilled by new pulses of hot primitive melts. This method is based on the changes is crystal contents and in phenocryst zoning resulting from temperature equilibration. Study of phenocryst zoning and the composition of interstitial glasses from various chilled magmatic inclusions and mafic bands in hybrid rocks supports the idea that equilibration of temperatures between two magmas is reached before their physical mixing. If the magmas are genetically related (for example magma-fractionate and initial hot melt), this leads to equilibration of the compositions of residual melts and phenocrysts rims. Intervals of reverse zoning in minerals, resulting from temperature increases, and intervals of normal zoning, caused by cooling, along with the changes in modal mineral contents enable estimation of the amounts of released and adsorbed heat and, therefore, assessment of the proportion of the volume of each magma. Application of the method to some hybrid volcanic series of the Kurile–Kamchatka island arc shows that the proportion of mafic magma intruded into a siliceous magma chamber reservoir never exceeds 10% of the volume of the host magma, based upon thermal considerations. However, the observed degree of compositional mingling requres 25–50% of the mafic end-member. It is shown that this contradiction is related to flotation of vesiculated magmatic inclusions towards the chamber roof, where they become concentrated near the exit and are further mingled with the host during the eruption (forced convection).

Introduction

Periodical refilling of magma chambers by new pulses of initially hot, primitive melts is now accepted by geologists as a common event in the life of many magma chambers. This process forms the basis of the theory of open magmatic systems (McBirney, 1980; Marsh, 1987). Refilling is thought to be responsible for many processes, such as magma mixing (Turner and Campbell, 1986), the triggering of volcanic eruptions (Anderson, 1976; Sparks et al., 1977) and the formation of layered intrusions (Campbell et al., 1983; Naldrett et al., 1990).

Geological evidence about the volume of new magma influxes in layered intrusions of the Kola Peninsula (Russia) suggests that the proportion of new magma does not exceed 2–4% of the total chamber volume (Sharkov, 1983). Evidence for multiple injections of primitive melts is also found in the cumulate layers of the Lukkullaysvaara pluton in Kareliya (Russia), where the volume of injected magma is 1–3% (E.V. Koptev-Dvornikov, pers. commun., 1991). The volumes of the new pulses of magma cannot be estimated directly when this magma undergoes any degree of mixing with the host magma, or for chambers at great depths (volcanic and plutonic).

In this paper a practical petrological method for the determination of the proportion volume of the total supplied by the new influx in a magma chamber (hereafter called the volume proportion) is proposed. This technique is based on the size and composition of the intervals of reverse and normal zoning in minerals, which result from thermal and chemical

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equilibration of two magmas, and the estimation of changes in the mineral contents with in them. This method is applied to some hybrid volcanic series of the Kurile-Kamchatka island arc to provide information about the mechanism of magma mingling. Chilled mafic magmatic inclusions and their host rocks, representing the two thermally equilibrated species of, originally hot, mafic magma in siliceous and intermediate hosts (Sparks et al., 1977; Eichelberger, 1980; Bacon, 1986; Davidson et al., 1990), are used for calibrating the method proposed.

The first aim of this article is to show that the volume proportion of a magma chamber refilled by a new influx of hot primitive melt in volcanic environments does not exceed 3–10% of the original volume of the host magma.

The other purpose is to stress the fact that the volume proportion of magma refilling a magma chamber does not correspond directly to the mixing proportion. A range of process for concentrating new pulses of magma at a particular density level in the magma chamber can be involved. This results in an increasing degree of hybridism at one particular level of the chamber or only in the channel.

Mineral zoning and thermal equilibrium

Thermal equilibrium between magmas is reached faster than compositional mixing (Eichelberger, 1980; Sparks and Marshall, 1986; Frost and Lindsay, 1988) because the thermal diffusivity coefficient is 4–5 orders of magnitude higher than the chemical diffusivity (Shaw, 1974). This means that, over short time scales, two volumes of hot and cool magmas behave as chemically, but not thermally, closed systems. Hot magma undergoes crystallization, whereas pre-existing minerals in the initially cool magma melt (Fig. 1). If the interacting magmas belong to one trend on the multicomponent *T-X* diagram (i.e. are genetically related), the temperature equilibration is followed by chemical equilibration of their interstitial melts and mineral compositions (Fig. 2). In nature, temperature equilibration leads to the formation of reverse zoning in phenocrysts in the initially cool melt and normal zoning in the minerals in the originally hot one. Melting of minerals in the cool melt and crystallization of them in the hot melt results in a difference in crystal contents (Fig. 3).

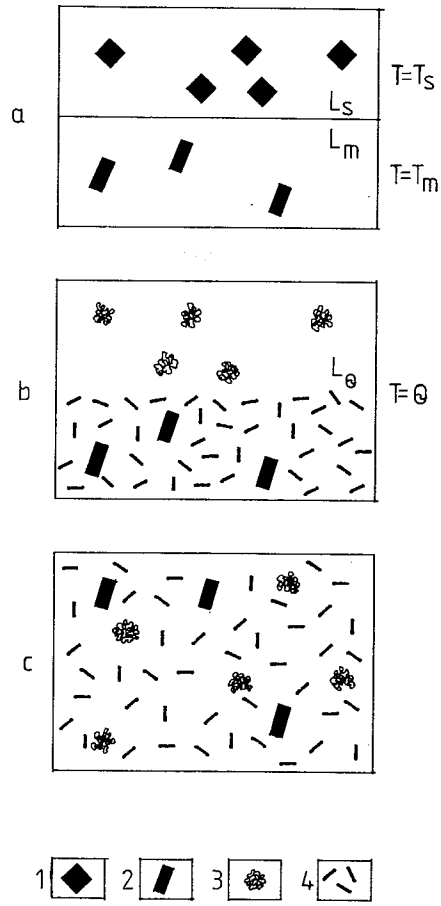

Fig. 1. Results of temperature equilibrations between two contacting magmas with initially different temperatures (T_m and T_s) and compositions of interstitial liquids (L_m and L_s); (a) before temperature equilibration; (b) after temperature equilibration; θ=final equilibrium temperature, L_θ=final composition of common equilibrium melt; (c) after mingling (mechanical homogenization). *1*=phenocrysts in the originally cool (siliceous) magma - L_s; *2*=phenocrysts in the originally hot (mafic) magma - L_m; resorbed, partly dissolved phenocrysts of the initially cool magma originated because of heating; *4*=crystals (microliths) precipitated in the initially hot magma in the course of cooling.

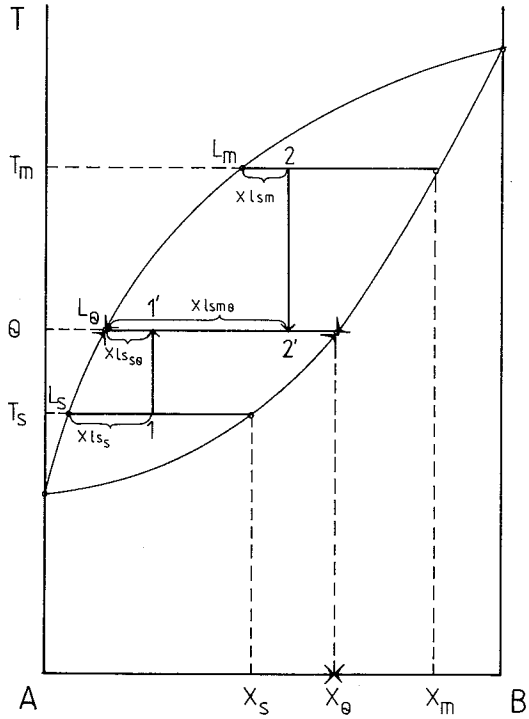

Fig. 2. A simple binary diagram showing results of temperature equilibration between two genetically related magmas, 1 = siliceous and 2 = mafic. T_m and T_s are initial temperatures; θ is the final common equilibrium temperature of the magmas; X_m, X_s and I_m, are initial compositions of coexisting minerals and residual melts; X_θ and L_θ are the final common compositions of mineral and melt; Xls_m and Xls_s are initial proportions of the minerals; $Xls_{m\theta}$ and $Xls_{s\theta}$ are equilibrium crystals contents in two thermally equilibrated magmas; X_s-X_θ and X_m-X_θ are possible intervals of reverse and normal zoning.

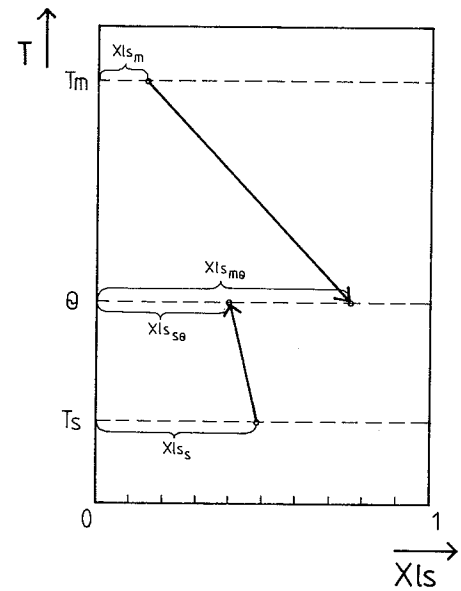

Fig. 3. Scheme showing changes in crystal contents in two contacting magmas after temperature equilibration.

Irreversible effects

Because of the large differences between heat and mass transfer rates, they do not proceed in equilibrium. Thermal equilibrium is affected by three processes: thermal diffusion, crystallization (including nucleation), and crystal dissolution. Transport coefficients for each of these processes differ greatly. The rate of crystallization increases with an increase in the degree of supercooling (Kirkpatrik, 1981). The rate of crystal dissolution is the slowest process, being a totally diffusion controlled process, weakly dependent on the degree of superheating (Zhang et al., 1989).

Pre-existing phenocrysts in the originally cool melt may have insufficient time to dissolve to be in equilibrium with the new value of θ (all variables are defined in Table 1). A possible example of this is plagioclase with "dusty" zones of incomplete dissolution (Tsuchuyama, 1985). The presence of strongly zoned crystals (from originally hot magma) and partially melted crystals, often with reverse zoning, within one rock are also petrographic evidence for these types of processes.

Microprobe study of mineral zoning in magmatic inclusions and in their host rocks shows close similarities between phenocryst rims and interstitial glass compositions (Linneman and Myers, 1990; Davidson et al., 1990; Bindeman, 1991). This implies that local equilibrium has been reached, which makes it possible to estimate the final equilibrium temperature using different equilibrium mineral-melt thermometers or thermometers for coexisting zones of phenocrysts.

TABLE 1

Symbols, parameters and their values

Symbol		Value and/or dimension
C_i	heat capacity of mineral i	kJ/mol
C_m	integral heat capacity of mafic magma	1.5 kJ/mol
C_s	integral heat capacity of silicic magma	0.7 kJ/mol
C^l	heat capacity of the mafic melt	kJ/mol
C_s^l	heat capacity of the silicic melt	kJ/mol
ΔH_i	heat of fusion of mineral i	kJ/g
ΔH_m	integral heat of fusion of mafic magma	400 kJ/g
ΔH_s	integral heat of fusion of silicic magma	300 kJ/g
H_m	heat released through crystallization of mafic magma	kJ
H_s	heat absorbed by the silicic magma	kJ
ρ_m	density of mafic magma	2.7 g/cm^3
ρ_s	density of siliceous magma	2.3 g/cm^3
V_m	volume of mafic magma	cm^3
V_s	volume of siliceous magma	cm^3
M_m	mass of mafic magma	g
M_s	mass of silicic magma	g
T_m	initial temperature of the mafic magma	°C
T_s	initial temperature of the silicic magma	°C
θ	final equilibrium temperature	°C
Xls_i	content of crystal i	
Xls_{im}	content of crystal i at $T=T_m$	
$\mathrm{Xls}_{m\theta}$	content of crystal i at $T=\theta$	
a_i	$=(\mathrm{Xls}_{im}\text{-}\mathrm{Xls}_{m\theta})/(T_m\text{-}\theta)$	grad^{-1}
A	$=(K\text{-}\Sigma\mathrm{Xls}_i)/(T_m\text{-}\theta)$	grad^{-1}
K	total crystal content at $T=\theta$	
p	proportion of end-member which is necessary for producing observable hybrid rock	
%Xls	volume proportion of crystals originally crystallized in each magma	

The integral value of heat capacities and heat of fusions are given after Huppert and Sparks (1988).

Description of the method

Consider two volumes of magma (for example silicic and mafic), with initially different temperature, composition and crystal content. The heat absorbed by the cool, silicic magma (H_s) and the heat released through crystallization of hot, mafic magma (H_m) in the course of temperature equilibration can be given as a sum of three integrals (see Table 1 for parameters):

$$H_m = M_m\left[\int_{\theta}^{T_m} \Sigma(\mathrm{Xls}_i C_i)\mathrm{d}T + \int_{\theta}^{T_m} (1-\Sigma(\mathrm{Xls}_i)C_m)\right]\mathrm{d}T + \int_{\theta}^{T_m} \Sigma\left(\frac{\mathrm{dXls}}{\mathrm{d}T} H_i\right)\mathrm{d}T \quad (1)$$

$$H_s = M_s\left[\int_{T_s}^{\theta} \Sigma(\mathrm{Xls}_i C_i)\mathrm{d}T + \int_{T_s}^{\theta} (1-\Sigma(\mathrm{Xls}_i)C_s)\right]\mathrm{d}T + \int_{T_s}^{\theta} \Sigma\left(\frac{\mathrm{dXls}}{\mathrm{d}T} H_i\right)\mathrm{d}T \quad (2)$$

The first integral reflects the heat released or

absorbed due to heating (1) or cooling (2) of the crystals in magmas, the second reflects the same process for the residual melt of both magmas; the third is the heat released or absorbed due to crystallization (1) or melting (2) of minerals in each magma. Practically, we can estimate only the initial and the final equilibrium temperatures and crystal contents, whereas the real dependence of Xls with temperature is unknown. Therefore, each of these integrals can be given as a difference between the initial (T_m and T_s) and the final (θ) states. If this is true, the only way to calculate these integrals is the assumption of linear dependence of Xls with temperature (Fig. 3). Therefore, for mafic magma:

$$\mathrm{Xls}_i = a_i T + \mathrm{Xls}_{im} \tag{3}$$

$$K = \sum \mathrm{Xls}_i = AT + \sum \mathrm{Xls}_{im} \tag{4}$$

$$1 - \sum \mathrm{Xls}_i = 1 - K = 1 - AT - \sum \mathrm{Xls}_{im} \tag{5}$$

$$\left(\frac{\mathrm{dXls}}{\mathrm{d}T}\right)_i = a_i \tag{6}$$

$$\sum\left(\frac{\mathrm{dXls}}{\mathrm{d}T}\right) = \left(\frac{d\sum Xls}{dT}\right)_i = A \tag{7}$$

The coefficients a and A may be estimated if the initial T_m, T_s and the final temperature θ are known as $a = (\mathrm{Xls}_{im} - \mathrm{Xls}_{m\theta})/(T_m - \theta)$ $A = (K - \sum \mathrm{Xls}_i)/(T_m - \theta)$ (Fig. 3).

Substitution of Eqs. (3-7) into Eqs. (1,2), assuming constant values of heat capacities and heat of fusions with temperature, gives:

$$H_m = M_m[(T_m - \theta)(\sum 0.5 a_i C_i (T_m + \theta) + \sum C_i \mathrm{Xls}_{im}) + C^l(1 - \sum 0.5 A (T_m + \theta) - \sum \mathrm{Xls}_{im}) + \sum a_i H_i] \tag{8}$$

$$H_s = M_s[(\theta - T_s)(\sum 0.5 a_i C_i (T_s + \theta) + \sum C_i \mathrm{Xls}_{is}) + C^l(1 - \sum 0.5 A (T_s + \theta) - \sum \mathrm{Xls}_{is}) + \sum a_i H_i] \tag{9}$$

If necessary, the dependence of C and H on temperature is given in the review of Frost and Lindsay (1988).

Because the convective heat loss from the hot mafic magma to the siliceous one is much more effective than the conductive heat loss from the chamber walls, the heat released through crystallization proves to be equal to the heat absorbed by cool magma:

$$H_m = H_s \tag{10}$$

and after substitution (see Table 1):

$$V_m/V_s = \frac{\rho_s[(\theta - T_s)(\sum 0.5 a_i C_i (T_s + \theta) + \sum C_i \mathrm{Xls}_{is})}{\rho_m[(T_m - \theta)(\sum 0.5 a_i C_i (T_m + \theta) + \sum C_i \mathrm{Xls}_{im})} + \frac{C^l(1 - \sum 0.5 A (T_s + \theta) - \sum \mathrm{Xls}_{is}) + \sum a_i H_i]}{C^l(1 - \sum 0.5 A (T_m + \theta) - \sum \mathrm{Xls}_{im}) + \sum a_i H_i]} \tag{11}$$

The last formula can be used for the determination of the volume proportion of magmas after temperature equilibration.

Therefore, to establish the ratio V_m/V_s it is first necessary to establish the temperature changes T_m-θ, θ-T_s and to estimate the initial and the final crystal contents for each magma at θ.

T_m-θ, θ-T_s can be determined using one of the following methods:

(1) mineralogical thermometers for cores and rims of phenocrysts;

(2) mineral-melt thermometers for pairs: crystal–bulk groundmass composition (for end-member) and rim of crystal–residual glass (for thermally equilibrated rocks);

(3) simple binary diagrams for different mineral solid-solutions (see Fig. 1, for example).

Each method has its specific shortcomings. Determination of temperature differences using mineralogical and liquidus thermometry is strongly dependent up on the correct determination of coexisting compositions and compositions of end-members. The last approach is the most reliable in practice, because the ranges of crystal zoning determined by microprobe may be easily inverted into temperature

differences. Moreover, the simultaneous use of simple binary diagrams for plagioclase, pyroxenes, olivine, etc., would greatly improve the final result. It should be taken into account that the values of temperature differences are in the nominator and denominator respectively in eqs. (11,13). This implies that the relative error should be strongly reduced.

To determine the changes in crystal contents of each mineral and therefore, the values of a_i and A, we have to assess first the initial Xls_i and final content $\mathrm{Xls}_{i\theta}$ of the crystals. In the case where information about end-members is available and we can observe the contacting, thermally equilibrated, magma (for example magmatic inclusions and host rock), there is no difficulty in determining changes in crystal contents in thin sections. Let us consider the case in which magma was homogenized (mingled) after temperature equilibration (Fig. 1c). First, using chemical mass-balance calculations we estimate the proportion of each end-member (p), which is necessary for producing observable hybrid rock, assuming end-member compositions are known. If they do not, one may establish the compositions of end-members using the technique, described by Langmuir et al. (1978) and Popov et al. (1991). Then the volume proportions of crystals (%Xls) originally crystallized in each magma (with reverse and normal zoning) can be calculated separately. Finally, using the simple equation:

$$\mathrm{Xls} = \%\mathrm{Xls}/p \tag{12}$$

we can estimate the volume proportion of crystals in each thermally equilibrated magma (Fig. 1b) as if they had not been mingled after temperature equilibration (Fig. 1c).

Assuming that the error in the temperature determination and the error in the calculation of changes in mineral contents seem to parallel the errors in C and H for various minerals and melt compositions (Frost and Lindsay, 1988), we can use the simplified formula similar to Sparks and Marshall (1986):

$$\frac{V_\mathrm{m}}{V_\mathrm{s}} = \frac{\rho_\mathrm{s}[(\theta - T_\mathrm{s})C_\mathrm{s} + (\mathrm{Xls}_\mathrm{s} - \mathrm{Xls}_{\theta\mathrm{s}})H_\mathrm{s}s]}{\rho_\mathrm{m}[(T_\mathrm{m} - \theta)C_\mathrm{m} + (\mathrm{Xls}_{\theta\mathrm{m}} - \mathrm{Xls}_\mathrm{m})H_\mathrm{m}]} \tag{13}$$

where (Xls_s-$\mathrm{Xls}_{\theta\mathrm{s}}$) and ($\mathrm{Xls}_{\theta\mathrm{m}}$-$\mathrm{Xls}_\mathrm{m}$) are integral values of changes in crystal amounts for siliceous and mafic magmas, respectively.

Procedure and practical application of the method

Three hybrid volcanic series of the Kurile-Kamchatka island arc were used for the determination of $V_\mathrm{m}/V_\mathrm{s}$ using eq. (13). Non-equilibrium phenocryst assemblages, strong compositional heterogeneities of volcanics and abundances of chilled mafic inclusions are characteristic of these volcanic rocks. Its origin is consistent with the model of refilling of the silicic magma chamber by a new pulse (or pulses) of hot mafic magma (Bindeman, 1992a,b; Bindeman and Bailey, 1992). It is important to note that explosions of silicic magma at the beginning stage of hybrid eruptions of Baranskiy and Mendeleev volcanoes (Table 2) show no non-equilibrium mineral assemblages (Bindeman, 1992a,b). They can be considered as a host magma, pushed out by the new pulse of hot mafic magma before mixing began (Blake, 1981). Products of these explosions are considered to be silicic end-members.

At Dikii Greben' Volcano, the initial rhyodacites have neither non-equilibrium phenocryst assemblages nor magmatic inclusions. Dacites and andesites are hybrid and contain both magmatic inclusions and non-equilibrium phenocryst assemblages. Therefore, initial rhyodacites can be considered as a silicic end-member.

Magmatic inclusions in the eruptive products of these volcanoes cannot be considered directly as mafic end-members. They contain phenocrysts of quartz and An-poor plagioclase captured from the silicic host rock. To estab-

TABLE 2

Volume proportion of mafic magma refilled the siliceous magma chamber.

Volcano	End-members	Crystal content (%)		Ranges of phenocryst zoning			Estimation of temperature differences, method		
		initial	final	Pl (%An)	Opx (%Fs)	Cpx (%Fs)	Pl	Opx	Pl-melt
Mendeleev (Kuriles)	rhyodacite	45	~43	40–50	40–37	–	50	15	80
	basaltic andesite (inclusion)	5–7	70	80–50	27–37	16–29	170	50	358
V_m/V_s (%)							6.8	4.3	6.6
Dikii Greben' (Kamchatka)	rhyodacite	35	32	30–45	35–31	–	70	25	65
	basalt (inclusion)	3–5	60	75–45	23–31	15–20	180	40	378
V_m/V_s (%)							10.2	8.7	7.1
Baranskii (Kuriles)	dacite	25	24	40–47	n.d.	n.d.	45	–	39
	basaltic andesite (inclusion)	15	70	74–50	n.d.	n.d.	165	–	203
V_m/V_s (%)							6		4.9

Temperature estimation: Pl,Opx - using plagioclase and orthopyroxene binary diagrams (Chukhrov, 1974); Pl-melt after Kudo and Weill (1970)
n.d. = no appropriate data: crystals are weakly zoned.

lish the initial state of mafic magma (end-member) at the moment of refilling we excluded captured phenocrysts from quantitative mineral calculation.

Intervals of reverse and normal zoning of minerals were established on different grains, both in inclusions and in host rocks. Mineral, glass and bulk groundmass compositions were determined using "CamScan" electron microscope with Link AN 10.000 microprobe analyser at the laboratory of the Moscow State University.

The composition of crystal rims in magmatic inclusions and in host rocks, as well as the composition of residual glasses are practically the same (Table 2), which is in good agreement with Figure 2. These compositions were considered to be in equilibrium for θ. Cores of phenocrysts nearly always exhibit only minor zoning; only the narrow rims are strongly zoned. Such a relationship, reflecting high rates of temperature equilibration, has been observed elsewhere (Sakuyama, 1981; Nixon, 1988).

For determining T_m-θ and θ-T_s, using data on mineral zoning, we used two approaches (Table 2):

(1) Plagioclase-melt thermometer (Kudo and Weill, 1970) for pairs: crystal–bulk groundmass composition (for end-members) and the rim of crystal–residual glass (for thermally equilibrated rocks and inclusions);

(2) Simple binary diagrams for plagioclase and orthopyroxene (Chukhrov 1974).

The second step is the determination of changes in crystal contents ($Xls_{\theta m}$-Xls_m) and (Xls_s-$Xls_{\theta s}$). For mafic end-members (magmatic inclusions), the total amount of microl-

ites can be considered to be produced during rapid cooling in the course of temperature equilibration and therefore approximates ($Xls_{\theta m}$-Xls_m). The small increase in volume of phenocrysts due to narrow, strongly zoned rims was also taken into account. Phenocrysts interpreted as having been captured from the cold magma were excluded.

For the silicic end-member, the proportion of phenocrysts was estimated on products erupted before mixing began. As shown in Table 2, the volumetric decrease in crystals in the silicic end-member after temperature equilibration is very small (1–5 vol. %). It is explained by the fact that plagioclase of "dusty" type and partially decomposed amphibole preserve their original volume. Decreased volume fractions of phenocrysts are mainly due to the dissolution of quartz, pyroxenes and opaque minerals, and the shrinking of completely decomposed amphibole and biotite in the course of its dehydration.

Proportion of magma chamber refilling

As we can see from Table 2, in all cases the proportion of the injected hot mafic magma does not exceed 10% of the resident silicic magma (note that the influx of mafic magma constitutes 4–10% of the resident silicic magma but not of the entire chamber volume). As shown by mass balance calculation (Bindeman, 1992a,b), the silicic magma, itself, is originated from the fractionation of an initial mafic magma on Baranskiy and Mendeleev volcanoes. The volume of the silicic magma can constitute no more than 30% of the original mafic magma volume after fractionation. Therefore, the actual proportion of the new mafic magma in the entire chamber is approximately 1–3%.

The siliceous end-member shows only a slight decrease in crystal contents in all cases. The degree of crystal dissolution in silicic magma can be emphasized by the dimensionless parameter α, changing from 0 (no dissolution) to 1 (equilibrium dissolution). The parameter α relates to the degree of non-equilibrium dissolution, resulting from the slow rate of dissolution, V_d, relative to high rate of temperature equalization, V_t. If heat transfer is fast, $V_d \ll V_t$: if it is slow $V_d \approx V_t$. Both these regimes may exist in nature. If a new influx of hot mafic magma is then injected forcefully and disrupted in the course of refilling, giving birth to magmatic inclusions (cf. Bacon and Metz, 1984), crystals in the resident magma have insufficient time to dissolve before θ is reached, therefore $\alpha=0$. If mafic magma slowly refills the silicic chamber, ponding on the bottom and slowly cooling as an independent convective layer, crystals of the resident magma may have time to dissolve providing $\alpha \approx 1$.

Substitution of α into eq. (13) yields:

$$\frac{V_m}{V_s}=\frac{\rho_s[(\theta-T_s)C_s+(Xls_s-Xls_{\theta s})H_s\alpha]}{\rho_m[(T_m-\theta)C_m+(Xls_{\theta m}-Xls_m)H_m]} \quad (14)$$

Accepting $T_m=1200°C$ and $T_s=900°C$ and assuming linear variation in Xls_m and Xls_s with temperature: for mafic magma at 1200°C $Xls_m=0$ at 900°C $Xls_s=1$; for silicic magma at 1000°C $Xls_s=0$, at 800°C, we can solve eq. (14) for θ:

$$\theta=\frac{3400V_m/V_s+1350\alpha+630}{2.833V_m/V_s+1.5\alpha+0.7} \quad (15)$$

In Figure 4 the final temperature θ is calculated as a function of α for different values of V_m/V_s.

As we can seen from Figure 4, silicic magma is heated significantly at low values of α. For example, 10% of added mafic magma can increase the temperature of the resident silicic magma by 90°C at $\alpha=0$, but only by 36°C at $\alpha=1$. Therefore, the actual equilibrium temperature should be higher than the equilibrium temperature of the mixture computed from simple theory. If so, even small volume proportions of resident hog magma can give rise to a significant temperature increase in the host siliceous magma.

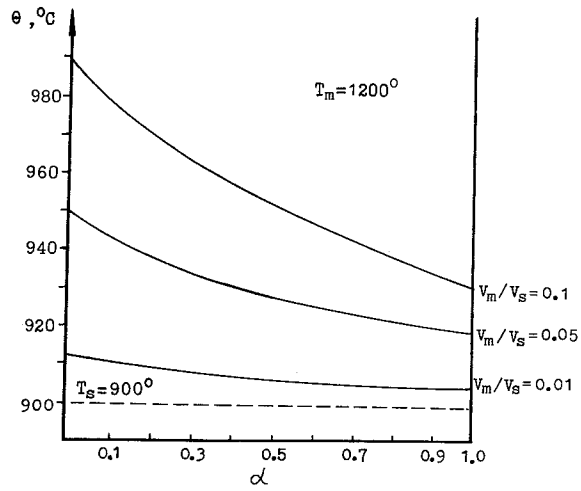

Fig. 4. The final equilibrium temperature θ as a function of the parameter α (degree retention of crystal dissolution) at different initial proportions of mafic magma (see text for further explanation).

Proportion of magma mixing and proportion of magma chamber refilling

The estimated proportion of magma chamber refilling contradicts the calculated proportion of the basic end-member needed to produce the average composition of the hybrid rocks studied. Simple mass-balance calculations, using end-members or even Sr and Nd isotope balance calculation (Davidson et al., 1990) require 20–40 vol.% of added basic magma to produce dacites and andesites from basalts and rhyolites.

This contradiction may be explained in two ways. First, periodic refilling of magma chambers results in a gradually increasing degree of hybridism in the whole chamber. This process can be recognized in the complicated recurrent zoning of plagioclase (cf. Nixon and Pearce, 1987), reflecting the periodic heating and cooling of magma. We observed such relationships in inclusions and their host andesites from the Bezymyanny and Kizimen Volcano, Kamchatka (Frolova et al., 1992). However, plagioclase with recurrent zoning is not typical of the volcanic series reported here. The second mechanism is concentration of mafic magma near the exit, before eruption. This can occur by vesiculation of initially dense mafic magma (Eichelberger, 1980; Huppert et al., 1982), leading to flotation of mafic inclusions and their concentration near the chamber roof (Bindeman, 1992; Bindeman and Podladchikov, 1992). The other possibility is the buoyant rise of the influxing magma to some neutral density level in a density-stratified magma chamber (Campbell et al., 1983). The different mechanisms of concentration due to the dynamics of magma withdrawal may also be involved (Turner and Campbell, 1986).

For the volcanoes studied, the densities of magmatic inclusions measured are 0.1–0.4 g/cm^3 less than their host magmas, due to the strong porosity of inclusions (Bindeman and Podladchikov, 1992), implying the possibility of the flotation of inclusions in the host magma (cf. Eichelberger, 1980).

The size distribution (Fig. 5) of inclusions on these volcanoes shows that it is always log normal, with an elevated volume proportion of the large inclusions. Figure 5 shows a typical distribution plot. This shape of distribution can be interpreted as the result of faster flotation of large inclusions according to Stokes' law and the concentration at the chamber roof before eruption (Bindeman and Podladchikov, 1992). Moreover, changes in the inclusion size distribution for consecutive eruptions of the Dikii Greben' Volcano (Kamchatka) show successive shifts in the distribution toward smaller sizes (Fig. 5b). This is interpreted as representing the gradual flotation of smaller inclusions, with their lower rise velocities. This evidence supports the notion that inclusions are concentrated at the channel roof before eruption.

Direct observations of the concentration of inclusions in the heads of extrusive domes and lava flows are often mentioned (Bacon, 1986; Frolova et al., 1992). In many cases, the volume proportion of magmatic inclusions exceeds 4–10 vol.% which is indicated by the thermal considerations described above (Yer-

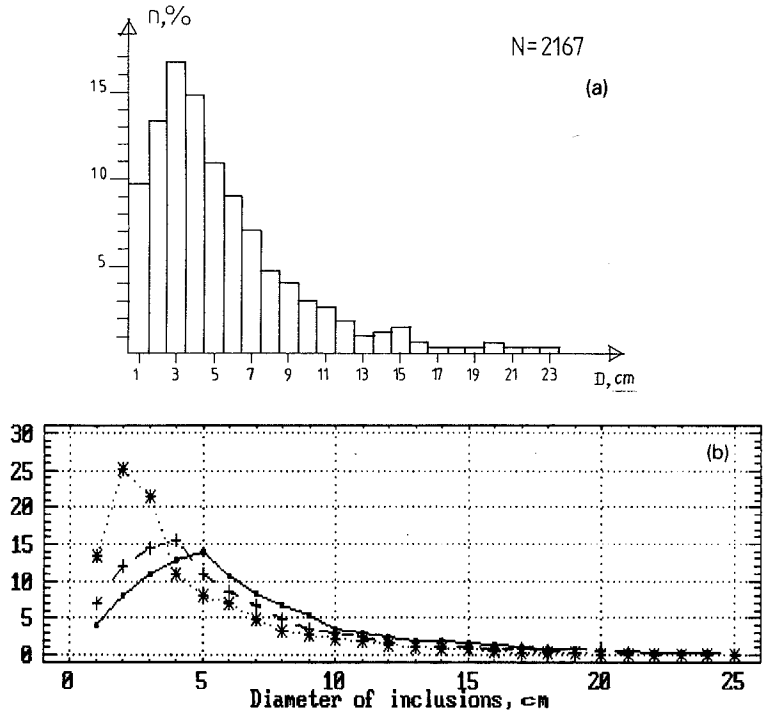

Fig. 5. The character of diameter distribution of magmatic inclusions in extrusive domes and lava flows of the Dikii Greben' Volcano (Kamchatka). (a) Average distribution plot, N is the total number of measurements. (b) Diameter distribution in the series of consecutive eruptions, numbers correspond to the order of eruption.

makov and Koloskov, 1969; Yermakov, 1973; Bindeman and Bailey, 1992). V.A. Yermakov (pers. commun., 1991) discovered that magmatic inclusions comprise up to 30 vol.% in the extrusive dome Lokhmatiy in Benzymianny Volcano (Kamchatka) and about 10–20 vol.% in some extrusive domes of Dzenzur Volcano (Kamchatka). I observed 12% inclusions on some extrusive domes on the Dikii Greben' Volcano and 3–5% inclusions on the extrusive dome of Mendeleev Volcano. These observations lead us to conclused that inclusions concentrate near the chamber roof before eruption. If so, the observable proportion of mingling reflects the proportion in the channel but not the proportion of the entire chamber.

It is interesting to compare the proportion of mingling with the proportion of mafic inclusions. As we can see from Table 2, the volume of inclusions is, in all cases, less than the required proportion of mafic end-member required. Apparently, most inclusions disintegrate in the host magma to small fragments and even microliths. Small fragments of the chilled margins of inclusions are common. Moreover, statistical comparison of microlith compositions in the groundmasses of host rock and in inclusions on Dikii Greben' and Mendeleev Volcanoes show their similarities (Bindeman, 1992; Bindeman and Bailey, 1992). Microliths in dacites and andesites have an unusually basic composition, identical to the inclusions. This proves that microliths were originally formed in mafic inclusions during the chilling and then disintegrated into siliceous host.

Several possible approaches for the mechanisms of magma mingling have been proposed. Concentration of the vesiculated parcels of magma (inclusions) near the chamber roof implies that the mingling model in the eruptive conduit during the course of eruption (forced convection) is the most appropriate for

the volcanoes studied (cf. Kouchi and Sunagawa, 1985; Koyagychi and Blake, 1989).

Proportion of refilling and volcanic eruption

The process of chamber refilling is traditionally considered to be a cause of volcanic eruptions (Anderson, 1976; Sparks et al., 1977; Blake, 1981). The small estimated proportion of chamber refilling corresponds to the calculations performed by Blake (1981), which showed that even 0.1% of influx is enough to exceed the cracking limit of the chamber roof and initiate an eruption. The initial pyroclastic overbursts (described above) of the resident silicic magma before mixing beings are probably connected with this overflowing of the chamber. If so, the relatively small volume of erupted pyroclastic rocks detected on the Mendeleev and Baranskiy volcanoes are roughly comparable with the small volume of new influx injected into chamber.

Flotation of inclusions with gas bubbles can be considered as an important factor leading to an increase in fluid pressure, according to the gas-lifting model (Steinberg et al., 1984; Michaelides, 1989). Therefore, if the new influx is responsible for the initial explosion, the process of its vesiculation (Eichelberger, 1980; Huppert et al., 1982) and the flotation of magmatic inclusions (Bindeman and Podladchikov, 1990) could be responsible for the extrusive eruption which follows.

In the other cases, a significant temperature increase, as a result of non-equilibrium heating (Fig. 4), could cause volume vesiculation of the resident silicic magma (cf. Sparks et al., 1977) and result in powerful explosive volcanic eruptions. Therefore, the flow regime of injected magma affects the character of a forthcoming eruption. If refilling is fast, $\alpha \rightarrow 0$, temperature increases rapidly and vesiculation is significant, causing a relatively powerful eruption; if refilling is slow $\alpha \rightarrow 1$ a relatively weak eruption may follow.

Summary

A practical petrological technique for determining the volume proportion (the proportion of the total volume of magma in a chamber) of magma chamber refilling has been proposed. It is shown that this proportion does not exceed 4–10% of the resident silicic magma, at least in the examples presented. The volume proportion of magma chamber refilling does not correspond directly to the mixing proportion. One of several different mechanisms of concentrating the influxed magma at a neutral density level of the chamber or near the exit may be involved. In the field examples presented here, the mechanism of concentration is connected with the flotation of vesiculated mafic inclusions to the chamber roof, leading to an increase fluid pressure in, volcanic eruption, and mingling in the course of eruption.

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