

Carbonatite magmas

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ABSTRACT. Normal carbonatite magmas are characteristically associated with ijolites and produce extensive fenitization by alkali metasomatism, commonly feldspathization or phlogopitization. The sequence of carbonatites developed from this magma are characterized by igneous isotopic ratios, high contents of incompatible elements, their distinctive pattern of differentiation from sövite to beforsite, and a late-stage rare-earth element, baryte, fluorite mineralization. Carbonatite rock is composed mainly of calcite, but fluid inclusion and related data indicate that the original magma was highly alkaline and chemically similar to that known at Oldoinyo Lengai volcano in Tanzania. It is suggested that this carbonatite magma is secondary magma produced from carbonated nephelinite magma by liquid immiscibility. Carbonatite magma can also occur associated with kimberlites but these carbonatites are alkali-poor and do not produce the characteristic fenitization.

CARBONATITES are defined as igneous rocks which contain more than 50% of carbonate minerals (Streckisen, 1980). The classes of carbonatite recognized are: (i) calcite-carbonatite, including sövite if coarse grained, alvikite if medium or fine grained; (ii) dolomite-carbonatite, sometimes also termed beforsite; (iii) ferrocarbonatite which carries essential iron-rich carbonate minerals; and (iv) natrocarbonatite which is composed of Na-Ca-K carbonates. The carbonate mineral in the bulk of carbonatites, the sövites and alvikites, is calcite. Dolomite, mostly in the beforsites, is less common. Ferroan carbonate, restricted mainly to the ferrocarbonatites is quite uncommon. The alkali content of carbonatites is negligible. The normal sequence of cationic abundances in the rock carbonatite is: $Ca > Mg > Fe > (Na + K)$. Carbonatite magma, as exemplified by the recent eruptions at Oldoinyo Lengai in northern Tanzania, is rather different with $Na > Ca > K > Mg > Fe$.

This contribution reconciles these two apparently contradictory statements, and attempts to characterize the properties of carbonatite magmas and their products.

Carbonatitic magmas

Mapping carbonatites in the field leaves little doubt of their magmatic origin. Many carbonatites occur as igneous bosses and plug-like bodies with margins disturbed by the attendant fenitization and carrying xenoliths, often flow oriented, from the surrounding country rocks. Many other carbonatites occur as dykes, veins, and cone-sheets with well-marked cross-cutting relationships and chilled margins. These and other igneous intrusive characters are documented in numerous publications, Eckermann (1948), Pecora (1956), Garson (1962), Heinrich (1966), Tuttle and Gittins (1966), and Le Bas (1977), to quote but a few.

But not all emplacements of carbonatite are said to be entirely magmatic. The large (20 km²) carbonatite at Sokli in northern Finland is described by Vartiainen and Paarma (1979) as having a magmatic carbonatite core of sövite surrounded by a metacarbonatite zone 1-2 km wide of sövite carrying abundant amphibole and other mafic minerals, the whole enclosed in a fenite aureole. The northern partial arcuate collar of aegirine-sövite around the main and pure sövite of the Wasaki carbonatite intrusion near Homa Bay in western Kenya, could also be interpreted to bear a similar relationship (Le Bas, 1977, p. 118). Replacement carbonatites are described by Armbrustmacher (1979) from the McClure Mountain, Gem Park, and Democrat Creek areas of Colorado, where primary magmatic carbonatite also occurs. All these replacement carbonatites are considered by the authors to be products of materials emanating from magmatic cores. On the other hand, some Russian authors consider carbonatites to be entirely metasomatic. Bulakh and Iskoz-Dolinina (1978) describe the carbonatite of the Turiy Peninsula of Kola as 'the results of repeated hydrothermal metasomatism', but admit 'the original rocks may have been the crystallization products from an alkalic-carbonate lava' (magma). Bowen

also held the opinion that carbonatites could not be magmatic since the melting temperature of calcite was too high (1340°C) and that they were entirely the product of carbonate replacement of earlier silicate minerals (1926). However, the experimental work of Wyllie and Tuttle (1960) and others have shown that such magmas are possible at small vapour pressures of water.

Another form of carbonatitic material is that found in the kimberlites. There it is mainly as calcite-rich globules or as calcite building the bulk of the kimberlitic groundmass (Dawson and Hawthorne, 1973). This carbonate material is apparently of igneous origin, but it is a matter of some debate whether the carbonates are genetically related to normal carbonatites (e.g. those of Uganda and Scandinavia) as advocated by Heinrich (1967) and Eckermann (1967) amongst others, or whether they are petrogenetically distinct as crisply stated by Mitchell (1979).

These kimberlitic carbonatites differ in many of their geochemical and mineralogical properties from normal carbonatites, and this is best seen by the lack in the former of the characteristic minerals such as pyrochlore, aegirine-augite, and the RE phosphates and carbonates, as well as the lack of any appreciable fenitization. The carbonatites of kimberlitic diatremes are almost completely devoid of any alkalis, nor are there any sodi-potassic igneous rocks intimately associated with them, unlike normal carbonatites which have a characteristic association with alkali-rich igneous and metasomatic rocks. The opinion of the author is that kimberlitic carbonatites are only very distantly related to normal carbonatites. They are not connected by a magmatic differentiation process, but they do both originate deep in the upper mantle as the result of differing styles of partial melting of mantle peridotite which has been subjected to different metasomatic processes. Bailey (1978) describes it as fluxing and metasomatism along different geotherms.

Extrusive magmatic carbonatites of the typical dominant calcite mineralogy have also been described. Carbonatitic tuffs occur capping the Got Chiewo satellitic volcanic cone on the south flank of Homa Mountain carbonatite complex in western Kenya (Le Bas, 1977). Carbonatitic tuffs and agglomerates are found interbedded with Miocene sediments under the large volcano of Tinderet in central Kenya (Le Bas and Dixon, 1965), and although these carbonatites are pyroclastic they maintain the typical high contents of incompatible elements Sr, Ba, Nb, and RE elements shown in Table I. Similar extrusive carbonatites are known from Kerimasi volcano in northern Tanzania (Dawson, 1964) and vesicular carbonatitic

lavas have been described from the Fort Portal area in western Uganda (von Knorring and Du Bois, 1961).

The natrocarbonatite lavas from the active volcano of Oldoinyo Lengai in northern Tanzania, only 10 km from Kerimasi, are chemically quite distinct from the above carbonatitic lavas which are dominantly of calcite with or without minor proportions of iron oxides and some silicate minerals. Natrocarbonatite is, as its name implies, a dominantly alkali carbonate rock (Table II) composed of two minerals: nyerereite, $\text{Na}_2\text{Ca}(\text{CO}_3)_2$, with some fairchildite, $\text{K}_2\text{Ca}(\text{CO}_3)_2$, in solid solution and gregoryite which is approximately the anhydrous equivalent of nahcolite, NaHCO_3 , plus a little CaCO_3 , both embedded in interstitial glass. Cooper *et al.* (1975) and McKie and Frankis (1977) have given precise descriptions of some of the mineralogical relations.

The problem which arises is: what is the genetic relation, assuming there is one, between the chemistry of the observed natrocarbonatite molten lava and that of the various solid and largely calcite-rich carbonatites observed in volcanic and subvolcanic complexes the world over? The chemical differences are so large, as shown in Table II, that early thoughts on the subject were that it was quite a separate rock. Indeed some authors (e.g. Milton, 1968) consider this alkali-rich carbonate lava to be derived from alkali-rich sediments such as are known in the vicinity, at Lake Natron and Lake Magadi, which have been pierced by a volcanic pipe. However, the other rock types such as nephelinite, phonolite, pyroxenite, ijolite, apatite-sövite, and fenite, are all perfectly normal associates of the Ca-rich carbonatite igneous complexes which suggests that Oldoinyo Lengai is an ordinary carbonatitic complex apart from the production of natrocarbonatite.

To examine this problem, it is instructive to consider the intrinsic properties of what we can term 'true' carbonatites, i.e. the normal solid carbonatites seen in the field, commonly associated with ijolites and accepted by most petrologists as being of igneous origin.

Hallmarks of 'true' carbonatites

Mineralogy. Carbonatites are typically composed of calcite or dolomite as essential constituents. One or more of the minerals apatite, phlogopite, magnetite, aegirine, aegirine-augite, and pyrochlore are frequently present. Less common accessories include arfvedsonite, orthoclase, olivine, fluorite, baryte, strontianite, bastnäsite, parisite, monazite, zircon, and baddeleyite.

Table I. Trace elements in p.p.m. in intrusive and extrusive carbonatites

	INTRUSIVE				EXTRUSIVE			LST
	1	2	3	4	5	6	7	
V	105	19	51	109	200	15	50	20
Li	15	55	23	29	19	50	30	5
Nb	560*	<53	50	1000	550	230	360	0.3
Y	114	130	10	91	65	85	90	9
Ce	} 5300†	-	950	2320	-	<300	>300	<20
La		630	320	1090	600	160	390	8
Sr	7500	>1000	1275	1000	5300	2200	2500	600
Pb	27	17	10	28	40	30	150	9
Ba	400	1000	1300	2900	1350	1100	1600	90

*includes Ta, † total REE.

1. Average carbonatite (Gold, 1966)
2. Average of 3 sovites from Tundulu, Malawi, (Garson, 1962)
3. Average of 6 sovites from Wasaki, W.Kenya (Barber, 1974)
4. Average of 20 alvikites from Homa Mt. W.Kenya (Barber, 1974)
5. Carbonatite lava, Fort Portal, W.Uganda (von Knorring and Du Bois, 1961)
6. Sovitic carbonatitic tuff, Legetet, Kenya (analyst: J.M.Rooke, Leeds Univ.)
7. Alvikitic carbonatitic tuff, Legetet, Kenya (analyst: J.M.Rooke, Leeds Univ.)
8. Average limestone (Wedepohl, 1978).

Table II. Chemical analyses of the range of igneous carbonatites

	1	2	3	4	5	6	7	8
SiO ₂	0.58	25.23	12.99	5.67	4.86	0.88	1.49	6.12
TiO ₂	0.10	0.46	1.74	0.50	0.17	0.18	0.22	0.68
Al ₂ O ₃	0.10	5.72	3.03	1.77	0.30	1.40	1.11	1.31
Fe ₂ O ₃	0.29	8.71	12.81	8.00	1.71	4.21	6.97	7.09
MnO	0.14	0.38	0.40	0.78	0.27	0.65	1.58	0.75
MgO	1.17	1.62	8.55	6.10	0.31	0.38	1.62	12.75
CaO	15.54	14.10	35.97	37.06	50.45	50.83	46.29	29.03
Na ₂ O	29.56	18.65	0.73	1.09	1.09	0.48	0.34	0.14
K ₂ O	7.14	4.89	0.20	0.87	0.17	0.04	0.22	0.79
P ₂ O ₅	0.95	1.49	3.32	1.73	1.00	1.02	1.63	2.66
CO ₂	31.72	8.75	14.79	32.16	34.84	37.57	33.97	37.03
H ₂ O ⁺	5.15	1.15	3.45	1.42	1.27	1.24	1.46	1.08
F	2.26	0.62	0.3	0.38	-	0.23	0.53	0.09
Cl	2.90	1.13	-	0.31	-	-	-	-
SO ₃	2.48	1.23	0.88	0.91	0.4	0.58	0.57	0.89
SrO	2.09	0.77	0.63	0.89	0.26	0.15	0.12	0.01
BaO	1.04	1.00	0.15	0.45	0.13	0.46	1.93	0.11

1. Natrocarbonatite, Oldoinyo Lengai volcano, N.Tanzania. Average of 4 samples extruded in 1960 and 1961 (Dawson, 1962; Du Bois et al. 1963).
2. Volcanic carbonatitic-nephelinitic ash from Oldoinyo Lengai eruption of August 1966 (Dawson et al. 1968)
3. Vesicular carbonatitic lava flow. Fort Portal, W.Uganda (von Knorring and Du Bois, 1961).
4. Average carbonatite (Gold, 1966).
5. Typical aegirine-sovite, Homa Mountain, W.Kenya (Le Bas, 1977; Barber 1974).
6. Average alvikite, W.Kenya (Le Bas, 1977).
7. Average ferrocarbonatite, W.Kenya (Le Bas, 1977).
8. Typical apatite-beforsite dyke, Alno, Sweden. (No.99, Eckerkmann, 1948).

Table III. Feldspathic end-products of fentization by carbonatite.

	1	2	3	4	5
SiO ₂	58.84	56.96	66.00	63.89	68.38
TiO ₂	0.37	0.26	0.45	0.08	-
Al ₂ O ₃	17.48	20.24	20.54	18.36	19.30
Fe ₂ O ₃	5.33	1.11	0.70	0.31	0.12
FeO	0.07	1.18	-	-	-
MnO	0.30	0.07	0.04	-	-
MgO	0.24	0.29	0.17	-	-
CaO	0.30	1.93	0.30	0.58	0.04
Na ₂ O	0.40	0.18	10.60	1.04	11.88
K ₂ O	14.96	14.08	0.24	16.08	0.14
P ₂ O ₅	0.26	0.06	0.02	0.01	-
CO ₂	-	1.47	0.36	-	-
H ₂ O ⁺	1.17	1.25	0.21	0.04	-
H ₂ O ⁻	0.13	0.36	0.27	-	-
	99.85	99.91	99.90	100.39	99.96

1. Orthoclase. Massive development near carbonatite on W.Ridge, Toror Hills, E.Uganda (SUT 197; Sutherland, 1965a).
2. Borengite. Dyke, Alno, Sweden (Eckerkmann, 1960). Includes BaO 0.13, SrO 0.01, F 0.37, S 0.26, total 100.21 less OqsF 0.30
3. Albitite. Derived from Burundian muscovite-schist by intrusion of Kirumba beforstic carbonatite (Denaeyer, 1966).
4. Orthoclase (Or₉₉Ab₈An₃) from orthoclase near sovite, S.flanks of Homa Mt., W.Kenya (HC798F). 'Wet' analysis by M.W.L. Blackley.
5. Albite (Or₁Ab₉₉An₀) from albitite in Ajui-Solapa carbonatite complex of Fuerteventura, Canary Islands (75#177). Electron Micro-Probe Analysis. Includes BaO 0.02, SrO 0.09.

Table IV. Estimate of bulk composition of carbonatitic magma trapped in fluid inclusions in apatite crystals.

	2.1	norm	
SiO ₂	2.1		
Al ₂ O ₃	1.0	Orthoclase	3.3
FeO	0.9	Corundum	0.4
CaO	19.8	Pyrrhotite	0.5
BaO	4.4	Magnetite	0.5
SrO	0.5	Halite	7.0
Na ₂ O	26.3	Sylvite	5.4
K ₂ O	6.1	Fluorite	2.9
CO ₂	29.2	CaCO ₃	30.1
SO ₃	4.4	SrCO ₃	0.7
Cl	6.9	BaCO ₃	5.7
F	1.4	Na ₂ CO ₃	32.8
S	0.2	K ₂ CO ₃	3.0
	102.4		
less O ≡ Cl, F, S	2.4	Na ₂ SO ₄	7.8
	100.0		100.0

Igneous association. Carbonatites occur in volcanic and subvolcanic diatremes, within continental and oceanic tectonic plates, commonly in association with ring-complexes comprising alkali-pyroxenitic, ijolitic, syenitic, and fenitic units and, unless subsequently removed by erosion, mantled by olivine-poor nephelinitic lavas. Continental examples are numerous and many are listed in the two books by Heinrich (1966) and Tuttle and Gittins (1966). The two oceanic examples, and both include the above range of rock types, are the Canary Island of Fuerteventura and several of the Cape Verde Islands including Brava, Fogo, Santiago, and Maio on which the author is currently working.

Fenitization. Intense alkali metasomatism is characteristic of carbonatites, particularly sövitic intrusions, but it is also known bordering alvikitic carbonatitic dykes and cone-sheets. Excellent examples have been described at Alnö, Sweden by Eckermann (1948), at Songwe Scarp in western Tanzania by Brown (1964), at Rufunsa and Chilwa in Malawi by Bailey (1966a) and Woolley (1969), and in Fremont, Custer, and Gunnison Counties of Colorado by Heinrich and Moore (1970).

The most frequent metasomatic product of fenitization by carbonatite is orthoclase feldspar together with a small proportion of iron oxides (commonly limonite or hematite) and often with a matrix of cementing calcite. The texture varies from cataclastic, turbid feldspar crystals being coated with iron oxides, to a rock with fresh Carlsbad-twinning interlocking plates of orthoclase. The latter seems to be the end product of the metasomatic process, and that orthoclase is strongly potassic. The norm of the analysis given in Table III, column 1, carries 5% hematite, 89% orthoclase, 1% albite and 1% nepheline. Perthosite, pseudotrachyte (Sutherland, 1965b) and microcline-rock (the 'burnt rock' of Heinrich and Moore, 1970) are also common depending on the stage at which the metasomatic process was 'frozen'. These fenites occur in zones sometimes exceeding 100 m wide but usually much less around sövite intrusions, and give strong evidence that an abundance of potassium has been transferred from an igneous source which must have been the carbonatite or its precursor, to the surrounding country rocks. The magma which filled the space now occupied by solid carbonatite must have been strongly potassic. In some instances within many complexes, instead of K-feldspar developing, phlogopite grew in its place as a result of higher water vapour pressures prevailing (Bailey, 1966b; Gittins *et al.*, 1975). The degree of potassium metasomatism bears little or no relation to whether the country rocks affected were acid, basic or otherwise (King and Sutherland, 1960).

In many carbonatite complexes, for example Toror in eastern Uganda, at Kaiserstuhl in W. Germany and at Alnö in Sweden, the K-rich feldspathic fenites appear to have become locally mobilized. The product formed dykes, some beyond the fenites and others cutting the carbonatites. The analysis (Table III, column 2) by Eckermann (1960) is typical of such dykes, and is very similar to that of the orthoclase given in column 1. He named this dyke rock borengite, but it is essentially a trachyte plus small amounts of calcite and iron oxides.

Unpublished X-ray and published optical data (Sutherland, 1965b) indicate that the potassium feldspar in the dykes is middle- to high-orthoclase sometimes with low-sanidine overgrowths. This contrasts with the potassium feldspar of the unmoved fenitic orthoclases in which the orthoclases usually have structural states near the orthoclase-microcline inversion. Sometimes, particularly in the brecciated orthoclases, they give obliquities (determined from the difference between the (131) and ($\bar{1}\bar{3}1$) peaks) of up to 0.5 indicating the microcline structural state although the feldspar optically appears to be orthoclase. It would seem, assuming a low-pressure subvolcanic regime, that the potassic metasomatism can take place at temperatures up to 400–500 °C.

With some carbonatites the feldspathization is sodic rather than potassic and leads to the development of albitite. Examples of this are the Lueshe and Kirumba carbonatites in eastern Zaire (Meyer and Bethune, 1960; Denaeyer, 1966), and some of the complexes of Fuerteventura in the Canary Islands and of the Cape Verde Islands. The norm of the albitite analysis given in Table III, column 3, is quartz and corundum normative with 90% albite and only 1% orthoclase.

Except for the Canary and Cape Verde Islands, the sodic metasomatism is usually associated with carbonatites exposed at deep levels of erosion. Three further examples of sodic fenites recently described are the early Proterozoic carbonatites of the Ukraine (Kapustin *et al.*, 1978), the late Proterozoic carbonatite at Newania in Rajasthan, India (Viladkar, 1980) and the large Palaeozoic carbonatite at Sokli in northern Finland (Vartiainen and Woolley, 1976) which all develop eckermannite- (and/or aegirine-) bearing syenitic fenites with $Na > K$. This could be interpreted to indicate that there is a vertical zonation of the fenitization, with sodic fenites developed at deeper levels and potassic fenites formed at shallower depths near to the volcanic surface. Such a zonation could be largely temperature controlled, with the mobility of K being greater than that for Na as the temperature decreases as found experimentally by Orville (1963).

Differentiation. Work on the carbonatites of western Kenya, particularly that by Dixon and Flegg (in Le Bas, 1977), brought out the feature that carbonatites show an almost constant sequence of emplacement. The earliest carbonatite emplaced within a complex is usually sövite (C_1), and invariably cutting this and the attendant feldspathic fenites described above, are the alvikites and any beforites (C_2) as swarms of dykes and cone-sheets. Ferrocarbonatite (C_3) dykes and veins transgress the C_1 and C_2 carbonatites, and in some instances a late-stage residual calcite-carbonatite (C_4) can be recognized. The latter occurs as thin veins and stringers composed usually entirely of a mosaic of small calcite crystals; sometimes the calcites are large and the vein can look sövitic. A similar C_1 - C_2 - C_3 sequence is seen among the carbonatites of the USSR (Kapustin, 1971).

The intrusive sequence C_1 - C_2 - C_3 is also the fractionation sequence with increasing iron content of the carbonatites (secondary effects often cause much of this iron in the carbonates to exsolve as iron oxides), and it corresponds to the crystallization differentiation sequence determined by Wyllie (1965) and Rosenberg (1967) in their study of the system CaCO_3 - MgCO_3 - FeCO_3 .

Igneous differentiation is also evident in the changing proportions of the main trace elements Sr, Ba, RE, and Mn. Sr decreases on passing from sövites to the later carbonatites whilst Ba, Mn, and RE elements rise by an order of magnitude, and most carbonatite complexes show a baryte-fluorite-REE-uranium-thorium mineralization associated with the C_3 stage of carbonatite emplacement. The RE enrichment in some of the Ontario carbonatites of Canada is fully documented by Culers and Medaris (1977). Such mineralization is not seen in kimberlitic carbonatites.

The values of the isotopic ratios of O and C again indicate igneous derivation. The $\delta^{18}\text{O}$ per mil values of the least evolved sövitic material are frequently as low as +6 relative to SMOW (and usually less than +10), which is characteristic of mantle origin. Unaltered natrocarbonatite gives similar values. However, later-stage fractionates of carbonatites can reach values as high as +30 (Suwa *et al.*, 1975), indicating extensive equilibration with other oxygen reservoirs at low temperatures (Dontsova *et al.*, 1977).

The $\delta^{13}\text{C}$ per mil values for most carbonatites are restricted to the range -2 to -9 P.D.B. (Deines and Gold, 1973; Sheppard and Dawson, 1975) which, while being very similar to the range in diamonds (Deines, 1980) does not unequivocally rule out involvement of carbon reservoirs other than those of the Earth's mantle (Vinogradov *et al.*, 1979).

The strongest evidence for mantle origin however comes from radioisotope data. Sr, Nd, and Pb isotope ratios which have been determined for a wide spectrum of carbonatites (Bell *et al.*, 1973; Pineau and Allègre, 1972; Richard *et al.*, 1976; Lancelot and Allègre, 1974), are very similar to those of the alkali basalts and the nephelinites with which carbonatites are associated, whereas other possible source materials such as sedimentary carbonates would be different.

Fluid inclusions. The apatites of both carbonatites and their associated ijolites frequently abound in primary fluid inclusions (Rankin and Le Bas, 1974a; Nesbitt and Kelly, 1977; Aspden, 1980). The determination of temperatures of homogenization by Rankin (1975), Romanchev (1972) and others indicate normal magmatic temperatures in the region of 1000-1100 °C for the ijolites, but a much lower set of temperatures for the carbonatites: sövites about 400 to 600 °C, and alvikites, beforites and ferrocarbonatites in the region of 200 to 400 °C.

There is now experimental evidence (Freestone and Hamilton, 1980; Verwoerd, 1978; Rankin and Le Bas, 1974b) that carbonatite magma may be one of the products of liquid immiscibility which could take place in a parental carbonated nephelinitic magma, the other product being nephelinitic/ijolitic magma in accordance with the field evidence (Le Bas, 1977). If this is so, then the homogenization temperatures show there must have been an interval and considerable temperature drop after ijolite formation and before the main crystallization of the carbonatite magma could begin. It is suggested that during that interval, two significant processes were able to take place in this superheated carbonatite magma: the crystallization of apatite over a wide range of liquidus temperatures; and the attainment of a fair degree of chemical equilibrium by exchange processes enveloping country rock.

Micro-probe studies on apatites from a range of carbonatites indicate a cryptic variation in the apatite compositions as crystallization developed from sövite to alvikite and ferrocarbonatite. Preliminary results (Le Bas and Handley, 1979) show that Ca and P decrease and Ce, La, and F increase in the apatites as crystallization proceeded in the western Kenyan carbonatitic magmas, confirming that we are dealing with a fractionation sequence. Similar results for F have been obtained for carbonatitic apatites in the USSR (Chernysheva *et al.*, 1976). Apatites crystallizing over such a big temperature interval would have an unusually great opportunity of trapping fluid inclusions, and it would be greater still if there were spurts of nucleation resulting from undercooling. The most common daughter products of cooling primary fluid inclusions trapped from carbonatitic magma

are Na-Ca-K carbonates including nahcolite, alkali sulphates, alkali halides, some barium salts and silicate glasses, and captured mineral phases including iron oxides and sulphides, and calcite (Rankin, 1977; Aspden, 1980). The trapped fluids, rich in Na and K, provide us with a clue to the nature of the original carbonatite magma which was rich in Ca, Na, K, also in sulphate (sulphide in the earlier stages) and halide molecules. The high

F and Cl contents are in accordance with an origin by liquid immiscibility (Kogarko and Ryabchikov, 1978).

A preliminary calculation of the composition of this alkali-rich carbonatite magma, deduced from the bulk composition of the trapped minerals, is given in Table IV. The calculation was based on the mean modal proportions, determined photographically, of the various phases. Aspden (1977)

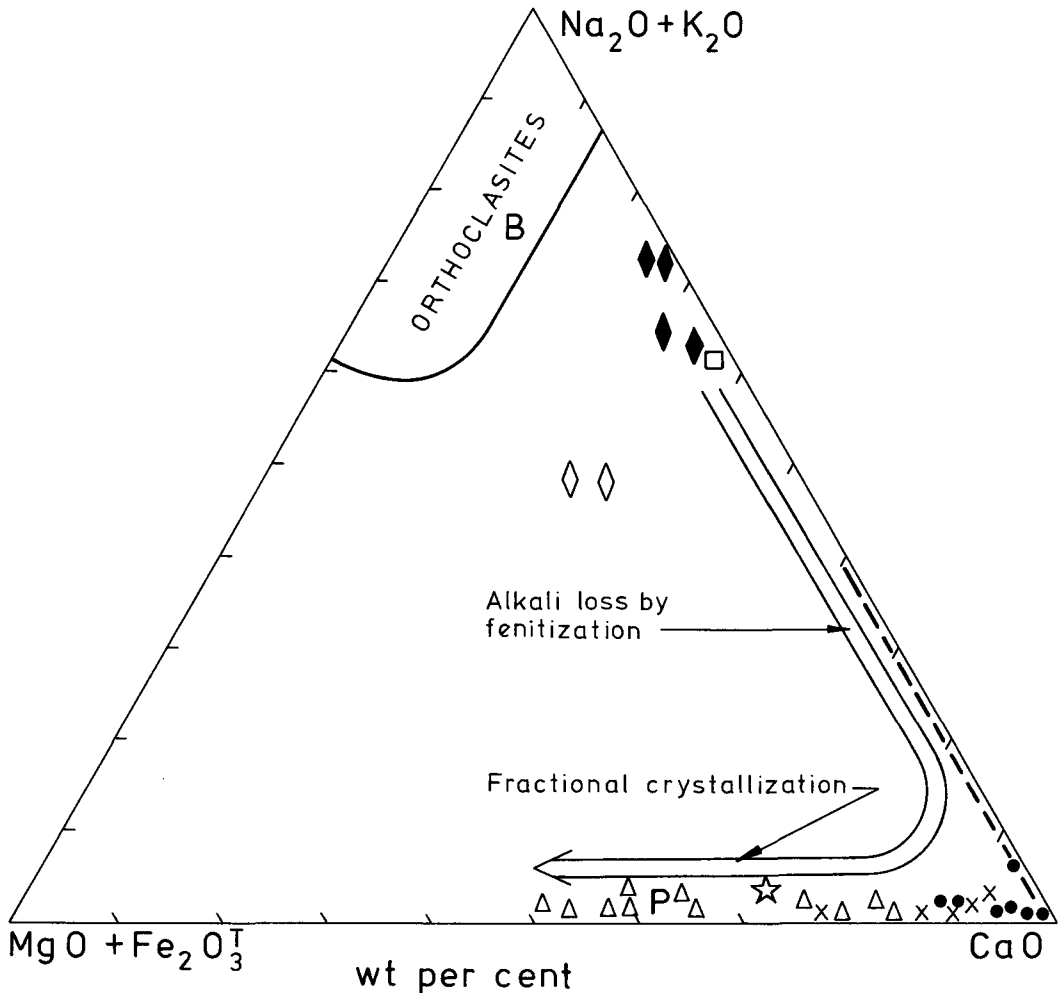


FIG. 1. Plot of carbonatites on $(\text{Na}_2\text{O} + \text{K}_2\text{O})$ - CaO - $(\text{MgO} + \text{Fe}_2\text{O}_3^{\text{T}})$ wt.% variation diagram showing the trend schematically indicated of derivation from natrocarbonatite to sövite by alkali loss, then to alvikite and to ferrocarbonatite and beforite by fractional crystallization and late stage processes. Symbols: closed diamonds = natrocarbonatite; open diamonds = Oldoinyo Lengai ash which includes clasts of nepheline and pyroxene crystals; square = bulk composition of trapped materials in carbonatitic apatite crystals (Table IV); dots = sövite; crosses = alvikite; triangles = ferrocarbonatite and beforite; (sövites, alvikites, ferrocarbonatites, and beforites all taken from examples at Alnö, Tundulu and Kenya); P = Fort Portal lava; star = average carbonatite of Gold (1966). Field of carbonate-free orthoclases is also shown, B = borengite (Eckermann, 1960). Dashed line between CaO and $\text{Na}_2\text{O} + \text{K}_2\text{O}$ is range of compositions of trapped solid carbonate inclusions in apatites in eastern African carbonatites (Aspden, 1977).

determined the composition of many of the individual phases by beam analysis using the electron microprobe, others were characterized by X-ray back-scatter imagery. These analytical data, combined with thin section information, permitted chemical identification of the phases so that the mode could be calculated into a chemical analysis. A paper is in preparation giving the details. The composition calculated resembles that of natro-carbonatite lava, and the conclusion may be drawn that the carbonatite magma at the time of growth of the early apatite crystals had a chemical composition not far removed from that of the natro-carbonatite lava.

It is also apparent that if a carbonate magma of such alkali-rich composition existed at temperatures of 500–1000 °C deep in the continental crust, it would certainly react with the silicates there. The standard way in the laboratory to dissolve granite or other silicate material is to add sodium carbonate and heat. The natural chemical laboratory of the Earth's crust has done just this over the interval mentioned above. Thus it is not surprising that by the time a carbonatite magma has cooled and crystallized, it has lost the majority of its alkalis by reaction, and the solid carbonatite encountered in the field is alkali-depleted and relatively enriched in Ca and the incompatible elements.

Fig. 1 shows the extent of this alkali depletion from natrocarbonatite magma to sövite and the concomitant concentration of the alkalis to give feldspathic fenites. Only the potassic fenites are plotted since there are very few data as yet on albitites of undoubted carbonatitic parentage. The sodian calcite compositions shown in fig. 1 of the early solid trapped inclusions in the apatites is another clue that the early carbonatite magma was alkali-rich. Fig. 1 also indicates that some of the ferrocarnatites are not as 'differentiated' as previously supposed (Le Bas, 1977), but that they may be late-stage carbonatitic fluids metasomatically imposed on earlier carbonatitic intrusions. Beforsites however do plot as the extreme of fractional crystallization.

A further feature can be argued if it is accepted that carbonatite magma, arising from liquid immiscibility, would be superheated. According to the Cann (1970) effect, a superheated magma would be able to intrude more easily to higher levels in the Earth's crust than a silicate magma on the liquidus. This would explain the observed greater frequency of intrusions of carbonatite alone than ijolite alone. Normally both occur together.

Conclusions

Not only do the mineralogy and differentiation features track out the history of the carbonatite

magma, but the alkali metasomatic fenitization processes and the fluid inclusions provide further vital clues of the nature of 'true' carbonatite magmas. These are interpreted as showing that the chemical composition of natrocarbonatite is approximately the same as that of the parental magmas which must have existed at all 'true' carbonatitic igneous centres.

Finally, it might be noted that if carbonatite magma is the product of liquid immiscibility from an alkaline silicate liquid, there is every likelihood that the precise composition of the carbonatite liquid separating would change progressively with change in temperature and pressure, and recent experimental work by Hamilton *et al.* (1979) shows this. The solvus defining the boundary between the one-liquid state and the two-liquid state is at least divariant, and thus it is possible that the Ca:Na:K ratio of a carbonatitic liquid separating from a nephelinitic magma might differ from that for a later carbonatitic liquid separating from some differentiation production of the original nephelinitic magma. Such dynamic un-mixing of carbonatitic and nephelinitic melts could give rise to several chemically distinct carbonatite magmas, each of which could subsequently differentiate along their own individual paths. This could explain the slightly different patterns of fenitization seen at some carbonatite complexes.

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