Ba partitioning and the origin of anorthoclase megacrysts in basaltic rocks

ALKALI FELDSPAR megacrysts are commonly found in continental alkalic basalts and their derivatives as euhedral–subhedral crystal fragments and often occur together with other megacrysts, such as aluminous clinopyroxene, aluminous orthopyroxene, kaersutitic amphibole, spinel, titanomagnetite and ilmenite, biotite, apatite, zircon and, more rarely, corundum. Mineralogical studies show that these megacrysts are Na-rich and relatively homogeneous. The majority of megacrysts have an orthoclase content (Or mol %) between 20 and 30 in the ternary feldspar system (Fig. 1). X-ray studies also show that these megacrysts possess a high degree of disorder, with the structural state belonging to the high-albite–high-sanidine series (Hoffer and Hoffer, 1973; Chapman, 1976). Previous interpretations of the origin of anorthoclase megacrysts fall into two categories: either (1) anorthoclase together with the associated pyroxene and amphibole megacrysts crystallised from basaltic magmas as phenocrysts at high pressure and were transported with the same magma during eruption; according to this hypothesis, the present host basaltic rocks represent the parent magmas for the anorthoclase megacrysts (Binns et al., 1970; Kutolin and Frolova, 1970; Hoffer and Hoffer, 1973; Laughlin et al., 1971); or (2) unlike the associated pyroxene and amphibole megacrysts, anorthoclase crystallised from a more evolved liquid at relatively lower pressure and was then picked up and carried to the surface by a later surge of alkalic basaltic magma in the overall eruptive episode. Thus the present host basaltic rocks do not represent the parent magmas of the anorthoclases (Irving, 1974; Chapman and Powell, 1976; Bahat, 1979; Upton et al., 1983; Irving and Frey, 1984; Aspen et al., 1990). The nature of this 'evolved melt' is speculative.

The partition coefficient (D), defined as the weight ratio of the element concentrations in the crystal phase and in the coexisting silicate liquid, is very useful in characterising element distribution in magmas. Given the partition coefficient of an element, the abundance of this element in the parent magma for a mineral of known composition can be calculated. The partitioning of Ba between alkali feldspar and silicate liquid may be applied to the characterisation of the chemistry of the parent magma for the anorthoclase megacrysts found in basaltic rocks. Following an earlier report of the experimental study of Ba partitioning between alkali feldspar and silicate liquid (Guo and Green, 1989), detailed evaluation of the partition coefficients (D_{Ba}) for Ba has allowed us to test the models of anorthoclase megacryst genesis.

Previous work

Berlin and Henderson (1969) first reported the direct measurement of alkali-feldspar–matrix partition coefficients ($D_{Ba}^{af/matrix}$) for Ba obtained by separating and analysing both phenocrysts and matrix. A large data-base of $D_{Ba}^{af/matrix}$ determined by such a method has accumulated since then, and the published $D_{Ba}^{af/matrix}$ data are summarised in Table 1. The partition coefficient shows marked variation; e.g. $D_{Ba}^{af/matrix}$ in trachyte and phonolite shows a range of 2.7 to 18.1 and 3.4 to 9.75, respectively. The variation in D_{Ba} is generally considered to be a function of temperature and pressure, as well as the volatile content of the magma. However, determining the physical–chemical condition of formation of specific minerals is not possible at our present level of knowledge of thermodynamics of complex silicate systems, and there is no unequivocal method of selecting numerical values for petrogenetic applications. Guo and Green (1989) crystallised alkali feldspars from a natural trachyte doped with Ba (hydrous and anhydrous) under controlled P–T

conditions (900–1100 °C, 10–25 kbar). The orthoclase contents of the synthesised alkali feldspars varied from 38 to 61, shown in Fig. 1. Partition coefficients for Ba between alkali feldspar and silicate liquid determined experimentally varied from 1.5 to 8.7, depending on pressure, temperature, the composition of alkali feldspar and the presence or absence of water. Although all but one of the alkali feldspars crystallised in the above experimental study are within the field of sanidines (Fig. 1), evaluation of the effect of feldspar composition on D_{Ba} allows reasonable extrapolation.

Choice of partition coefficient

Reconnaissance experiments carried out by Chapman (1976) suggested that anorthoclase is stable in a host basanite composition at near solidus temperatures and pressures up to 10 kbar under anhydrous conditions. These near-solidus experiments were conducted at temperatures of 1000–1100 °C. Stephenson (1990) recently measured the densities of fluid inclusions within corundum of an anorthoclase–corundum xenolith found in basaltic rocks and suggested that the assemblage may be formed at high pressure, with 10 kbar being the minimum. The partition coefficients ($D_{Ba}^{af/liq}$) determined at 15 kbar and 1000–1050 °C are therefore chosen to represent the conditions of anorthoclase formation, with a small amount of water permitted (Table 2). Considering the effect of feldspar composition on Ba partitioning, the natural anorthoclases must have lower partition coefficients for Ba since D_{Ba} decreases with increasing Na/K ratio of alkali feldspar (Guo and Green, 1989). Direct extrapo-
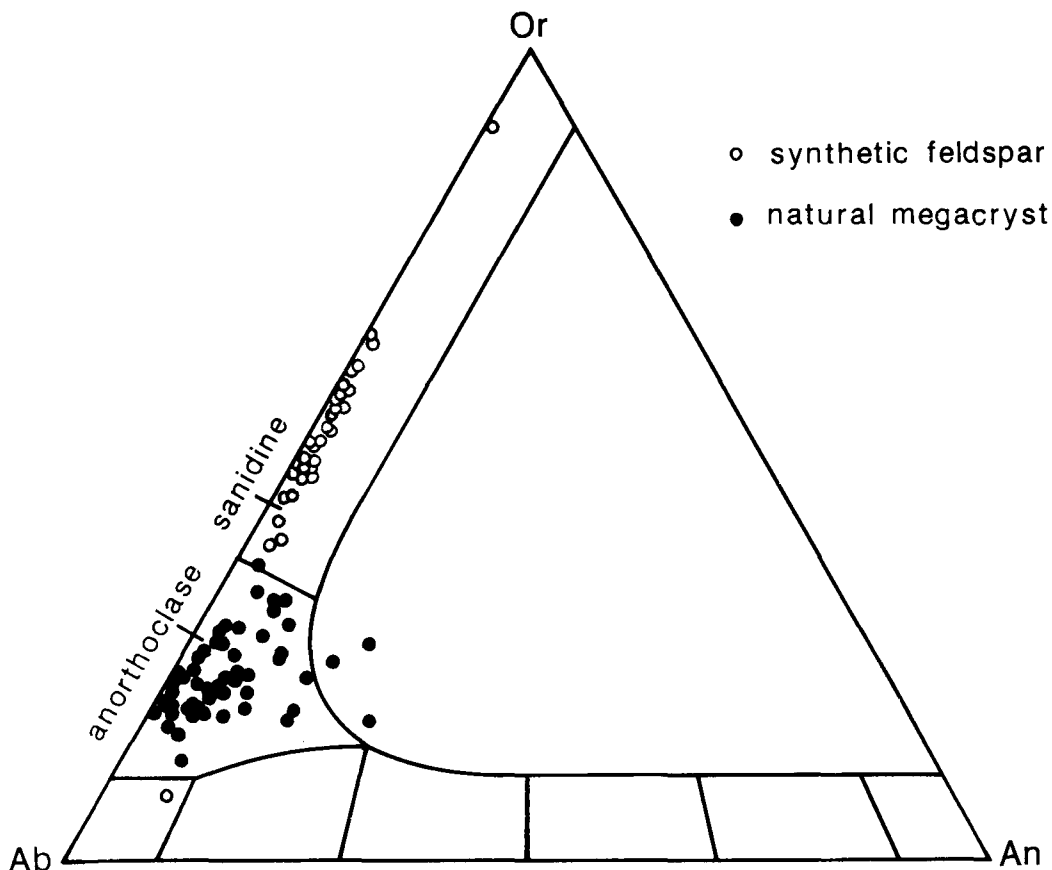

FIG. 1. Compositions of natural alkali feldspar megacrysts in alkalic basaltic rocks and synthetic feldspars crystallised from trachyte at 10–25 kbar in the temperature range of 900–1100 °C. Data sources: Binns et al. (1970), Kutolin and Frolova (1970), Hoffer and Hoffer (1973), Irving (1971, 1974), Chapman (1976), Bahat (1979), Mason et al. (1982), Irving and Frey (1985), and this work.

lation of the results of runs 1083 and 1269 in Table 2 gives a D_{Ba} value of 2.9 for an alkali feldspar of Or_{25} (equivalent to the average composition of natural anorthoclase megacrysts) at 15 kbar and 900 °C. The ln $D-T$ relationship for alkali feldspar of Or_{45} can be constructed using D_{Ba} of run 1119

Table 1   Compiled partition coefficients for Ba between alkali feldspar phenocryst and the matrix in natural volcanic rocks

Rock types	D_{Ba} af/matrix	Or (mole %)	Ab	An	Locality	Reference
Phonolite	5.04	n.a.	n.a.	n.a.	Ross Island, Antarctica	Sun & Hanson,1976
Peralkaline phonolite	2.83-9.75[1]	52.6-60.7	45.9-31.5	1.5-7.8	Laacher Sea Volcano, Germany	Wörner et al.,1983
Trachytic phonolite	3.4	40.9	56.5	2.6	Gardar, S. Greenland	Larsen,1979
Trachyte	13.5	58.5	38.6	2.9	Drachenfels, Rhine, Germany	Berlin & Henderson,1969[2]
Trachyte	18.1	60.4	37.8	1.8	Siebengebirge, Germany	(as above)
Trachyte	5.9	33.5	59.5	7.1	Jan Mayen, N. Atlantic	(as above)
Trachyte	2.7	49.0	45.7	5.3	Gough Island, S. Atlantic	(as above)
Trachyte	3.1-9.2	48.7-43.8	48.9-52.1	2.4-4.1	Tristan da Cunha, S. Atlantic	(as above)
Kenyte	4.4	16.6	65.0	18.4	Antarctica (BM 1910-199)	(as above)
Trachyte	3.6	n.a.	n.a.	n.a.	Massif, Central France	Villement et al.,1981
Trachyte	3.5	80.5	17.4	2.0	Roma Province, Italy	Francalanci et al.,1987
Rhyodacite	6.12	74.5	19.0	6.5	Naples, Italy	Philpotts & Schnetzler,1970
Rhyolite	20	65.6	33.3	1.1	Tweed Volcano, S. QLD, Australia	Ewart et al.,1977
Rhyolite	20-27	50.0-66.9	47.4-31.6	1.6-3.0	Yellowstone, USA	Leeman & Phelps,1981
Rhyolite	1-6.7	64.1-65.6	35.0-33.2	0.8-1.4	Bishop Tuff, Calif., USA	Mahood & Hildreth,1983
Rhyolite	3.4-7.5	n.a.	n.a.	n.a.	Twin Peak, Utah, USA	Crecraft et al.,1981
Rhyolite	7.2-2.4	47.4-69.1	51.2-29.8	1.1-1.4	Twin Peak, Utah, USA	Nash & Crecraft, 1983

n.a. Data not availabe
[1]   0.36 in the original paper is omitted because of the great disparity with all other results;
[2]   Directly calculated from phenocryst/matrix pairs in the original report to be consistent with the rest of the data in the table.

Table 2 Proposed crystallization conditions for anorthoclase megacrysts allowing testing via measured D_{Ba}

run no.	P (kb)	T (°C)	H$_2$O (wt%)	Or (af)	D_{Ba}
1083	15	900	4	58	7.8
1269	15	900	3	8.3	1.4
1119	15	1000	0.8	45	3.6
[extrapolation]					
step 1	*15*	*900*		*25*	*2.9*
*step 2**	*15*	*1000-1050*		*25*	*1.5*

* Using the correlation LnD=7871/T-5.66 for alkali feldspar of Or$_{45}$. See text for full discussions on the selection of D value.

and the corrected D_{Ba} of run 1083. Assuming the slope of ln $D–T$ regression does not vary with changing alkali feldspar composition, the D_{Ba} above (2.9) can be further extrapolated to estimate a D_{Ba} value at 1000–1050 °C. From this calculation, an approximate value of $D_{Ba} = 1.5$ is suggested as appropriate for anorthoclase megacrysts (Or$_{25}$). The nearly equal partitioning of Ba between silicic and basic melts in two-liquid partitioning experiments by Watson (1976) suggests that the partition coefficient for Ba between a given solid phase and magmatic liquid might remain nearly constant with changing liquid composition. Thus, the compositional differences between magmas in terms of the silica contents should not have a significant effect on the partitioning of Ba. The D_{Ba} value so derived (1.5) can then be applied directly to calculate the Ba concentration of the parent magma for anorthoclase megacrysts.

Discussion

Irving (1978) noted that equilibrium between megacrysts and host lavas could be evaluated by comparing megacryst/host trace element ratios with experimentally determined partition coefficients in systems of similar bulk composition under appropriate physical conditions. Comparisons of this type led to the conclusion that megacrysts of pyroxene and amphibole were in equilibrium with the host magma (Irving and Frey, 1984). However, Irving (1974, 1986) argued that anorthoclase, like zircon and corundum, could never be produced near the liquidus of alkali basalt composition under any $P–T$ conditions and further suggested that they may represent phases precipitated from highly evolved melt derived from basaltic magmas. Magmas,

residual after a large degree of crystallisation of mafic minerals such as olivine, clinopyroxene and orthopyroxene, would be enriched in silica and alkalis, enabling alkali feldspar to crystallise as the near-liquidus phase. The composition of this alkali feldspar would depend on the compositional characteristics of the parent magma.

Fig. 2 shows Ba abundances in alkali volcanic rocks from different provinces in relation to the well-established fractionation sequence from basalt through hawaiite, mugearite towards benmoreite and trachyte. Crystal fractionation of an alkalic basaltic magma by removing mafic phases and calcic plagioclase may result in the upper part of the trend. The lower part of the trend may be attributed to the participation of alkali feldspar during crystal fractionation. Ba abundance data in anorthoclase megacrysts compiled from different sources are illustrated in an inserted histogram in Fig. 2. This shows that the majority of anorthoclase megacrysts have BaO of about 0.12–0.32%, with all lying in the range of 0.01–0.55%. By applying a D_{Ba} of 1.5, the Ba content of silicate melt in equilibrium with the majority of the megacrysts can be calculated to be in the range 700–1870 ppm. The solid line at the bottom of Fig. 2 was reproduced from the inserted BaO histogram. It is noted that, near the end of the upper trend, the amount of Ba in the fractionated magma equivalent to benmoreite in composition (500–1400 ppm) is close to the whole range of Ba anticipated for magma from which anorthoclase megacrysts crystallised. Chapman and Powell (1976) investigated the density differences between anorthoclase and various types of magma following the theory that magmas of density similar to the anorthoclase megacryst allow the mineral to grow to a substantial size while the latter is suspended in the liquid. The density calculations indicated that the most appropriate magmas for anorthoclase crystallisation are similar to trachyandesite, the less K-rich counterpart of benmoreite. In Fig. 2, although some hawaiites and mugearites contain as much Ba as the predicted anorthoclase parent magmas, experimental studies of the phase relationships for a mugearite composition demonstrate that anorthoclase may appear near the solidus when the system is hydrous, but could not crystallise out of mugearite magmas as a near-liquidus phase under any conditions (Irving, 1971).

It is also noted from Fig. 2 that some trachytes are compositionally consistent with the likely parent magmas for anorthoclase megacrysts. Although anorthoclase has been observed as a common type of phenocryst in trachyte, the chemical composition of these phenocrysts

reveals a more complex growth history. The anorthoclase phenocrysts of trachyte are typically strongly zoned with variable K/Na ratios, (Table 3). These complex zoning patterns of anorthoclase phenocrysts in trachytes with sharp compositional changes contrast with the relatively homogeneous nature of anorthoclase megacrysts. Ewart (1985) thoroughly discussed these complex zoning patterns and argued that they are the result of a convective fractionation within a magma chamber. Furthermore, the crystallisation of trachyte at very deep level may not be feasible according to the experimental data of Guo and Green (1989). These data show that increase in pressure (above 15 kbar) tends to favour the crystallization of alkali feldspar of higher Or content from trachyte, which is in the opposite direction to that needed to produce anorthoclase megacrysts.

Other evidence for the origin of anorthoclase megacrysts comes from isotopic studies. Stuckless

and Irving (1976) summarised the existing isotopic data for anorthoclase megacrysts and their host basalts. These data do not demonstrate a consistent genetic relationship between anorthoclase and the host basalt. In some cases, the (^{87}Sr/^{86}Sr) ratios for anorthoclase megacrysts and the host are too different for them to be cogenetic (Dasch, 1969). In other cases, the (^{87}Sr/^{86}Sr) ratios are not significantly different or overlap each other, allowing a cogenetic relationship (Laughlin et al., 1971; Stuckless and Irving, 1976). However, more recent isotopic data (O'Reilly and Griffin, 1990) clearly demonstrate that within a given volcanic province, although individual anorthoclase/host pairs exhibit isotopic disequilibrium, the variations in the (^{87}Sr/^{86}Sr)$_i$ ratio for anorthoclase are within the range of (^{87}Sr/^{86}Sr)$_i$ ratios of basaltic rocks in that province. Aspen et al. (1990) also compared the Sr isotopic characteristics of both anorthoclase megacrysts and their host basalts. Despite the
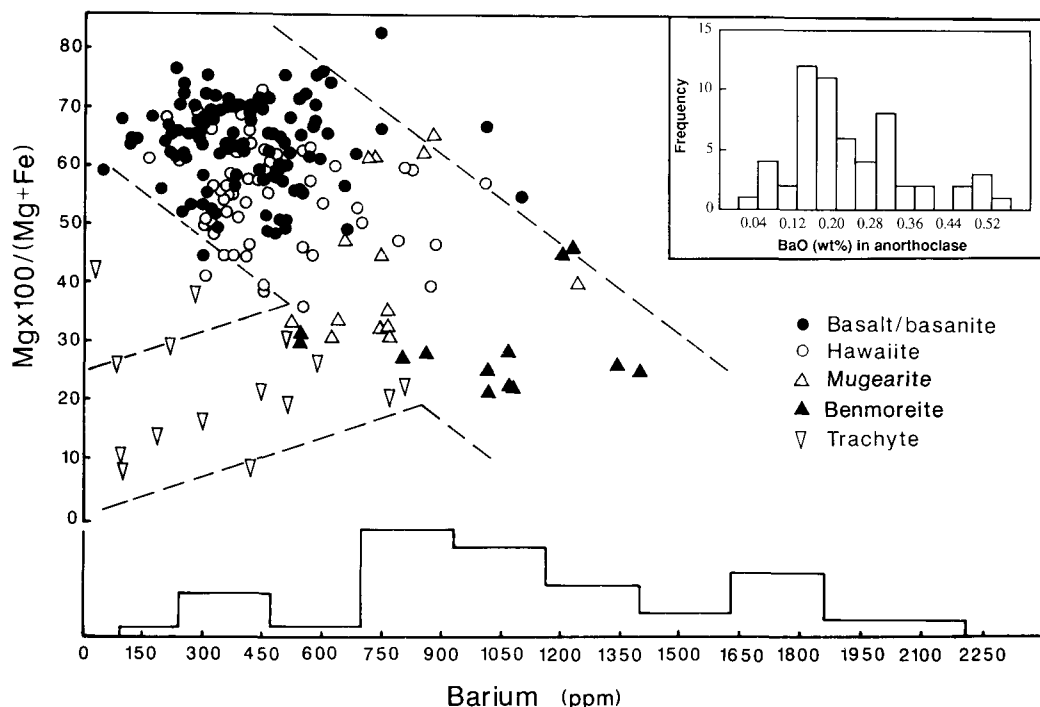

FIG. 2. Ba versus Mg $\times$ 100/(Mg + ΣFe^{2+}) atomic ratios plot for alkalic volcanic suites containing megacrysts and xenoliths. The inserted histogram shows Ba distribution in anorthoclase megacrysts from alkalic basaltic rocks and the solid line at the bottom of the diagram was deduced directly from BaO histogram, illustrating the anticipated Ba content distribution of the parent magma for the anorthoclase megacryst. Ba analyses were taken from Kutolin and Frolova (1970), Irving (1971), Bahat (1979), Mason et al. (1982), Irving and Frey (1984) and this work. Data sources for Ba in rocks: Knutson (1975), Baker et al. (1977), Rafferty and Heming (1979), White et al. (1979), Ghent et al. (1980), Wass (1980), Day (1983), Briggs and Goles (1984), Irving and Frey (1984), Nelson and Carmichael (1984), McDonough et al. (1985) and Nelson et al. (1987).

relatively large differences between the (^{87}Sr/^{86}Sr)$_i$ ratios for megacrysts and the host basalts (0.70287–0.70326 vs. >0.70326), both data sets fall into the entire range of regional (^{87}Sr/^{86}Sr)$_i$ ratio variations. Furthermore, isotopic and trace element analyses of a large number of anorthoclase megacrysts from a single locality reveal the presence of more than one generation of anorthoclase formation (O'Reilly and Griffin, 1990). These data suggest that anorthoclase megacrysts may have crystallised from a number of magmas, which are broadly similar in major element composition but different in trace element and isotopic composition, and were accidentally caught up in a later but different magma, produced during the same igneous episode.

It may be concluded from this study that anorthoclase megacrysts found in basaltic rocks crystallised from an evolved derivative of basaltic magma, which was most probably similar to benmoreite in composition. The crystallisation of such a magma at high pressure, around 45 km depth, may yield localised pegmatitic veins composed of large anorthoclase and other phases (cf. Irving, 1974). These anorthoclases appear to belong to the 'non-cognate' megacryst suite in basalt and an understanding of its genesis may have some general implications for other less

Table 3   Representative electron microprobe analyses of an anorthoclase phenocryst in trachyte and some average analyses of anorthoclase megacrysts in basaltic rocks

	Phenocryst in trachyte							Megacrysts in basaltic rocks			
	1	3	5	8	9	11	12	AK2	AK3	AK4	AK5
	(core --> rim)							(19)	(18)	(30)	(6)
SiO2	65.90	60.69	66.18	60.64	64.84	65.10	62.00	65.39	65.95	65.96	65.69
Al2O3	20.23	24.38	19.01	24.68	21.41	20.30	23.55	20.90	20.48	20.66	20.85
FeO	0.36	0.24	0.50	0.25	0.28	0.46	0.30	0.14	0.17	0.16	0.15
CaO	1.15	5.85	0.22	6.03	2.24	1.26	4.77	1.29	0.98	1.09	1.17
Na2O	7.33	7.42	6.29	7.32	8.08	7.64	7.75	8.01	8.22	8.29	7.95
K2O	5.27	1.40	7.23	1.28	3.31	4.88	1.75	3.74	3.88	3.68	3.90
BaO	n.d.	0.17	n.d.	n.d.	0.28	0.16	0.29	0.32	0.24	0.28	0.34
Total	100.24	100.15	99.43	100.20	100.44	99.80	100.41	99.79	99.92	100.12	100.05

Calculated on the basis of 8 oxygens

Si	2.936	2.711	2.986	2.703	2.878	2.919	2.760	2.914	2.934	2.928	2.920
Al	1.062	1.283	1.011	1.296	1.120	1.073	1.235	1.098	1.074	1.081	1.092
Fe	0.013	0.009	0.019	0.009	0.010	0.017	0.011	0.005	0.006	0.006	0.006
Ca	0.055	0.280	0.011	0.288	0.107	0.061	0.227	0.062	0.047	0.052	0.056
Na	0.633	0.643	0.550	0.633	0.695	0.664	0.669	0.692	0.709	0.713	0.685
K	0.300	0.080	0.416	0.073	0.187	0.279	0.099	0.213	0.220	0.208	0.221
Ba	-	0.003	-	-	0.005	0.003	0.005	0.006	0.004	0.005	0.006
Sum	4.999	5.009	4.992	5.002	5.003	5.016	5.007	4.989	4.994	4.993	4.987

An	5.6	27.9	1.1	29.0	10.8	6.0	22.8	6.4	4.8	5.3	5.8
Ab	64.1	64.1	56.3	63.7	70.3	66.2	67.2	71.6	72.7	73.3	71.2
Or	30.3	8.0	42.6	7.3	18.9	27.8	10.0	22.0	22.6	21.4	23.0

[1] Trachyte sample (MU6627) is from Mt Bulli, Comboyne Plateau, New South Wales, Australia, and described in details by Knutson (1975).

[2] Anorthoclase megacrysts are from Mt Anakie of western Victoria, Australia and the petrology of the host basalt was studied by Irving (1974); Numbers in bracket indicate the total number of analyses in averaging; Statistically, the variation of major elements is very small for each sample (2 standard deviation: SiO_2<1%, Al_2O_3<2%, Na_2O<3% and K_2O<8%).

[3] Counting time on Ba is extended 15 times longer than on other elements to achieve better counting statistics and to lower the limit of detection of Ba. L.L.D.s of Ba in phenocrysts and megacrysts are BaO 0.11% and BaO 0.13% respectively. n.d. denotes "not detected".

common megacrysts such as ilmenite, mica, zircon, apatite and corundum.

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