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The origin and significance of crystal rich inclusions in pumices from two Chilean ignimbrites

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Abstract – Crystal rich (~ 70–98 % phenocrysts) magmatic inclusions in pumices for compositionally heterogeneous ignimbrites from the Central Andes of northern Chile are interpreted as the products of crystal accretion at the sidewalls of the magma chambers. The inclusions are typically andesitic in composition and are found as 'peppery textured' lenses and bands, or as discrete ovoid 'blobs' within dacitic pumices from the early erupted portions of the ignimbrites. The inclusions have a bimodal gain size with large phenocrysts (> 1 mm), typical of those of the host pumice, set in a dominant fine-grained framework (< 0.5 mm) of plagioclase, with lesser amounts of hornblende and biotite in equal proportions, and ubiquitous titanomagnetite in a matrix of vesiculated high-Si rhyolite glass. An igneous microgranular texture is defined by this framework. The mineralogy of the inclusions, as well as the compositions of the phenocrysts and glass, are very similar to those of the host pumices. These characteristics, in addition to available major, trace and REE data, are best reconciled if the inclusions represented samples of fractionated crystals and glass from the same magma as the host pumice. The restricted occurrence of these inclusions in the early erupted portions of the ignimbrites suggests that these crystal accumulations occurred in the upper portions of the magma chambers, at the sidewall; the dominantly fine grain size and crystal rich nature of the inclusions are considered to be the result of the higher thermal gradient in the boundary layer.

These inclusions may be an important link between the experimental and geochemical models for the origin of compositional layering in magma chambers by sidewall crystallization. The presence of similar inclusions in other ignimbrites and volcanics, as well as plutonics, suggest that they may be a common feature of silicic magmas.

1. Introduction

Mafic inclusions of magmatic origin are a common feature of silicic volcanic rocks. The most commonly described are inclusions which result from the mingling of magmas (e.g. Walker & Skelhorn, 1966; Eichelberger, 1975; Bacon & Metz, 1984; Bacon, 1986). These are typically spheroidal, vesicular, have crenulate margins which are convex towards the more silicic host, and show crystal morphologies (quench textures) consistent with incorporation in at least a partially molten state (Bacon & Metz, 1984). Less commonly described are inclusions which have a cumulus origin. These are generally very phenocryst rich and have characteristics suggesting that they are samples of the accumulative crystal mush resulting from fractional crystallization of their host or precursor magmas (e.g. Wörner & Schmincke, 1984; Grove & Donnely-Nolan, 1986; Tait, 1987; Bacon & Druitt, 1988). This report describes such phenocryst-rich inclusions in pumices from two ignimbrites from the 22° 30′ S to 23° S region of the Central Volcanic Zone of northern Chile (Fig. 1).

Guest (1968) first identified these inclusions in pumices from the Cajon ignimbrite – the Purico ignimbrite of this report (after Hawkesworth *et al.*, 1982 and Francis *et al.*, 1984) – and he described them as 'banded pumices' The banding, he noticed, was due to dark crystal-rich bands, which he termed 'andesitic' on account of their bulk chemistry (~ 60 % SiO_2), in a vesicular dacite (~ 65 % SiO_2) pumice host. In contemplating a possible origin for these 'banded pumices' Guest (1968) suggested that they may be a result of either magma mixing, or partial melting and remobilization of material stopped from the walls or roof of the magma chamber. While the present author agrees with many of Guest's petrographic and field observations, it is suggested that there are several inconsistencies with these mechanisms as:

(a) There are no crenulate margins between inclusion and host as found in other typical magmatic inclusions incorporated in a molten state (e.g. Bacon & Metz, 1984).

(b) The euhedral crystal shapes and igneous texture of the inclusions suggests that an origin by partial melting is unlikely.

(c) It is unlikely that a dacite would partially melt andesitic material.

(d) The mineral assemblage of the inclusions, except for the lack of quartz, is very similar to that of the host.

More detailed work on the inclusions from the Purico ignimbrite and similar material from the

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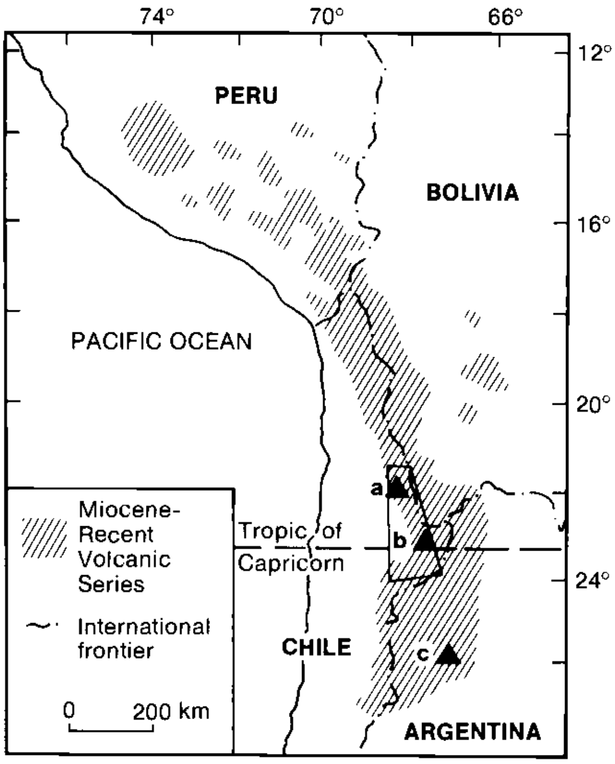


Figure 1. The Central Volcanic Zone of the Andes. The study area is outlined. Also shown are well known volcanic centres. (a) Volcans San Pedro & San Pablo. (b) Cerro Purico. (c) Cerro Galan.

Puripicar ignimbrite in the same region of northern Chile suggests an alternative origin. In this report, the dark bands are interpreted as inclusions of accumulated crystals resulting from crystal fractionation from the same magmas as the host pumice. The accumulations are suggested to result from boundary layer crystallization at the side-walls of magma chambers.

2. Occurrence

The inclusions typically occur as microgranular, holocrystalline, 'salt and pepper' textured streaks, lenses, and bands up to 15 cm long and 4 cm wide, or as discrete ovoid 'blobs' within the host dacite pumice (Fig. 2).

The dominant volume, over 90 %, of the juvenile material from both the Purico and Puripicar ignimbrites is homogeneous high-K dacite (de Silva, unpublished Ph.D thesis, Open University, 1987). However, both ignimbrites also contain small volumes of compositionally distinct pumice. The field and petrologic characteristics of the various pumice populations are consistent with derivation of the ignimbrites from compositionally-layered magma chambers. The compositional heterogeneity of the Purico and Puripicar ignimbrites is described by de Silva (unpublished Ph.D. thesis, Open University, 1987) and is to be discussed in a forthcoming paper. However, in considering the origin of the inclusions it is important to establish at what stage during the eruption they were ejected and therefore a brief description of the ignimbrites is warranted.

Figure 3 shows a stratigraphic section through the Lower Purico ignimbrite. This ignimbrite has a K–Ar age of 1.35 ± 0.15 Ma and an erupted volume of $\sim 70\ km^3$. The earliest erupted material is an airfall of rhyodacitic composition (71 % SiO_2), followed by three major flow units of a homogeneous dacite (~ 65 % SiO_2). In the uppermost flow units, spectacular mixed, andesite–dacite, pumices are found (Francis *et al.*, 1984; de Silva, unpublished Ph.D. thesis, 1987). These are not the 'banded-pumices' described by Guest (1968) but are the result of mingling of the dacite and andesite from the compositionally-layered magma chamber of the Purico ignimbrite (de Silva, unpublished Ph.D. thesis, Open University, 1987) The inclusions described here, and by Guest (1968), occur only in dacitic pumice from the lowest flow units of the Lower Purico ignimbrite (Fig. 3).

The Puripicar ignimbrite (Guest, 1969) has a K–Ar age of 4.18 ± 0.07 Ma and has an estimated erupted volume of at least 250 km^3 within northern Chile alone and an equivalent volume is thought to occur in southwest Bolivia (de Silva, unpublished Ph.D. thesis, Open University, 1987). The ignimbrite contains two compositionally distinct pumice types. The dominant type (90 % by volume) is a higher K, low temperature (673 ± 13 °C), homogeneous, dacite (Type 1) and the other (Type 2) is a lower K, high temperature (740–796 °C), streaky, dacite. De Silva (unpublished Ph.D. thesis, Open University,1987) has shown that the physical, chemical, mineralogical and Sr-isotopic characteristics of these pumices are consistent with eruption of the Puripicar ignimbrite from a compositionally-layered magma chamber which comprised an upper dominant volume of type 1 dacite underlain by type 2. The crystal-rich inclusions from this ignimbrite occur in the dominant type 1 pumice.

3. Petrographic features

The inclusions are typically crystal-rich aggregates consisting predominantly of plagioclase, with lesser amounts of hornblende, and biotite in equal proportions, and ubiquitous titanomagnetite in a matrix of slightly devitrified vesiculated glass; total phenocryst contents (calculated vesicle free) vary from 70 to 98 %. (Table 1). The main petrographic features are;

(i) The inclusions in both the Purico and Puripicar ignimbrites contain no quartz but have an otherwise similar mineralogy to the host pumice. In addition, ilmenite and allanite are not found in the inclusions although they are present in the host pumice of the Puripicar.

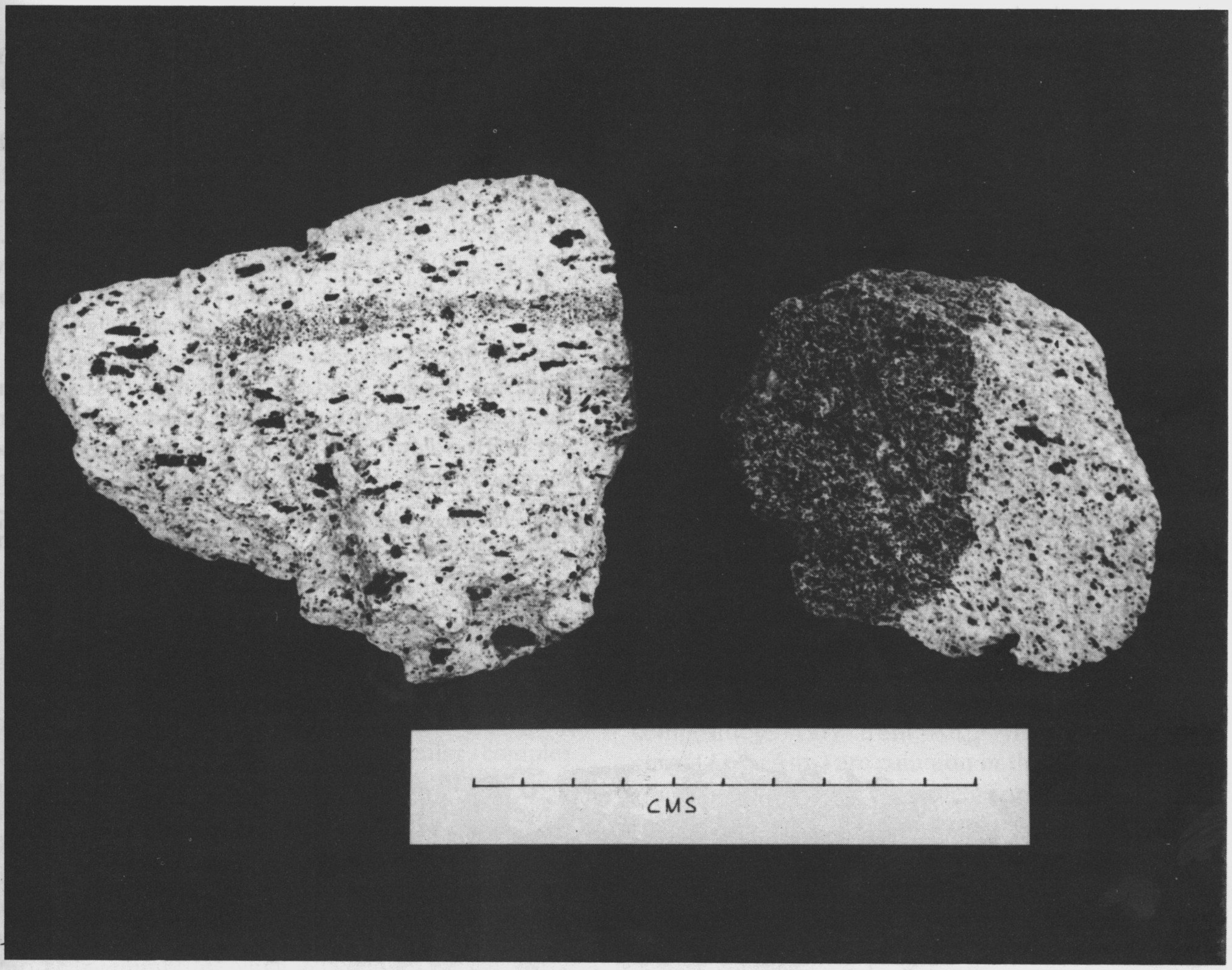


Figure 2. Photograph of hand specimens showing two of the common forms of the crystal-rich inclusions in pumices. Note the sharp contacts and the lack of crenulation which is commonly observed in inclusion formed by magma-mingling (e.g. Bacon & Metz, 1984). The bi-modal grain size of the inclusions as well as their 'peppery' texture is readily apparent. The sample on the left is from the Purico ignimbrite (supplied by Dr. J. E. Guest), the sample on the right is from the Puripicar ignimbrite.

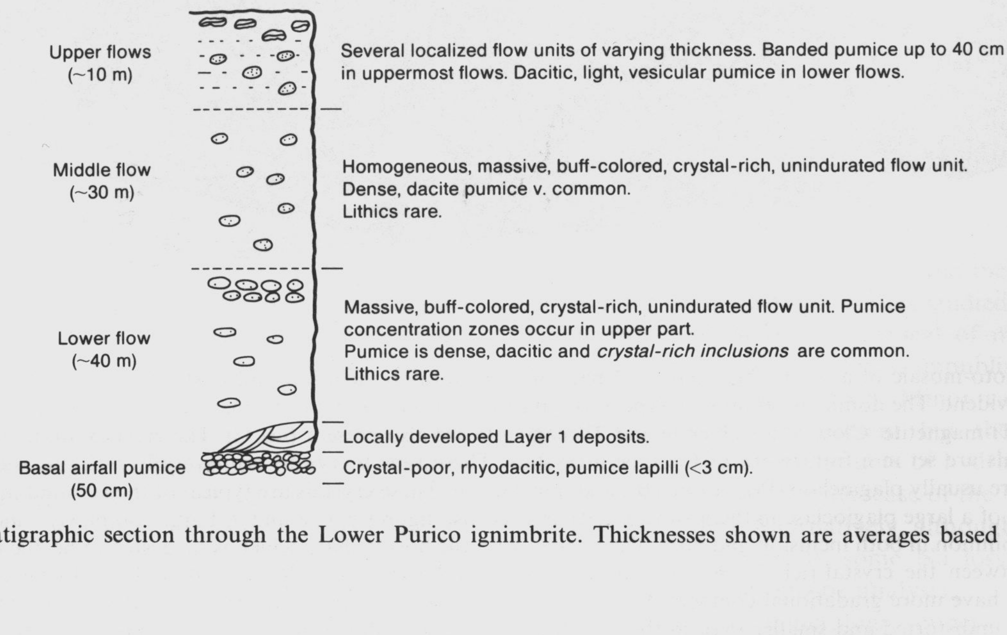


Figure 3. Stratigraphic section through the Lower Purico ignimbrite. Thicknesses shown are averages based on measured sections.

Table 1. Representative modal analyses of inclusion – host pairs from the Purico and Puripicar ignimbrites

	Puripicar						Purico					
	83018		83019		83029		83041		83073		83042	
	H	I	H	I	H	I	H	I	H	I	H	I
Glass	50	27	52	18	50	32	45	39	47	2	47	12
Plagioclase	25	37	25	41	23	31	27	29	26	48	27	46
Quartz	10	n.d.	14	n.d.	11	tr	6	n.d.	6	n.d.	6	tr
Biotite	10	13	6	20	9	16	9	13	9	21	11	18
Hornblende	3	20	2	16	5	16	11	18	10	24	11	21
Fe-Ti oxide	2	4	2	6	2	4	2	1	2	5	2	3
allanite*	tr	n.d.	tr	n.d.	tr	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
apatite*	tr	n.d.	tr	tr	tr	n.d.	tr	n.d.	tr	n.d.	tr	n.d.
zircon*	tr	n.d.	tr	n.d.	tr	n.d.	tr	n.d.	tr	n.d.	n.d.	n.d.
sphene*	tr	n.d.	tr	n.d.	tr	n.d.	n.d.	n.d.	tr	n.d.	tr	n.d.
% crystals**	50	73	48	82	50	68	55	61	53	98	57	88

All modal abundances are calculated on a vesicle free basis. * abundance as discrete microphenocrysts only – does not include occurences as inclusions in other crystals. ** on a vesicle free basis. n.d., not detected; tr, trace amounts; H, host; I, inclusion.

Figure 4. Photo-mosaic of a single thin section of part of a crystal-rich inclusion and host. The bi-modal grain size of the inclusion is evident. The dominant grain size, type a, is typically < 1 mm, and consist mainly of plagioclase with hornblende, biotite, and Ti-magnetite. Clots of hornblende and Ti-magnetite are also present (i.e. C). The crystals are typically euhedral. Larger crystals are set in a framework of the type a crystals. These type b crystals are typically > 2 mm and can be up to 1 cm. They are usually plagioclase (Pl), biotite (B), and hornblende. These crystals are typical of those found in the host: note the remnants of a large plagioclase in the host (top left) and see also figures 5, 6, 8 and 9. Large complexly zoned plagioclases (i.e. Pl) are common in both inclusion and host. The contact with the host is sharp which is suggested to be due to the viscosity difference between the crystal-rich (> 90%) inclusion and the relatively crystal-poor (50%) host. Inclusions with lower crystallinities have more gradational contacts. Vesicles in the inclusions are few and are restricted to 'pockets' of glass, and are generally undistorted and smaller than in the host (i.e. V). The long dimension of the section is ~ 4 cm.

(ii) The total phenocryst contents of the inclusions (vesicle free) are much higher than in the host pumice.

(iii) The inclusions have higher modal *proportions* of mafic phenocrysts than the host: hornblende and titanomagnetite are considerably enriched (Table 1).

(iv) Vesicles in the inclusions are typically less than 0.5 mm in diameter and smaller than in the host, in which the vesicles are typically > 1 mm (Fig. 4). The vesicle content in the inclusions varies sympathetically with the glass content.

(v) The inclusions have a bimodal distribution of crystal sizes which is evident both in hand specimen and in thin-section and two main types of crystals can be defined on the basis of size. The framework of the inclusions is formed by type A crystals, which comprise $\sim 80\%$ (by volume), and are typically 1–2 mm. This framework encloses type B crystals which are typically 5 mm to 1 cm and are typical of those found in the host pumice.

Type A plagioclases are simply twinned euhedral laths of intermediate composition (see below) with little or no zonation while the larger type B plagioclases exhibit more complex optical chemical zoning (Fig. 5), as also noted by Guest (1968). Similar complex plagioclases are found in abundance in the host pumice of both the Puripicar and Purico ignimbrites (Fig. 6). Some of the type B plagioclases have very Ca-rich cores and more intermediate, outer, normally zoned rims (see below). Hornblende in the inclusions, occurs as euhedral crystals and is typically green in the Puripicar ignimbrite, while it is mainly brown in the Purico ignimbrite. These characteristics are consistent with those of the hosts. While some of the type A crystals are zoned, type B hornblendes are generally homogeneous and may contain inclusions of plagioclase and titanomagnetite. This is also seen in the host pumice. Although generally randomly oriented, and even distributed throughout the inclusions, 'clots' of type A hornblende are also found (Fig. 7). Biotite typically occurs as randomly oriented flakes. Once again some of the type A grains are zoned, but type B crystals are homogeneous and inclusion ridden (mainly type A plagioclase, titanomagnetite, apatite and zircon) with pleochroic haloes, like those in the host pumices (Figs. 8 and 9). Titanomagnetite occurs as discrete 0.5 to 1 mm, sub-rounded to euhedral grains as well as inclusions in the main phenocryst phases. Concentrations of these are commonly associated with the hornblende 'clots' (Fig. 7).

The contact between inclusion and host can be sharp or gradational. At the sharp contacts, which appear to be most common with inclusions of crystal contents $> 80\%$, distortion of the vesicles may occur. This is generally more marked in the host than in the inclusion (Fig. 10) as noted also by Guest (1968). Occasionally disaggregation of the inclusion may occur and gradational contacts result where the crystal content decreases into the host while vesicle size, vesicle content, and glass content increase. These gradational contacts are more common where the inclusions have crystal contents $< 75\%$.

4. Mineral and glass chemistry

Representative phenocryst and glass compositions are given in Table 2.

4.a. Plagioclase

Histograms of plagioclase compositions for both the Purico and Puripicar ignimbrites are shown on Figure 11. In the Puripicar ignimbrite, the host pumices contain large normally zoned andesines and Na-labradorites with core compositions which vary from An_{35}–An_{55}, the majority are in the An_{40}–An_{52} range. Rim compositions vary from An_{26}–An_{52}. The type A plagioclases are homogeneous with compositions varying from An_{40} to An_{50}. The type B plagioclases in the inclusions show a similar range of compositions. The cores are typically An_{40}–An_{55}, while the rim compositions vary from An_{25}–An_{50}. Rare Ca-rich cores (An_{71}–An_{75}) are common in the large complexly zoned plagioclases in both inclusions and hosts (Figs. 5 and 6). The zoning pattern in these crystals, in inclusion and host in the Puripicar ignimbrite, is broadly similar. They have an intermediate zone of labradorite (An_{61}–An_{65}) riddled with melt-inclusions and a more sodic outer rim, which is normally zoned in the range An_{43}–An_{51}. A similar case is apparent for the plagioclase in the inclusion and host in the Purico ignimbrite.

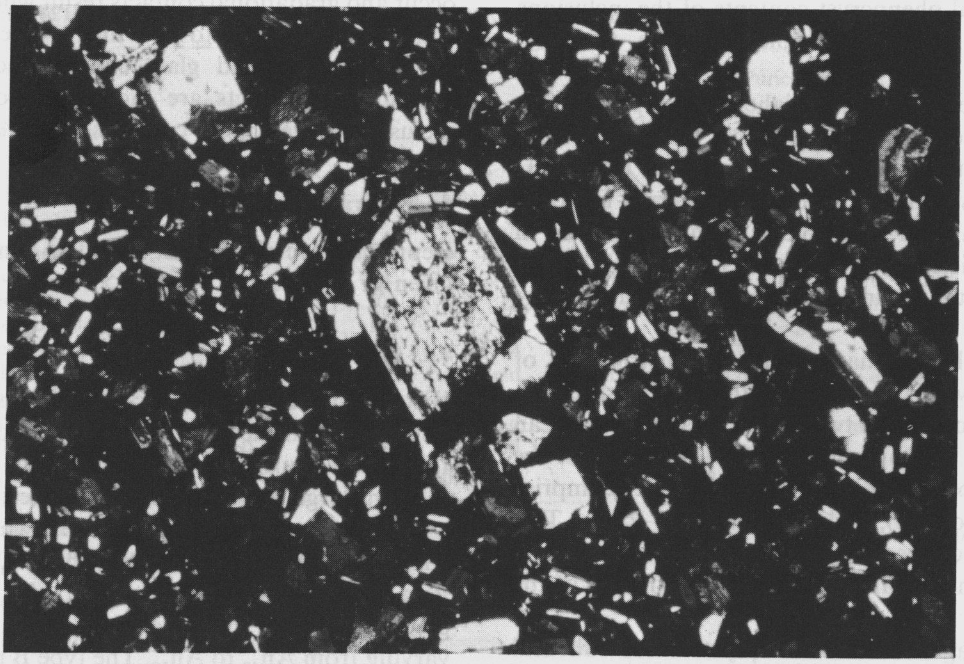
4.b. Fe–Mg minerals

Figure 12 illustrates the chemical characteristics of biotite and hornblende in inclusions and hosts for both the Purico and Puripicar ignimbrites. The compositions of the minerals are broadly similar in both inclusions and host, but the cores of the zoned crystals have slightly higher Mg/Fe ratios.

4.c. Glass compositions

Glass compositions of both hosts and inclusions are very similar in the two ignimbrites studied. The glass is a high-K high-Si rhyolite typical of most of the ignimbrites in this area (de Silva, unpublished Ph.D. thesis, Open University, 1987). Minor variations in the alkali contents are present but these can be accounted for by minor amounts of hydration of the glass which can result in a decrease of the Na/K ratio (i.e. Lipman, 1965). Alternatively, although great care was taken to minimize it, some Na loss may have occurred during microprobe analysis.

The similarity in mineralogy, mineral chemistry,



Figures 5 and 6. Photomicrographs (crossed nicols) of complexly zoned plagioclases from inclusion (Fig. 5) and host (Fig. 6). These phenocrysts have very similar zoning patterns with; a turgid, Ca-rich, bytownite (An 71–71) core, an intermediate zone of labradorite (An 61–65), and a more sodic andesine (An 43–51) outer zone. The outer zone is more typical of the plagioclase compositions found in the pumices and inclusions. Both examples are from the Purico ignimbrite. Figure 6 shows part of a crystal which is about 2 cm long. Field of view in both plates is 3.5 mm.

Table 2. Representative microprobe analyses of phenocryst and glass from inclusions and host pumice from the Purico Puripicar ignimbrites

	Hornblende			Biotite		Glass*	
	h	i	i (c)	h	i	h	i
				Purico			
SiO2	46.51	46.91	47.27	36.70	37.15	78.02	77.65
TiO2	1.56	1.19	1.34	4.76	4.87	0.11	0.12
Al2O3	7.44	7.24	6.74	13.67	13.80	12.61	13.16
FeO	13.63	13.51	13.32	14.56	14.72	0.34	0.37
MnO	0.45	0.47	0.48	0.17	0.17	0.03	0.04
MgO	13.84	13.95	14.29	14.94	15.16	0.12	0.14
CaO	11.55	11.52	11.57	0.04	0.03	0.99	1.02
Na2O	1.30	1.15	1.20	0.53	0.53	2.37	2.19
K2O	0.80	0.75	0.63	8.91	9.14	5.41	5.32
F	—	—	—	0.21	0.21		
Cl	—	—	—	0.15	0.16		
Total	97.08	96.69	96.84	94.64	95.94		
				Puripicar			
SiO2	44.03	41.59	44.68	36.23	35.94	77.71	77.66
TiO2	1.56	2.06	1.71	4.58	4.57	0.09	0.10
Al2O3	8.88	10.68	8.69	13.56	13.46	12.73	12.83
FeO	18.70	19.28	16.77	21.66	21.31	0.29	0.29
MnO	0.48	0.56	0.49	0.31	0.26	0.05	0.03
MgO	9.75	9.00	11.23	10.35	10.49	0.02	0.05
CaO	11.27	11.45	11.42	0.01	0.04	1.07	1.13
Na2O	1.48	1.78	1.56	0.37	0.39	2.67	2.51
K2O	0.99	1.10	0.81	9.04	9.10	5.37	5.49
F	—	—	—	0.00	0.04		
Cl	—	—	—	0.22	0.21		
Total	97.14	97.40	97.36	96.33	95.81		

	Plagioclase								
	h core	int	rim	i-type b core	int	rim	h core	rim	i-type a core
				Purico					
SiO2	49.27	50.34	57.62	49.75	51.14	57.05	56.86	57.46	57.02
Al2O3	31.25	30.28	25.89	30.77	30.40	26.29	25.88	25.90	25.83
FeO	0.32	0.56	0.25	0.32	0.34	0.34	0.31	0.31	0.37
CaO	15.25	14.17	8.97	15.15	13.91	9.52	9.28	8.82	8.69
Na2O	2.84	3.39	6.08	2.99	3.61	5.97	5.98	6.28	6.24
K2O	0.15	0.21	0.57	0.15	0.22	0.54	0.54	0.58	0.55
Total	99.08	98.95	99.38	99.13	99.62	99.71	98.85	99.35	98.70
				Puripicar					
SiO2	49.37	51.54	56.50	49.62	50.79	56.88	56.08	56.95	56.19
Al2O3	31.60	30.43	27.28	31.41	30.37	26.71	26.70	26.34	27.15
FeO	0.27	0.23	0.18	0.27	0.28	0.24	0.23	0.23	0.23
CaO	15.44	13.87	9.91	15.36	13.82	9.26	9.63	8.98	9.75
Na2O	2.82	3.68	5.72	2.85	3.60	6.10	5.89	6.26	5.90
K2O	0.14	0.21	0.46	0.15	0.19	0.46	0.44	0.52	0.45
Total	99.64	99.96	100.05	99.66	99.05	99.65	98.97	99.28	99.67

h, host; i, inclusion; i (c), Mg-rich core from type a crystal; * Glass analyses recalculated to 100 % (anhydrous).

and glass chemistry of host and inclusion suggest that they evolved under very similar magmatic conditions which makes it extremely unlikely that the inclusions could be accidental xenoliths – for instance plutonic xenoliths which crystallized from another magma and have been incorporated into the host. The glass and mineral chemistry of the host and inclusions are therefore suggested to be best reconciled if the crystal-rich inclusions are parts of accumulated residues of a crystal fractionation process which took place in the host magmas (now represented by the host pumices).

5. Whole rock geochemistry

Representative analyses of the host and inclusion pairs are shown in Table 3. The compositions of the analysed inclusions are typically high-K andesite as also reported by Guest (1968). SiO_2 varies from 59.7

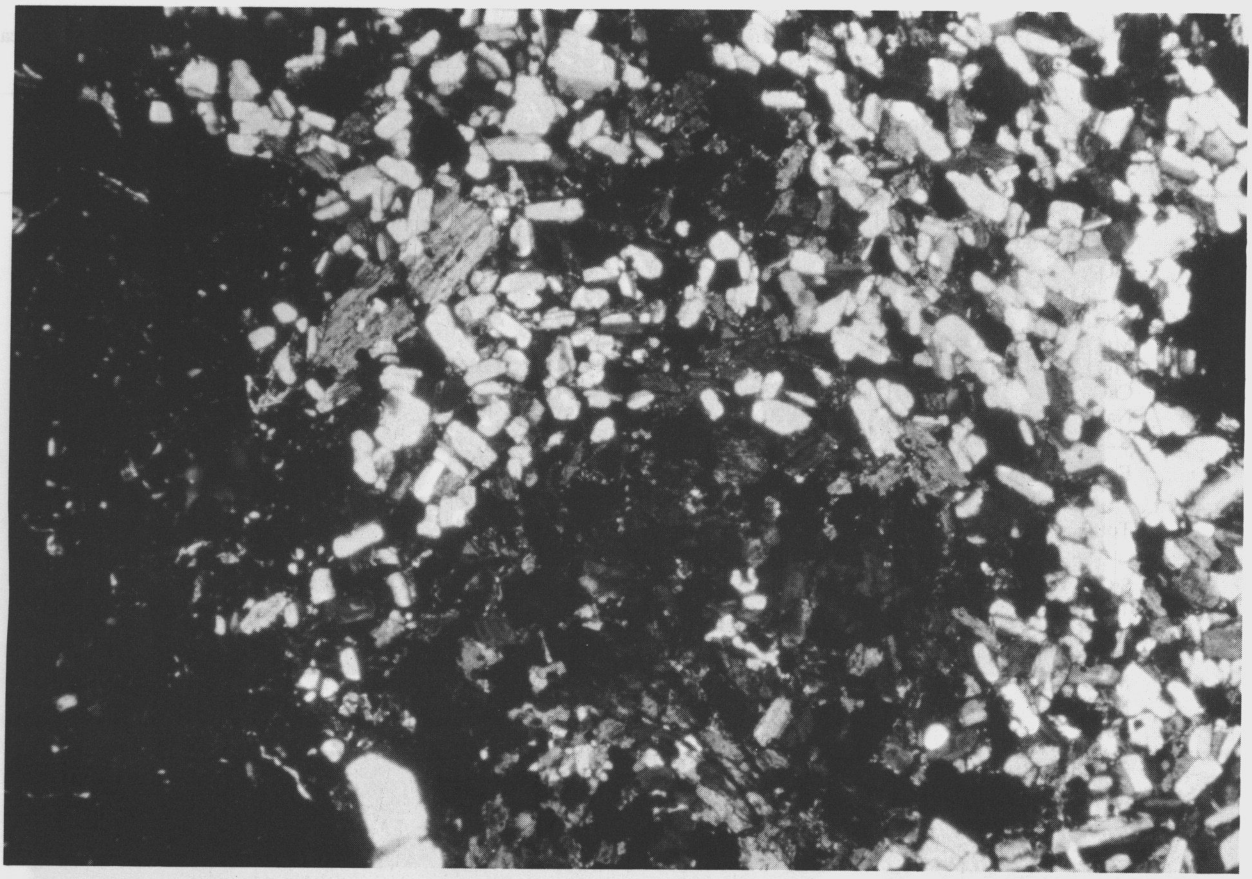


Figure 7. Photomicrograph (crossed nicols) of a typical hornblende 'clot'. Titanomagnetite (black) is the only other mineral present in these aggregations. Field of view 3.5 mm.

Table 3. Representative XRF whole-rock major and trace element analyses of inclusion – host pairs from the Purico and Puripicar ignimbrites

	Purico		Puripicar	
sample	Host 83042B	Inclusion 83042A	Host 83018	Inclusion 83018B
SiO2	65.94	60.59	67.83	59.67
TiO2	0.66	0.95	0.64	0.75
A1203	15.50	16.51	15.34	16.95
FeO	4.76	6.51	3.89	7.04
MnO	0.08	0.10	0.06	0.13
MgO	2.03	2.77	1.53	3.13
CaO	4.27	6.11	3.49	5.87
Na2O	2.87	2.79	3.11	2.93
K2O	3.68	3.41	3.94	3.39
P2O5	0.20	0.26	0.15	0.17
H2O–	1.69	2.02	2.10	2.53
Cr	55	65	45	63
V	76	149	84	166
Sr	324	387	345	400
Y	23	25	21	54
Nb	14	11	14	15
Th	20	13	17	10

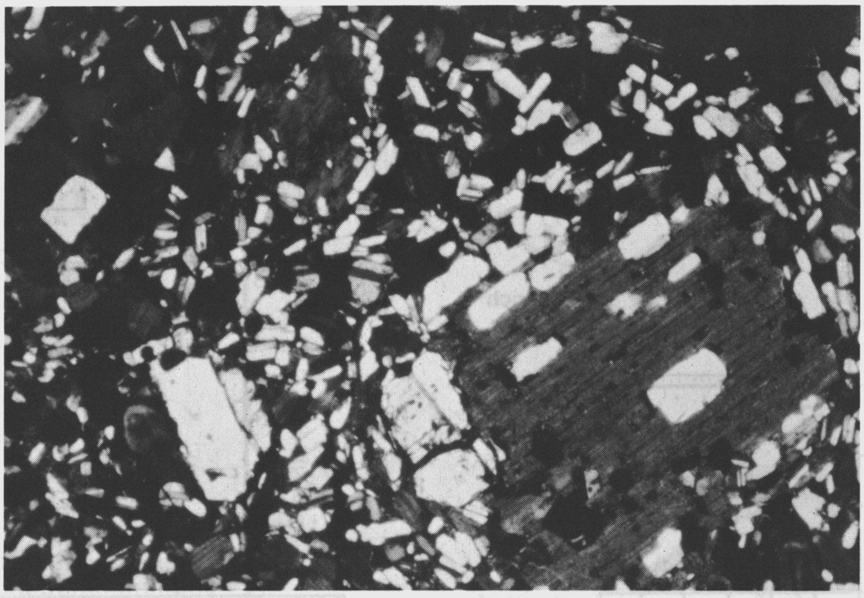
Major elements recalculated to 100% anhydrous.

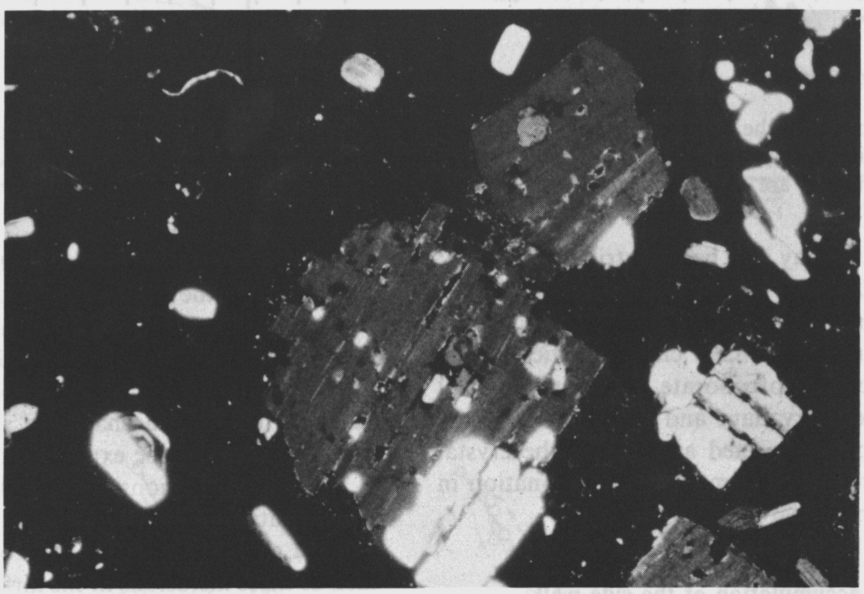
to 61.8%. However, these represent analyses of inclusions with crystal contents between 70 and 80% which is typical of the lower end of the crystallinity of these inclusions. More crystalline inclusions will be more mafic. Three main observations on the chemical characteristics of the inclusions are consistent with a cumulus origin.

(i) On Harker diagrams of the major and selected trace elements (Fig. 13), the inclusions plot at the mafic end of linear trends defined by the host pumices and glasses. These trends are consistent with crystal-liquid separation from the host magma (host pumice): the chemistry of the inclusions representing higher crystal contents than the host.

(ii) The compositions of the host pumices can be modelled simply as a mixture of the inclusions (accumulated crystals) and glass (the residual liquid).

(iii) The REE characteristics of sample 83018b – an inclusion from the Puripicar ignimbrite – show that the inclusion is enriched in MREE and HREE relative to the host pumice (Figs. 14a and b). Nagasawa & Schnetzler (1971) demonstrate that hornblende fractionation is a major control on REE concentrations in acidic magmas and that higher modal abundances of hornblende can lead to increases in absolute abundances of REE (particularly the MREE and HREE). In this light, the higher MREE and HREE contents of the inclusions is suggested to be due to the higher modal contents of hornblende in the inclusions compared to the host (section 3, Table 1). The difference in the LREE contents of the host and inclusion is most likely due to dilution of the LREE by accumulation of crystals, however, the presence/





Figures 8 and 9. Photomicrographs (crossed nicols) of type b biotite phenocrysts from inclusion (Fig. 8) and host (Fig. 9) of the Puripicar ignimbrite. Note the similarity in growth detail and the presence of type a phenocrysts as inclusion in both examples. Field of view is 3.5 mm in both plates.

Figure 10. Photomicrograph (partially crossed nicols) showing the distortion of vesicles in the host pumice (H) at the contact with a crystal-rich inclusion (I). The stretching of the vesicles is suggested to be due to the viscosity contrast of the two phases as a result of their differing crystal contents. Field of view is 3.5 mm.

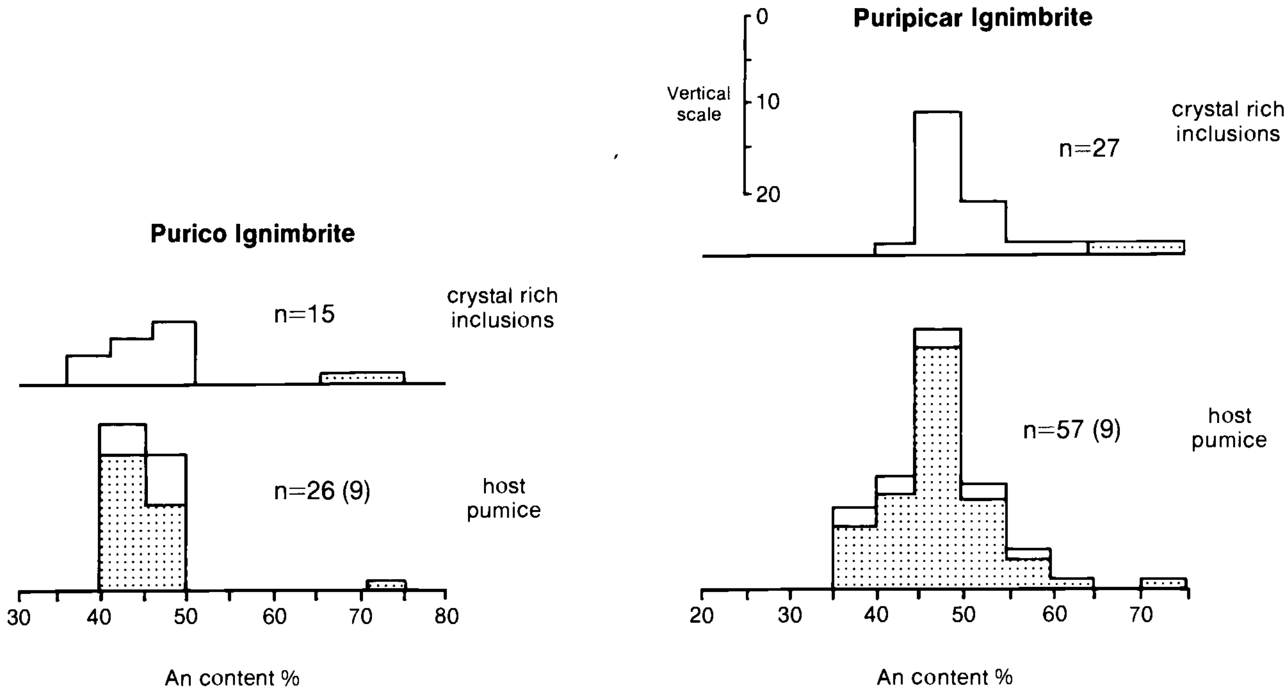


Figure 11. Histograms of the compositions (% An content) of plagioclases from inclusions and hosts from the Purico and Puripicar ignimbrites. Stippled, core compositions; clear, rims or unzoned phenocrysts; n, number of individual analyses. Number of rim analyses are given in parentheses.

absence of allanite may also be a contributory factor: the higher LREE contents of the host reflecting the presence of allanite.

The major, trace and REE characteristics of the inclusions therefore corroborate the mineralogical and mineral chemistry data and confirm that these inclusions can be interpreted as parts of the crystal accumulations resulting from crystal fractionation in the parent magma.

6. Evidence for accumulation at the side-wall

Where may these accumulations have occurred in the magma chambers? Crystal settling has been rejected convincingly by workers such as McBirney & Noyes (1979), Hildreth (1979), and Sparks, Huppert & Turner (1984), and this fractionation mechanism may be effectively ruled out for felsic magmas in general. Several observations on the inclusions and their hosts described here are also inconsistent with a crystal settling model for the origin of the accumulations.

(i) The host pumices in both ignimbrites contain quartz as a major phenocryst phase. This has a similar density to plagioclase (~ 2.6) and could reasonable be expected to occur in a cumulate containing plagioclase. Furthermore, the Puripicar host pumice contains allanite as discrete microphenocrysts. Even though they are 'late' crystallizers, occurring late in the fractionation history of the magma, these dense crystals (~ 4) may also reasonably be expected to occur in a cumulate from the host – it is possible, however, that their small size may have precluded them settling out.

(ii) The inclusions are restricted to early erupted portions of the ignimbrites. Assuming that the eruption tapped the magma chamber from the top downwards (i.e. Blake, 1981; Blake & Ivey, 1986), it is likely that these accumulations occurred in the upper parts of the magma chambers and not in the lower portions as would be expected from crystal settling.

These observations are best reconciled if the accumulations occurred at the upper margins of the magma chambers: at the roof or the sidewalls. The lack of these inclusions in the airfall base of the Lower Purico ignimbrite, suggests that the main crystal accumulations were not at the roof. Furthermore, in the light of the work of Brandeis & Jaupart (1986), it is unlikely that these inclusions are samples of convective crystal-rich plumes which descend from unstable boundary layers at the roof of the chamber, as these are unlikely to form in silicic magmas. Therefore, it is suggested that the crystal-rich inclusions are samples of the boundary layers which formed at the sidewalls of the magma chambers. In the discussion below, the following assumptions are made about the likely emplacement conditions and where and when crystallization may be expected to take place in a magma:

(a) The magma is emplaced at a higher temperature than the country rock and therefore will be cooled at the contact. The thermal gradient will be higher near the sidewall.

(b) The magma will start to crystallize in a boundary layer at the sidewall which will progress inwards.

(c) At some later stage in the cooling history of the chamber 'interior crystallization' would also begin in

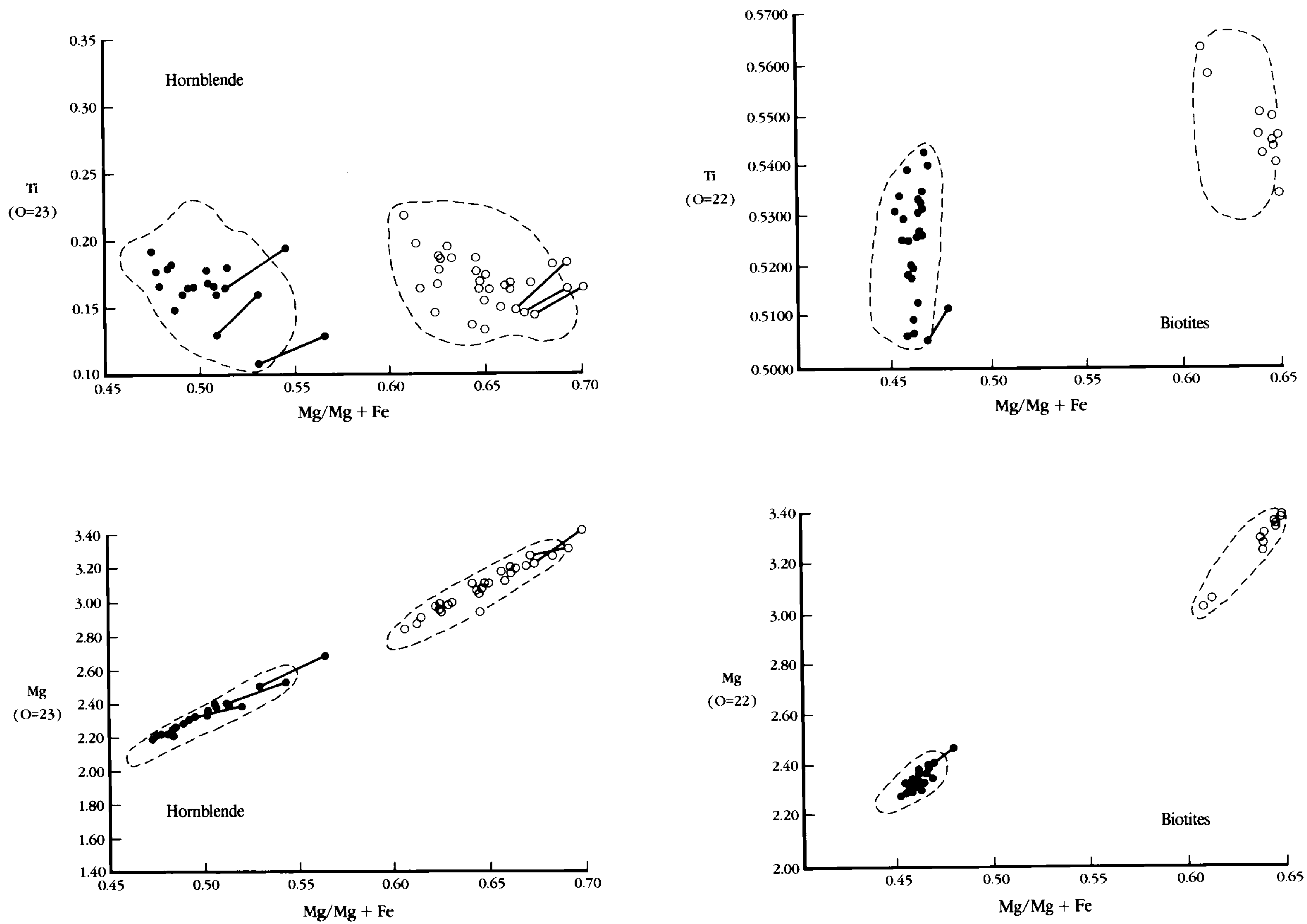


Figure 12. Compositions of hornblendes and biotites from inclusions from the Purico (○) and Puripicar (●) ignimbrites. The field defined by compositions of each mineral in the host pumice is outlined. The lines join core and rim analyses from grains where a compositional zonation was detected. Cores have higher Mg/Fe ratios. All data presented as formula units calculated on the basis of 23 oxygens for hornblende and 22 oxygens for biotite.

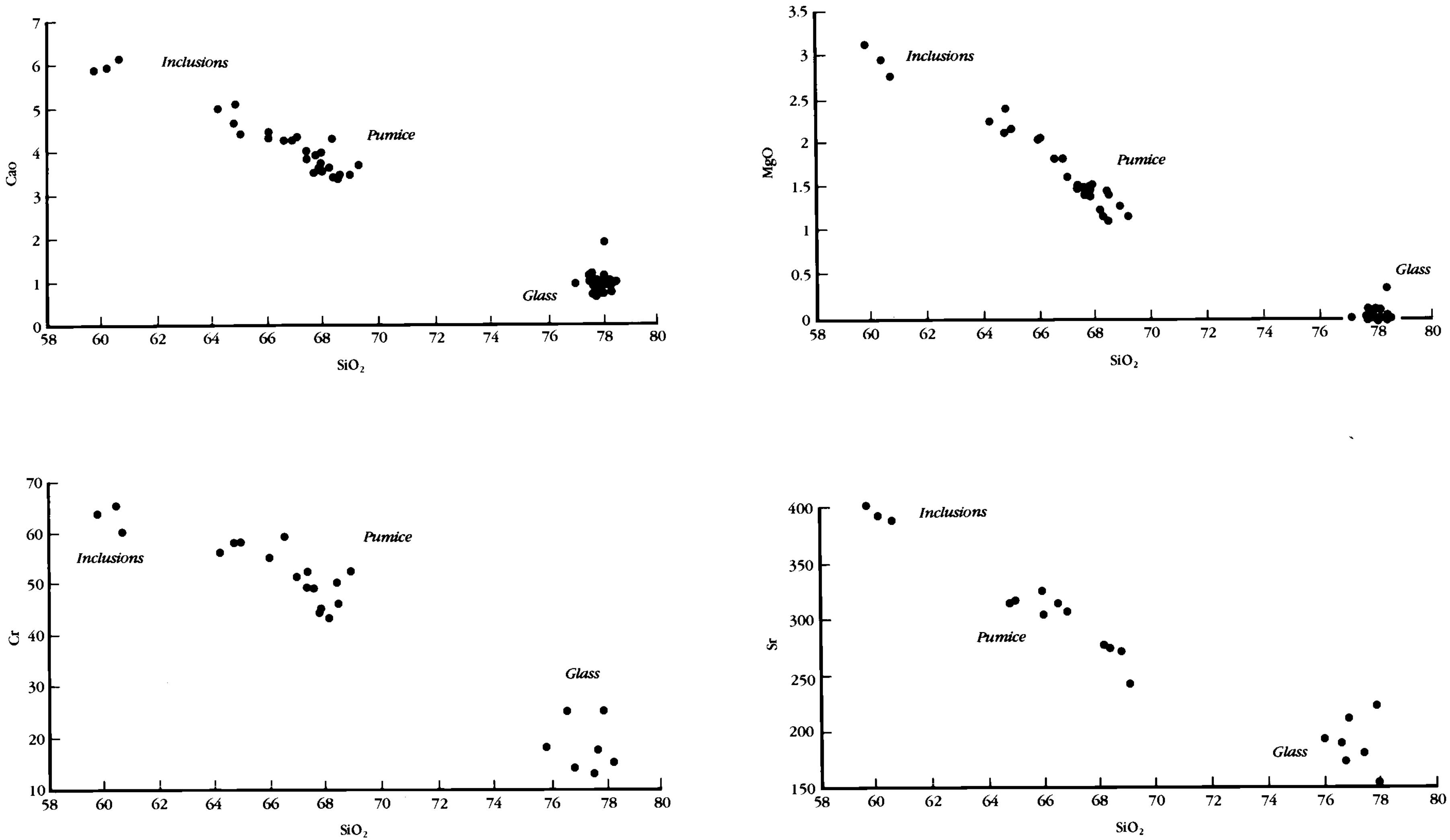


Figure 13. Harker diagrams of representative major and trace elements for inclusions, host pumice and glass from the Purico and Puripicar ignimbrites. Major elements for glasses are microprobe analyses of individual glass shards while trace elements for the glass were determined by XRF on bulk glass separates. All other data were obtained by XRF analyses of whole rock materials. Major elements are recalculated to 100% anhydrous.

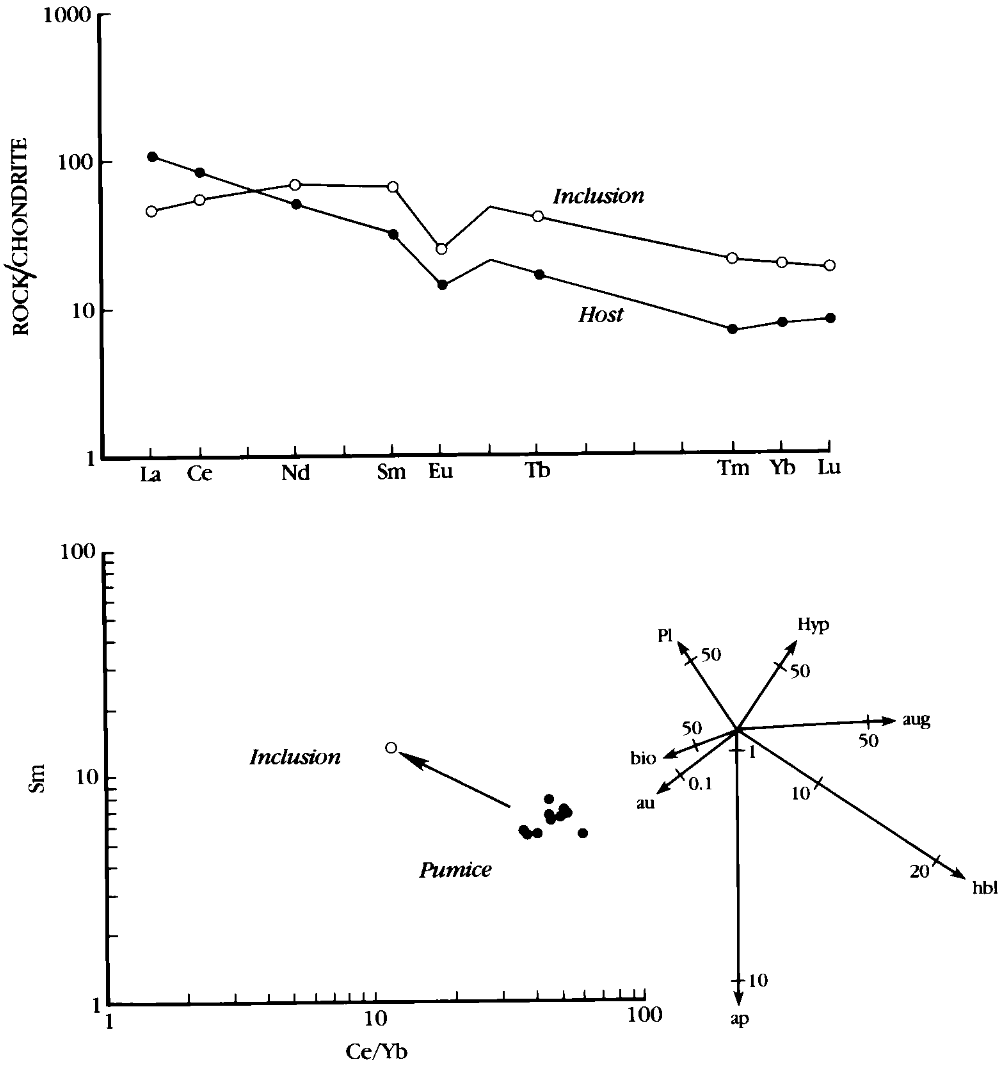


Figure 14. Available REE data for inclusion and host pumice. (a) Chondrite normalized REE patterns for inclusion 83018b and host pumice 83018 from the Puripicar ignimbrite. Note the depletion in LREE and the increase in overall REE contents of the inclusion. (b) Sm vs Ce/Yb for inclusion 83018b and all available data for the host pumices from the Puripicar ignimbrite. Arrow indicates direction of the fractionation process. These parameters provide a measure of MREE (Sm) enrichment in relation to the slope of the REE pattern (Ce/Yb). Comparison with the Raleigh fractionation vectors illustrates that the REE characteristics of the inclusion are best understood in terms of an increase in hornblende content relative to the host pumice.

the main mass of magma in the inner parts of the magma chamber. (Interior crystallization is used here to describe crystallization which occurs only in the main mass of magma as opposed to the sidewall).

(d) Both sidewall and 'interior crystallization' will then continue until the system is interrupted by eruption.

Several features of the inclusions and their host pumices are consistent with these assumptions;

(i) The larger proportion of hydrous minerals, biotite and particularly hornblende, in the inclusions, may be the result of the faster crystallization kinetics of these phases during the early stages of crystallization of Fe and Mg bearing granitic systems as predicted by crystal growth experiments (Naney & Swanson, 1980). Sidewall crystallates – the inclusions – represent the earliest stages of crystallization.

(ii) The presence of Mg-richer cores in some of the biotite and hornblendes of the inclusions would be consistent with the earliest crystallizing parts of the crystals being in equilibrium with a hotter more mafic liquid. As crystallization continues the liquid will evolve to the more felsic compositions that the crystals were in equilibrium with on eruption.

(iii) The absence of quartz and allanite in the inclusions, both of which are likely to crystallize at more advanced stages of fractionation, is consistent with the formation of the sidewall accumulations early in the fractionation history of the magma.

Observations made on the chilled margins of granitic intrusions, which are plutonic analogues for sidewall crystallates, support many of the assertions above. Chilled margins of many granitic intrusions often show a marked increase in biotite and amphibole

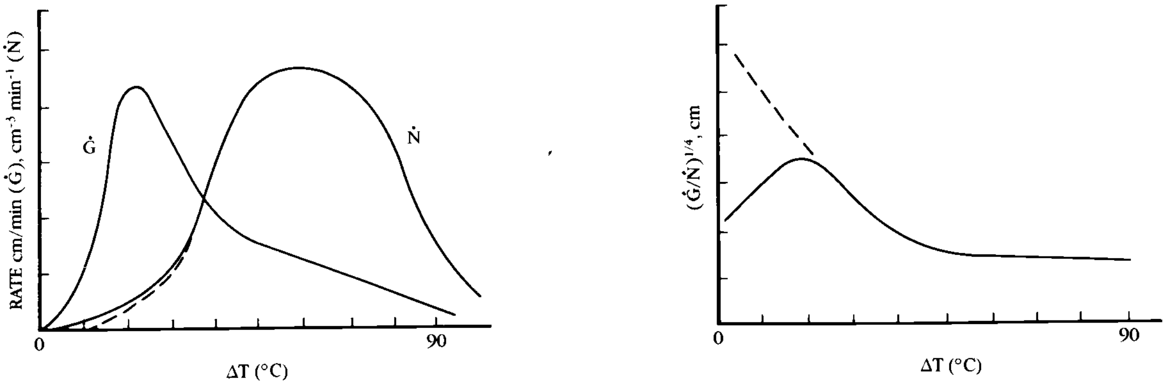


Figure 15. (a) Plots of nucleation rate ($\dot{N}$) and growth rate ($\dot{G}$) versus the degree of undercooling (ΔT) for nepheline determined experimentally by Winkler, as reported and modified (dashed line) by Shaw (1965). See also Lofgren (1972). (b) Plot of the grain size variation (represented by $\dot{G}/\dot{N}$) versus ΔT from the data of figure 7(a) (after Shaw, 1965).

contents relative to the interior. This is accompanied by a decrease in the amount of quartz towards the margins (e.g. – The Aigoual Massif, French Massif Central; Didier, 1973). In addition, chilled margins commonly display a microgranular texture while the interior of the intrusions have a much coarser porphyritic texture (Didier, 1973). These characteristics are very similar to those described above from the inclusions and hosts.

7. Discussion

Although much attention has been paid to understanding and predicting the chemical consequences of sidewall crystallization, little is known about the details of crystallization and the grain size spectra that may result during this process. Such grain size data is important in understanding the mechanism of release and segregation of melt from a growing crystal mush and other fundamental problems to igneous petrology. Recent work by Brandeis & Jaupart (1986), who address the issue of crystallization at the roof of a magma chamber, shows that the grain size spectrum which may result in a given situation is dependent on several factors, of which details of the thermal history of the magma, nucleation rates and crystal growth rates are the most important. A further complication is that crystallization will also contribute to the thermal history of the magma by latent heat release. Deduction of the grain size spectrum is therefore not straightforward. Nevertheless, a qualitative approach, based on observations in this study, is adopted below to try and offer some insights into the grain size spectrum that may have characterized the magma chambers of the Purico and Puripicar ignimbrites.

The difference in crystal content and grain size between the sidewall (inclusions) and the interior (host pumice) is likely to be a function of the nucleation rate and growth rate of crystals, and may therefore be understood in terms of Figs. 15a and b, which show the variation of these parameters with increasing undercooling (ΔT) in silicate liquids. It is noted however, that in detail the data represented in Fig. 15 is somewhat idealized as it doesn't take into account the effect of the changing composition of the melt, and hence viscosity, which is likely to have an effect on $\dot{G}$. Nevertheless the data has been shown to be adequate as a first approximation of the variation of $\dot{N}$ and $\dot{G}$ in magmas (Shaw, 1965; Lofgren, 1972). According to Figure 15a, at a small degree of undercooling, which would probably characterize the main mass of convecting magma represented by the host pumice, $\dot{N}$ is low while $\dot{G}$ is high (e.g. $\Delta T = 20$ °C), resulting in a coarse grain size. However, because of the forms of the two curves, a small increase in T (e.g. 40 °C) will cause a large increase in $\dot{N}$, a large decrease in $\dot{G}$ and therefore a marked reduction in grain size (Figure 15b). The larger values of ΔT, which are likely in a boundary layer at the sidewall of a magma chamber, will produce progressively finer grain sizes. Evidence of this may be found in the 'chilled margins' of many granitic intrusions (see for instance Didier, 1973), which show a marked decrease in grain size from the interior of the intrusions (porphyritic) to their margins (microgranular). The curves also predict that the crystal content at the sidewall will be considerably higher than in the interior. This is consistent with observations made by Spera, Yuen & Kirschvink (1982) and Thompson & McBirney (1985), who point out that the crystal content of the boundary layer will be considerably higher than in the central portions of the magma chamber.

The crystal content in the magma is likely to vary with distance from the sidewall. However, as shown on Figure 16, it is suggested that this variation is not uniform. The crystal content of the boundary layer will vary rapidly as a result of the high thermal gradient while in the interior of the chamber the crystal content is constant due to no temperature gradient as a result of convection. This may explain: (a) the range of crystal contents in the inclusions; inclusions derived from different distances from the sidewall will have different crystal contents, and (b) the homogeneous nature of the host pumices

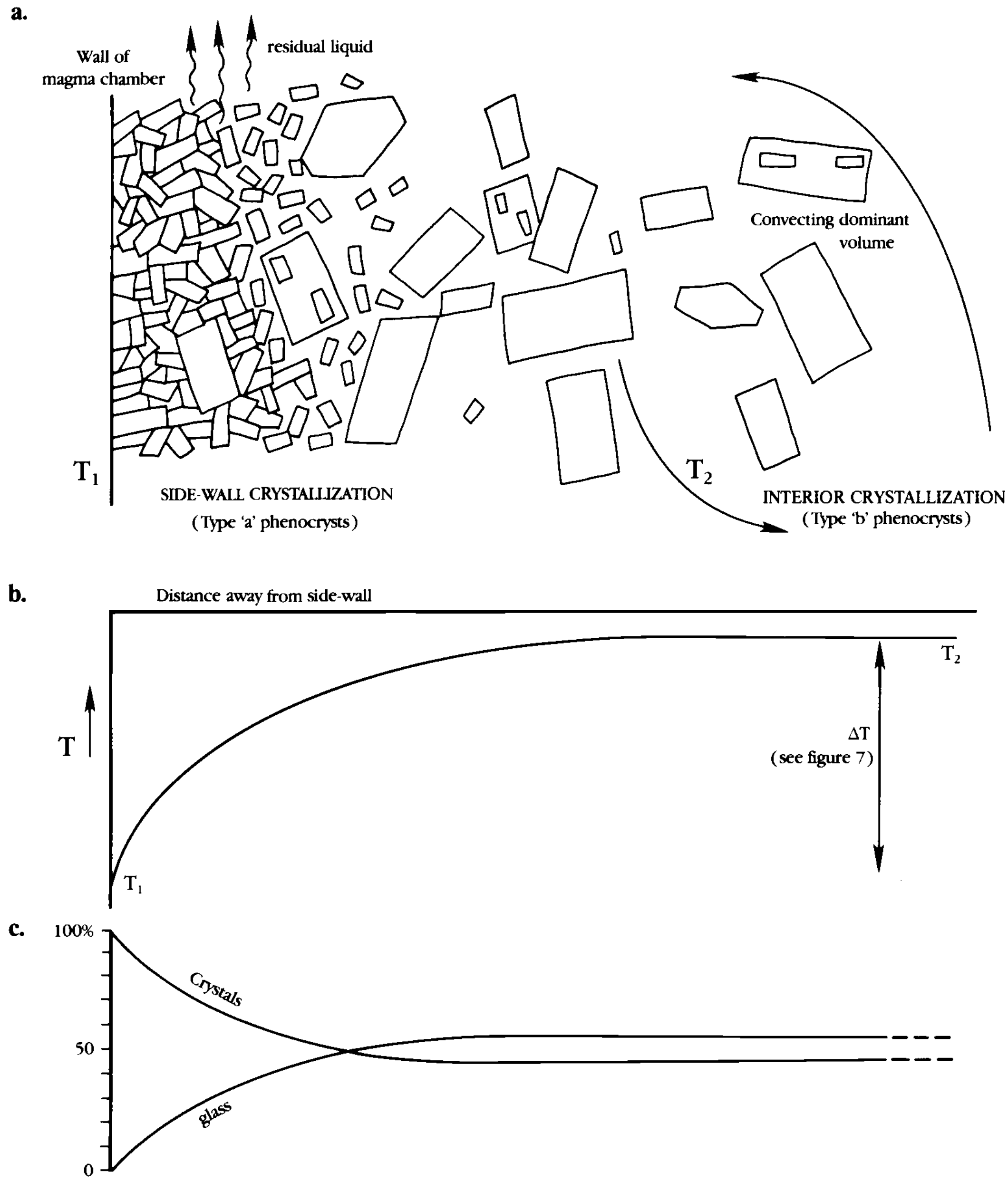


Figure 16. (a) Cartoon of a portion of a magma chamber, illustrating some of the inferred processes responsible for the characteristics of sidewall crystal accumulations. As $T1 < T2$, which are the temperature at the sidewall and the central portion of the magma at the time of intrusion respectively, rapid crystal growth at the sidewall results in a large number of type A phenocrysts. These form a crystal-rich carapace (mush) to the magma chamber. Buoyant residual liquids released will rise to the top of the chamber resulting in stratification of the chamber. No distinct interior to the boundary layer is shown here as it is envisaged that it will disrupt due to convection and redistribution of phenocrysts (see for instance Thompson & McBirney, 1985). Type B phenocrysts formed by (later) crystallization in the main magma (interior crystallization) are incorporated into the side-wall accumulations as the crystallization front grows inwards into the chamber. Type b phenocrysts may overgrow and include type a phenocrysts which are disaggregated from the crystallization front by convective currents. (b) Schematic representation of the temperature gradient under which the processes in a take place. ΔT is the degree of undercooling which may occur at the sidewall of a magma chamber. (c) Schematic representation of the variation of crystal and melt (glass) content across the portion of the magma chamber shown in (a), as inferred by petrographic observations.

(de Silva, unpublished Ph.D. thesis, Open University, 1987). Furthermore, if the crystal content of the inclusions is low, the viscosity contrast may be sufficiently small that disaggregation may take place due to shearing at the host-inclusion interface. This may also cause lower crystallinity in the inclusions. The contact between host and inclusion will in these cases be gradational. However, if the crystallinity is high, > 80 %, the inclusion may be sufficiently viscous to resist disaggregation by shear and contacts will then be sharp. Viscous coupling at the inclusion-host contact may lead to deformation of vesicles in the host in response to shear; for instance, in the conduit.

8. Concluding remarks

The role of side-wall crystallization in high-level magmatic processes has gained considerable importance through recent work on the fluid dynamics of magma chambers (Chen & Turner, 1980; McBirney, 1980; Sparks, Huppert & Turner 1984). This has lead to a consensus that crystal fractionation is the dominant process in generating layered magma columns in both calc-alkaline (e.g. Michael, 1983; Miller & Mittlefehldt, 1984; McBirney, Baker & Nilson, 1985) and highly alkaline systems (Wolff & Storey, 1984). There is, however, little geological evidence for sidewall processes. In this light, inclusions such as those reported here are an important link between the experimental and geochemical models. The occurrence of these inclusions in ignimbrites which have been erupted from layered magma chambers is undoubtedly significant and more detailed studies of these inclusions and their hosts may help further unravel the processes which generate compositional layering (e.g. Tait, 1987).

Apart from the ignimbrites described here, other ignimbrites contain similar textured inclusions. For instance, the 'aplitic' inclusions of the Young Toba Tuff, Toba, Sumatra (C. Chesner, M. Caress personal communication), and 'peppery textured' inclusions in the Ongatiti ignimbrite, North Island, New Zealand (authors observations), the Cerro Galan ignimbrite, northwest Argentina (P. W. Francis personal communication), and in most of the ignimbrites in the Central volcanic zone studied by the author. It is likely therefore that similar inclusions occur in other ignimbrites elsewhere. In addition, phenocryst rich 'cumulate' rocks erupted as inclusions in other volcanics have been described. Two particularly well documented examples are hornblende gabbro inclusions in ejecta from the 6845 yr b.p. climactic eruption of Mt. Mazama, Crater Lake, Oregon (Ritchey, 1980; Bacon, 1983; Bacon & Druitt, 1988) and Kaersutite bearing inclusions from the zoned phonolitic Laacher See Tephra (Worner & Schmincke, 1984; Tait, 1987). It is interesting to note that some examples of the latter are interpreted as samples of sidewall crystal accumulations. Other possible examples of cumulate inclusions in calc-alkaline volcanic rocks have been described from Soufriere Volcano, St Vincent (Wager, 1962; Lewis, 1973), Santorini, (Nicholls, 1971), Aso Caldera, Japan (Lipman, 1967) and Shikotsu Caldera, Japan (Katsui, 1963). Some microcrystalline enclaves in granitic plutons (Didier, 1973) maybe analogous material to the inclusions described here (de Silva, 1987). However, whether these represent sidewall accumulations of crystals will require reassessment of the field relations as well as the mineralogy and petrology of these inclusions.

Notwithstanding the fact that a crystal aggregate with interstitial vesicular glass may have a low preservation potential during explosive eruption, inclusion, and therefore representation, of such highly crystalline material (70–98%) among eruption products will depend on many factors of which magma viscosity and magma chamber geometry may be among the most important. Assuming similar magma chamber geometries and eruption dynamics the incorporation and eruption of these accumulations is likely to depend largely on the magma viscosity. Hence, although they may occur in many magma chambers they may not be erupted. In the Purico and Puripicar ignimbrites, these inclusions are found in crystal rich dacitic pumices ($\sim 50\%$ crystals), which may be due to the magma being sufficiently viscous to entrain and retain the crystalline material on eruption. Additionally, these inclusions are volumetrically insignificant ($< 0.01\%$) in relation to the total amount of pumice erupted and their presence may not actually be recognized unless detailed field observations are carried out.

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References

BACON, C. R. 1983. Eruptive history of Mt. Mazama and Crater Lake caldera, Cascade Range, U.S.A. *Journal of volcanology and Geothermal Research* **181**, 57–115.

BACON, C. R. 1986. Magmatic inclusions in silicic and intermediate volcanic rocks. *Journal of Geophysical Research* **91B**, 6091–112.

BACON, C. R. & DRUITT, T. H. 1988. Compositional evolution of the zoned calcalkaline magma chamber of Mount Mazama, Crater Lake, Oregon. *Contributions to Mineralogy and Petrology* **98**, 224–56.

BACON, C. R. & METZ, J. 1984. Magmatic inclusions in rhyolites, contaminated basalts and compositional zonation beneath the Coso volcanic field, California. *Contributions to Mineralogy and Petrology* **85**, 346–65.

BLAKE, S. 1981. Eruptions from zoned magma chambers. *Journal of the Geological Society of London*. **138**, 281–7.

Blake, S. & Ivey, G. N. 1986. Magma mixing and the dynamics of withdrawal from stratified reservoirs. *Journal of Volcanology and Geothermal Research* **27**, 153–78.

BRANDEIS, G. & JAUPART, C. 1986. On the interaction between convection and crystallization in cooling magma chambers. *Earth & Planetary Science Letter* **77**, 345–61.

CHEN, C. F. & TURNER, J. S. 1980. Crystallization in a double-diffusive system. *Journal of Geophysical Research* **85**, 2573–93.

DE SILVA, S. L. 1987. Microgranular enclaves in pumices – the products of crystal accretion at the sidewalls of silicic magma chambers (abstract) *EOS* **68(44)**, 1512–3.

DIDIER, J. 1973. *Granites and their enclaves*. Amsterdam: Elsevier, pp. 393.

EICHELBERGER, J. C. 1975. Origin of andesite and dacite: evidence of mixing at Glass Mountain and other circum-Pacific volcanoes. *Geological Society of America Bulletin* **86**, 1381–91.

FRANCIS, P. W., MCDONOUGH, W. F., HAMMILL, M., O'CALLAGHAN, L. J, & THORPE, R. S., 1984. The Cerro Purico volcanic complex, North Chile. In *Andean Magmatism – Chemical & Isotopic constraints* (eds R. S. Harmon and B. A. Barriero) pp. 107–124. Orpington U.K.: Shiva publishing.

GROVE, T. L. & DONNELLY-NOLAN, J. M. 1986. The evolution of young silicic lavas at Medicine Lake Volcano California: implications for the origin of compositional gaps in calc-alkaline series lavas. *Contributions to Mineralogy Petrology* **92**, 281–302.

GUEST, J. E. 1968. Banded Pumice in a Chilean ignimbrite. *Geological Magazine* **105**, 177–84.

GUEST, J. E. 1969. Upper Tertiary ignimbrites in the Andean Cordillera of part of the Antofagasta province of northern Chile. *Bulletin of the Geological Society of America* **80**, 337–62.

HAWKESWORTH, C. J., HAMMILL, M., GLEDHILL, A. R., VAN CALSTEREN, P. & ROGERS, G. 1982. Isotope and trace element evidence for late stage intra-crustal melting in the high Andes. *Earth and Planetary Science Letters* **58**, 240–54.

HILDRETH, E. W. 1979. The Bishop Tuff: Evidence for the origin of compositional zonation in silicic magma chambers. In *Ash flow tuffs* (eds C. E. Chapin and W. E. Elston). Geological Society of America Special Paper **180**, 43–76.

KATSUI, Y. 1963. Evolution and magmatic history of some Krakatoan calderas in Hokkaido, Japan. *Journal of the Faculty of Science Hokkaido University* Series IV, **11**, 1–58.

LEWIS, J. F. 1973. Petrology of the ejected plutonic blocks of Soufriere Volcano, St Vincent, West Indies. *Journal of Petrology* **14**, 81–112.

LIPMAN, P. W. 1965. Chemical comparison of glassy and crystalline volcanic rocks. *Geological Society of America Special Paper* **82**, 260–1.

LIPMAN, P. W. 1967. Mineral and chemical variations within an ash flow sheet from Aso Caldera, southwestern Japan. *contributions to Mineralogy and Petrology* **16**, 300–27.

LOFGREN, G. 1972. Temperature induced zoning in synthetic plagioclase feldspars. In *Dynamic crystallization and devitrification studies 1970–1978* (ed. G. Lofgren, 1979) pp. 35–50. Houston, Texas: NASA JSC.

MCBIRNEY, A. R. 1980. Mixing & unmixing of magmas. *Journal of Volcanology and Geothermal Research* **7**, 357–71.

MCBIRNEY, A. R. & NOYES, R. M. 1979. Crystallization and layering of the Skaergaard Intrusion. *Journal of Petrology* **20**, 487–554.

MCBIRNEY, A. R., BAKER, B. H. & NILSON, R. H. 1985. Liquid fractionation: Part 1: Basic principles and experimental simulations. *Journal of Volcanology and Geothermal Research* **58**, 51–64.

MICHAEL, P. J. 1983. Chemical differentiation of the Bishop Tuff and other high-silica magmas through crystallization processes. *Geology* **11**, 31–4.

MILLER, C. F. & MITTLEFEHLDT, D. W. 1984. Extreme fractionation in felsic magma chambers: a product of liquid state differentiation or fractional crystallization? *Earth and Planetary Science Letters* **68**, 151–8.

NAGASAWA, H. & SCHNETZLER, C. C. 1971. Partitioning of rare earth, alkali, and alkaline earth elements between phenocrysts and acidic igneous magma. *Geochimica et Cosmochimica Acta* **35**, 953–68.

NICHOLLS, I. A. 1971. Petrology of the Santorini volcano, Cyclades, Greece. *Journal of Petrology* **12**, 67–119.

RITCHEY, J. L. 1980. Divergent magmas at Crater Lake, Oregon: Products of fractional crystallization and vertical zoning in a shallow water-undersaturated chamber. *Journal of Volcanology and Geothermal Research* **7**, 373–86.

SHAW, H R. 1965. Comments on viscosity, crystal settling and convection in granitic magmas. *American Journal of science* **263**, 120–52.

SPARKS, R. S. J., HUPPERT, H. E. & TURNER, J. S. 1984. The fluid dynamics of evolving magma chambers. *Philosophical Transactions of the Royal Society of London* **A310**, 511–34.

SPERA, F. Y., YUEN, D. A. & KIRSCHVINK, S. J. 1982. Thermal boundary layer convection in silicic magma chambers: effects of temperature dependant rheology and implications for thermogravitational chemical fractionation. *Journal of Geophysical Research* **87**, 8755–67.

TAIT, S. R. 1987. Convective fractionation as a mechanism of magma chamber zonation: Evidence from cumulate nodules (abstract). *Terra Cognita* **7**, G11.24 357–8.

THOMPSON, M. E. & MCBIRNEY, A. R. 1985. Redistribution of phenocrysts by convective flow in viscous boundary layer. *Journal of Volcanology and Geothermal Research* **24**, 83–94.

WAGER, L. R. 1962. Igneous cumulates from the 1902 eruption of Soufriere, St. Vincent. *Bulletin Volcanologique* **24**, 93–4.

WALKER, G. P. L. & SKELHORN, R. R. 1966. Some associations of acid and basic igneous rocks. *Earth Science Reviews* **2**, 93–109.

WOLFF, J. A. & STOREY, M. 1984. Zoning in highly alkaline magma bodies. *Geological Magazine* **121**, 563–75.

WORNER, G. & SCHMINCKE, H-U. 1984. Mineralogical and chemical zonation of Laacher See Tephra (East Eifel, W. Germany). *Journal of Petrology* **25**, 805–35.