American Mineralogist, Volume 60, pages 566–573, 1975

Armalcolite-Ti-Phlogopite-Diopside-Analcite-Bearing Lamproites from Smoky Butte, Garfield County, Montana

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Abstract

Lamproite dikes and plugs occurring near Jordan, Montana, contain highly unusual mineral assemblages of three major types: (I) sanidine, armalcolite, and alkali amphiboles (potassium richterite and potassium riebeckite); (II) armalcolite, Ti-phlogopite, diopside, olivine (altered), and sanidine; (III) armalcolite, Ti-phlogopite, diopside, olivine (altered), analcite, and glass. These rocks are low in Al_2O_3, rich in TiO_2, and extremely rich in K_2O. They may be related in origin to the TiO_2- and K_2O-rich ultramafic rocks associated with the Montana diatremes.

Introduction

The Smoky Butte intrusives have been described by Matson (1960), where pertinent field data and a summary of previous work can be found. The outcrops of these potassic lavas are located 8 miles due west of Jordan (Montana), between Smoky Butte Creek to the north and the junction of Big Dry Creek and Lone Tree Creek to the south. The nearest eruptive rock, a nepheline-haüynite alnöite, has been described in Winnett (Ross, 1926; Powell and Bell, 1970) 90 miles south and west of Smoky Butte. A number of kimberlitic bodies and associated alkaline intrusions (characterized by forsterite phenocrysts in a groundmass of melilite, monticellite, nepheline, Ti-phlogopite, perovskite, and opaques) have been reported from north central Montana (Hearn, 1968) 100 miles west of the area considered here.

The Smoky Butte outcrop extends north-south for 1.9 miles, mainly as thin dikes and small plugs which metamorphose the Paleocene sediments of the Tullock member of the Fort Union formation (Matson, 1960). No extrusive rocks have been found.

The samples studied are from a quarry on the south side of Smoky Butte. Three major rock types (Matson, 1960) are present. Type III—composed of armalcolite, Ti-phlogopite, diopside, olivine, analcite, and glass—forms the narrow center part of a non-symmetric plug surrounded by type II, which is similar in composition to type III but for the presence of sanidine and the absence of glass. Type II is in turn surrounded by type I—a fine grained association of sanidine, armalcolite, and alkali amphiboles, with abundant inclusions of sedimentary rocks.

Mineralogical Descriptions

Sanidine

The only feldspar found at Smoky Butte is sanidine. It is present in types I and II. In type I it represents the greatest part of the rock, occurring as poorly-defined, very small crystals. In type II it is present as long euhedral crystals with square cross sections, often forming radial clusters (Fig. 1). These crystals often show hollow cores filled with glass which are quite similar in their morphology to the quench crystals of plagioclase described by Bryan (1972) in submarine basalts. They do not show any compositional zoning.

The sanidine was determined from a type II sample to be sodium-poor, with a high iron content; the percentage of iron feldspar molecule is close to 10 percent. Titanium was detected with the microprobe. The composition of the sanidine is similar to that reported by Carmichael (1967) for sanidines in a jumillite from Spain. The iron content of the Smoky Butte sanidine is significantly below that reported in the Leucite Hills feldspar (up to 18% iron feldspar; Carmichael, 1967) (Table 1).

Amphiboles

Table 2 lists representative microprobe analyses of amphiboles found in type I lamproite. They belong

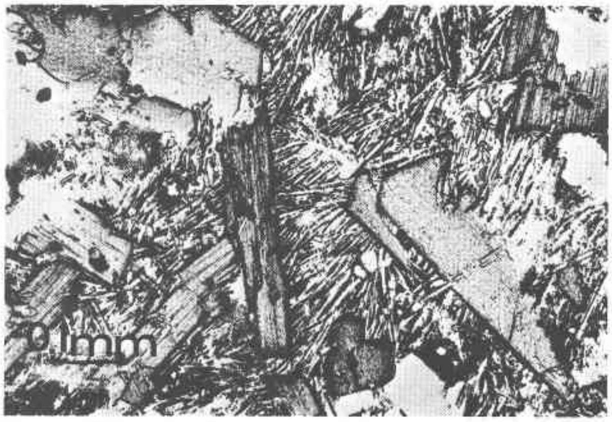

FIG. 1. Phlogopite phenocrysts and clusters of skeletal sanidine crystals in a type II lamproite.

for the most part to the potassium richterite group, but some are potassium riebeckites.

Potassium richterites have been reported from a number of lamproites (Prider, 1939; Hernandez-Pacheco, 1965; Carmichael, 1967) and from certain minettes (Velde, 1965; Nemec, 1973); the host rocks for both of these occurrences have very low sodium contents and hyperalkaline characteristics. Potassium richterites have also been reported in nodules found in kimberlites (Erlank and Finger, 1970; Aoki, 1974). Experimental work by Huebner and Papike (1970) has shown that complete solid solution exists between sodium and potassium richterite end-members at 1 kbar water pressure, between 775 and 850°C.

The potassium richterites from Smoky Butte are almost colorless in thin section, being slightly pinkish along β and very pale yellow along γ. This unusual optical characteristic can probably be related to their relatively low iron content compared to other reported members of the series. Potassium richterite appears here as a late crystallizing phase, occurring mostly in vugs. Sometimes it is rimmed by a thin zone of very strongly pleochroic amphibole (purple to dark prussian blue) towards the interior of the vug. Analyses reported in Table 2 indicate that this zone has the composition of a potassic riebeckite. This shows that potassium riebeckite can in fact exist in nature. However, the rarity of hyperalkaline potassic rocks, and the usual presence of calcium in these rocks, can explain why potassium riebeckite has not yet been reported. Its occurrence as rims on crystals in vugs suggests that it could have formed by precipitation from the vapor phase, similar to certain occurrences of acmite (see for example Smith and Lindsley, 1971).

Armalcolite

The name armalcolite (Anderson *et al*, 1970) has been proposed for a pseudobrookite-like mineral with the general formula $Fe^{2+}_{0.5}Mg_{0.5}Ti_2O_5$ found in a number of samples brought back by the Apollo 11 mission. Pseudobrookite is present in most samples from the Smoky Butte area. In type III rocks it occurs as euhedral crystals 0.03 by 0.3 mm (Fig. 2). These crystals are not opaque, showing a faint purplish tint in thin section and a weak birefringence. Table 3 gives the chemical analysis for the pseudobrookite from type III rock. The structural formula is:

$$(Ti_{1.80}Si_{0.03})(Al_{0.04}Cr_{0.02}Fe^{3+}_{0.20})(Fe^{2+}_{0.38}Mg_{0.57})O_5$$

Its general composition corresponds to that reported for lunar armalcolite, except for the presence of trivalent iron in the Smoky Butte specimen. The trivalent iron content of lunar armalcolite is thought to be zero, on the grounds that the mineral is in equilibrium—or at least coexists with—native iron. The mineral from Smoky Butte has low Fe_2O_3 and high MgO and TiO_2 content compared to most terrestrial pseudobrookites (Smith, 1965; Rice, Dickey, and Lyons, 1971; Anderson and Wright, 1972). It has, consequently, the closest composition to the lunar armalcolite thus far reported in terrestrial rocks. It could aptly be named ferri-armalcolite, rather than magnesium-rich ferro-pseudobrookite. Figure 3 shows the position of lunar armalcolite, Smoky Butte ferri-armalcolite, and Hawaiian pseudobrookite on a $(Ti,Si)O_2$—$(Fe,Mg,Mn)O$—$(Al, Fe^{3+}{}_2)O_3$ diagram (after Smith, 1965). The cell data, obtained from an X-ray powder diffractogram using a program written by C. W. Burnham (1962),

TABLE 1. Microprobe Analyses of Sanidine, Armalcolite, Ti-Phlogopite, Diopside, and Glass from Types II and III Lamproites

	Sanidine	Armalcolite	Ti-Phlogopite		Diopside	Glass
			Center	Edge		
SiO_2	65.20		38.33	40.79	54.72	55.93
Al_2O_3	16.53		11.40	8.89	0.17	8.87
Fe_2O_3	2.58*					
FeO		20.28*	5.78*	9.26*	3.59*	5.53*
MgO		11.68	19.18	17.25	17.56	5.25
CaO	0.01		0.02	0.03		0.78
Na_2O	1.14		0.28	0.34	0.30	1.10
K_2O	14.49		10.52	10.52		8.74
MnO		0.15				
TiO_2	0.33	67.56	11.27	11.05	0.99	2.21
Cr_2O_3		2.39				
Total	100.28	102.06	96.83	98.13	100.29	93.11**

* indicates total Fe calculated as Fe_2O_3 or FeO

** total includes 4.70% ignition loss, determined on a small sample of glass

TABLE 2. Microprobe Analyses of Alkali Amphiboles from Type I Lamproite Structural Formulas Computed following Czamanske and Wones (1973)

	Potassic Richterites		Potassic Riebeckites	
SiO_2	53.19	53.35	50.59	49.90
Al_2O_3	0.47	0.22	0.04	0.05
FeO	3.66*	4.09*	30.56*	29.50*
MgO	19.56	20.12	1.51	2.61
CaO	8.87	7.29	0.06	0.10
Na_2O	3.90	3.87	6.47	7.04
K_2O	3.96	4.16	4.76	4.71
TiO_2	3.45	3.43	4.50	3.71
Total	97.07	96.52	98.49	97.60
Si	7.626	7.676	8.014	7.973
Al	0.080	0.037		0.009
Ti	0.294	0.287		0.018
Σ (tet.)	8.000	8.000	8.014	8.000
Al			0.007	
Fe	0.439	0.492	4.048	3.941
Mg	4.180	4.314	0.356	0.622
Ti	0.078	0.084	0.536	0.428
Σ $(M_1\text{-}M_3)$	4.697	4.890	4.947	4.991
Ca	1.362	1.124	0.009	0.017
Na	0.638	0.876	1.988	1.983
Σ (M_4)	2.000	2.000	1.997	2.000
Na	0.445	0.204		0.197
K	0.724	0.764	0.963	0.959
Σ (A)	1.169	0.968	0.963	1.156

* indicates total Fe calculated as FeO

are quite comparable to those for lunar armalcolite or for synthetic armalcolite (Anderson *et al*, 1970). Values for Smoky Butte are a = 9.749(9) Å; b = 10.026(12) Å; c = 3.735(8) Å; V = 365.057(791) Å^3; for lunar armalcolite they are a = 9.743(30) Å; b = 10.024(20) Å; c = 3.738(30) Å; V = 365.077(619) Å^3, numbers in parentheses being the estimated standard errors.

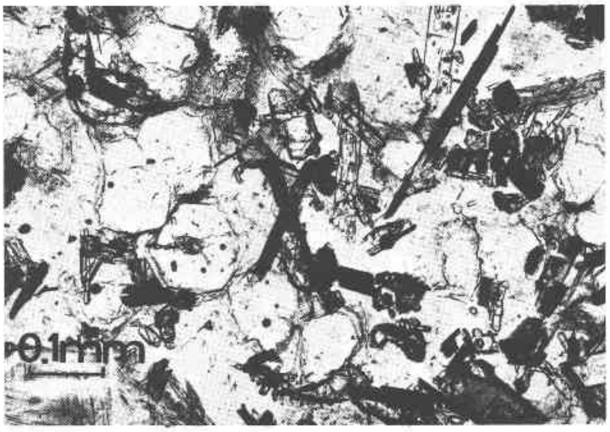

FIG. 2. Armalcolite crystals in a glassy analcite-rich groundmass of type III lamproite.

In lunar rocks (Haggerty *et al*, 1970) or in Hawaiian basalts (Anderson and Wright, 1972) pseudobrookite is always mantled by ilmenite, unless it occurs in another phenocryst. In the Smoky Butte intrusives, microprobe analyses for Fe, Mg, Ti, Cr, and Mn show armalcolite to be homogeneous and unzoned (a representative microprobe analysis is listed in Table 1).

The place of armalcolite in the crystallization sequence is of importance. In the Smoky Butte intrusives, armalcolite is found included in clinopyroxenes and in some mica phenocrysts, but never in the olivines. This indicates that it crystallized after olivine, and before some (if not all) of the Ti-phlogopite. In lunar rocks at atmospheric pressure, armalcolite is the first phase on the liquidus, followed by clinopyroxenes and then olivine (Akimoto *et al*, 1970). This same relation was also reported by O'Hara *et al* (1970) for compositions with 6.7 to 10 percent TiO_2. At pressures above 20 kbar, no Fe-Ti oxides are formed. Further, the Mg/Fe ratio of armalcolite decreases with decreasing temperature, from 0.81 at 1200°C to 0.59 at 1150°C at 1 bar; at 1125°C armalcolite is replaced by ilmenite (Akimoto *et al*, 1970). The Smoky Butte armalcolite, with a Mg/Fe ratio of 0.60, would then have crystallized at the lower pressure and temperature range of its stability in the chemical system in which it is found.

Analcite

Analcite occurs as euhedral microphenocrysts (Fig. 4) quite similar in aspect to the leucite in the Wyoming wyomingites (Carmichael, 1967). It is not birefringent, a fact which probably explains why it has been mistaken for leucite (Matson, 1960), since the Wyoming leucites are also isotropic. Analcite is found only in type II and type III rocks. In type II rocks, rich in sanidine, there is little analcite; however in type III samples the glass is abundant, and the analcite is fresh and homogeneous. It should be noted that, in Smoky Butte, analcite always coexists with glass.

Using Cu$K\alpha_1$ radiation, the *639* reflection for this analcite was displaced by 1.87° relative to the *331* reflection for the Si standard, indicating a Si-rich composition. This point falls outside the domain studied by Liou (Fig. 1, 1971) and would correspond to a crystallization temperature below 350°C. According to data given by Peters, Luth, and Tuttle (1966), the molar ratio SiO_2:Na_2O would be 4.5. Qualitative analysis with the electron microprobe indicates the absence of potassium.

As can be seen from the C.I.P.W. norms (Table 4), the Smoky Butte rocks are not undersaturated with respect to silica. Consequently they should contain alkali feldspar and no analcite, and indeed the holocrystalline rocks do not contain analcite. It would then appear that, in type III rocks, the analcite crystallized above the stability field of sanidine and quenching prevented the normal appearance of feldspar and any reaction of analcite with the liquid. From available experimental data (Peters *et al,* 1966) we know that analcite can coexist with a liquid at pressures above 5 kbar, and, according to very preliminary results by the same authors, above 3 kbar in potassium-bearing systems. The Smoky Butte analcite, however, is silica-rich, an unambiguous indication (Liou, 1971) of a low temperature of crystallization. It seems then logical to consider that a high temperature phase crystallized first but was replaced, at low temperature, by silica-rich analcite. This primary phase would have been cubic on the basis of morphological considerations. Its most likely identity is leucite, but it should be noted that in the glassy wyomingites leucite was not replaced by a low temperature phase, nor did it invert to the low temperature non-cubic polymorph of leucite (Carmichael, 1967). An unlikely possibility is the initial crystallization of a carnegieite. A description of carnegieite crystals formed in experiments (Bailey and Schairer, 1966) in a purely sodic system as "typical rounded octahedra" does not contradict this possibility.

Ti-Phlogopite

The phlogopite, perhaps the most obvious phase in hand specimen, is striking in thin section. Its pleochroism is strong, it is often zoned, and there seems to be a continuous variation in size from phenocrysts to small crystals. In the phenocrysts the edge is more intensely colored than the central part. The extinction angle can be as large as 5° and twinning with an (001) twin plane is frequent. Microphenocrysts show the same optical characteristics (Fig. 5). The bulk analysis of the phlogopite is listed in Table 3, and representative microprobe analyses of the outer and inner zones are presented in Table 1. The structural formula is:

$$(Na_{0.03}K_{0.73})O_{10}(OH)_2[Si_{2.75}Al_{0.99}Fe^{3+}_{0.21}Ti_{0.05}]$$
$$[Ti_{0.44}Fe^{2+}_{0.29}Mg_{2.04}]$$

It is outstanding because of its TiO_2 content of 9.36 weight percent. Carmichael reported relatively low TiO_2 values (from 1.8 to 2.1 percent) in the phlogopites from the Leucite Hills, but 5.9 percent in a Spanish jumillite mica, and as high as 7.8 percent in a west Kimberley fitzroyite (as compared with Prider's original figure of 8.97 percent). The contrast between Leucite Hills compositions and the Smoky Butte one is probably due to rock bulk composition, the Leucite Hills volcanics containing half as much TiO_2 as do those of Smoky Butte.

TABLE 3. Chemical Composition of Armalcolite, Ti-Phlogopite, and Diopside from Type III Lamproite

	Armalcolite	Ti-Phlogopite	Diopside
SiO_2	0.96	39.09	53.24
Al_2O_3	0.88	11.96	1.56
Fe_2O_3	7.45	4.04	2.25
FeO	12.65	4.95	2.67
MgO	10.53	19.49	16.37
CaO			20.90
Na_2O		0.24	0.31
K_2O		8.17	tr
MnO	0.02		0.15
TiO_2	65.99	9.36	1.42
P_2O_5		0.21	0.08
Cr_2O_3	0.67		
H_2O +		1.43	0.56
H_2O -		0.14	0.04
Total	99.15	99.08	99.55

The cell dimensions of the Smoky Butte Ti-phlogopite are: a = 5.323(7) Å; b = 9.163(3) Å; c =

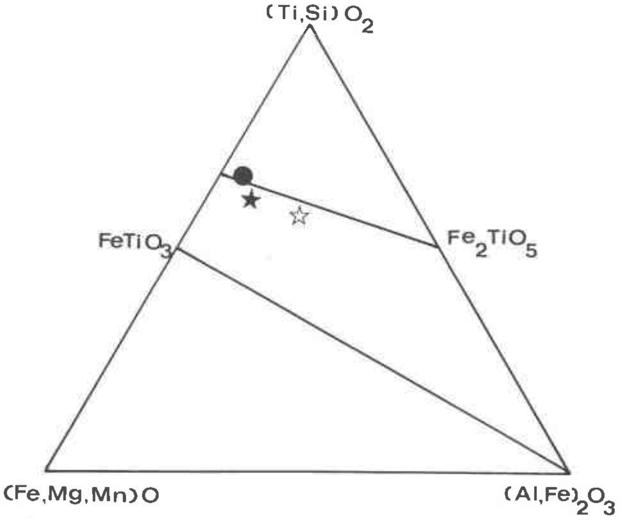

FIG. 3. Composition of Smoky Butte ferric armalcolite (solid star) compared to a lunar armalcolite (solid circle) analysis from Haggerty *et al.* (1970) and to a Hawaiian ferropseudobrookite (open star) analysis (FPSB 13 in Anderson and Wright, 1972).

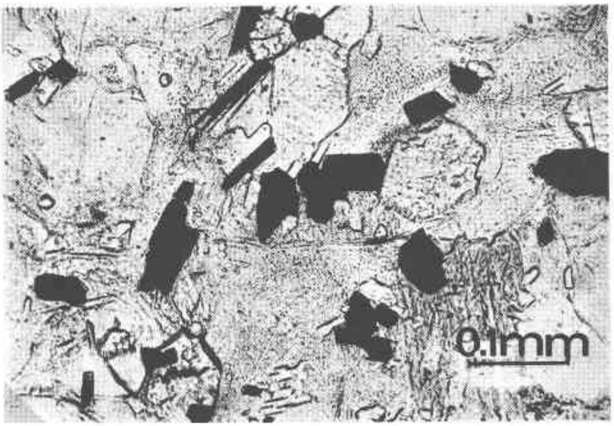

FIG. 4. Analcite crystals in a type III lamproite. Clear prisms of diopside, armalcolite (black) and groundmass phlogopite are also visible.

19.985(9) Å; β = 94°54′(38′); V = 971.2(1.2) Å^3, numbers in parentheses representing the estimated standard error. These values are comparable with those reported by Yoder and Eugster (1954) for a natural 2*M* phologopite. However, they are small compared with a number of dimensions reported for natural phlogopites from lamproites (Velde, 1969).

The only other micas thus far reported with such high titanium contents are phenocrysts in basaltic lavas (Flower, 1969; Velde and Coulon, unpublished), or crystals in mica-garnet lherzolite included in an ankaramitic lava (Dawson, Powell, and Reid, 1970). In these instances, phlogopite phenocrysts can be interpreted as high pressure phases. The conditions of crystallization for the latter mica were estimated at 1109°C and 24.2 kbar (Dawson *et al*, 1970).

TABLE 4. Chemical Compositions and C.I.P.W. Norms of Smoky Butte Lamproites

	Type I	Type II	Type III
SiO_2	52.19	53.53	51.89
Al_2O_3	9.04	9.08	9.05
Fe_2O_3	5.02	4.39	3.86
FeO	3.50	1.02	1.31
MgO	7.98	7.78	7.90
CaO	4.90	4.43	4.44
Na_2O	1.14	1.93	1.97
K_2O	7.85	7.85	7.63
MnO	0.08	0.08	0.08
TiO_2	4.96	5.23	5.17
P_2O_5	0.22	0.22	0.25
BaO	0.69	0.77	0.88
H_2O +	0.95	2.69	4.05
H_2O -	0.54	1.72	1.27
Total	99.06	100.72	99.75
Cu	25	28	22
Ni	600	415	405
Zn	87	87	82
Sr	2,700	2,790	2,950

	Type I	Type II	Type III
Q	0.22		
Or	46.39	46.39	45.09
Ab	2.79	3.00	4.07
Ac	6.04	11.74	11.10
Di	15.26	5.24	6.18
En	12.80	16.95	16.81
He	2.93	0.33	0.02
Il	7.56	2.33	2.94
Tn	2.40	7.38	2.04
Pf		1.70	4.75
Ap	0.48	0.48	0.55

Cu,Ni,Zn and Sr expressed in parts per million

Titanium substitutions in phlogopite have recently been studied experimentally. Robert (1973) found that the 2 Si^{IV}, Mg^{VI} = 2 Al^{IV},Ti^{VI} substitution is maximum at low pressures (1 kbar) and relatively high temperatures (1000°C). Forbes and Flower (1974) investigated the substitution 2 Mg^{VI} = Ti^{VI}, □, and deduced that it is favored by high pressures and temperatures (up to 30 kbar). Figure 6 is an Al-Fe (total iron) + Mg-Ti plot of the available analyses of titanium-bearing phlogopites, including those of the Smoky Butte phlogopites (from Tables 1 and 3) and five other analyses of their grain centers and edges. The two trends corresponding to the two types of substitutions mentioned above have also been indicated. It is obvious that these two substitutions can account for the compositions of the micas found in basalts and in the garnet lherzolite, but they alone cannot explain the composition of the low alumina phlogopites which seem to define a series roughly parallel to the join phlogopite–$K_2Mg_4TiAl_2Si_6O_{20}(OH)_4$. In the series of natural phlogopite compositions, Si is close to 6 ions per formula, but alumina is always lower than 2, even in titanium-poor phlogopites. It is necessary then to consider the substitution of iron for tetrahedral silicon. The formula $K_2(Mg,Fe^{2+})_6Al_{1.6}Fe^{3+}_{0.4}Si_6O_{20}(OH)_4$ would approximately represent the titanium-free end of the series. The titanium-rich micas (the Smoky Butte phlogopites for example) are also low in silicon (5.5 ions per structural formula) and in this case, since iron content is low, titanium probably substitutes also for silicon.

These alumina-deficient mica compositions have not yet been studied experimentally, so that nothing can be said concerning their stability. However, judging from their existence as phenocrysts in nodules in kimberlites (Aoki, 1974) and in hypabyssal rocks they should be stable both at high and low pressures.

Clinopyroxenes

Clinopyroxenes appear as euhedral crystals which sometimes contain glass inclusions. Chemical analysis by wet chemical methods shows their composition to be close to that of diopside, quite comparable to the analyses reported for clinopyroxenes from the Leucite Hills volcanics (Carmichael, 1967). The structural formula corresponding to the analysis listed in Table 3 is:

$$[Al_{0.02}Fe^{3+}_{0.06}Fe^{2+}_{0.08}Mg_{0.90}Ca_{0.82}Ti_{0.04}Na_{0.02}]\\(Si_{1.96}Al_{0.04})O_6$$

Microprobe analyses (see Table 1) indicate that the Si content would in fact be closer to 2 ions, and the alumina content is slightly overestimated in the wet chemical analysis.

Glass

Attempts have been made to determine the composition of the yellow glass present in type III rock with the electron microprobe. Results have shown consistently low totals (around 90%). Although reproducible, they cannot be considered totally accurate. Loss on ignition was determined on an isolated fraction of the glass and found to be 4.7 percent. It is possible that the glass is enriched in elements such as barium and strontium, which were not sought for but are abundant in the whole rock.

A representative composition (the glass proved to be very homogeneous) is reported in Table 1. It differs from that of a wyomingite glass determined by Carmichael mostly in its potassium and calcium content. Leucite is a phenocryst phase in the wyomingite, and as a result the glass is depleted in potassium relative to the bulk rock (2.69% against 9.69%). On the other hand, the crystallization of diopside in the Smoky Butte rocks depleted the glass in CaO, whereas wyomingites contain no diopside phenocrysts and the glass is calcium-rich.

Petrographic Descriptions

The *type I* rock is extremely fine-grained and inhomogeneous. It is gray in hand specimen with no mineral distinguishable with the naked eye. Inclusions of sedimentary rocks (less than a centimeter in diameter) are abundant. They are mostly sandstones with varying amounts of a phyllite-rich matrix. These inclusions never show any alteration or metamorphism.

Microscopically, the only visible phases are poecilitic, poorly formed crystals of phlogopite, quartz grains, barite, calcite, and alkali amphiboles. The amphibole, for the most part potassium richterite, occurs in vugs; potassium riebeckite forms the apices of the richterite crystals, or small independent prisms in the fine-grained groundmass of the rock. The vugs are filled by calcite, barite, and quartz in variable proportions. Armalcolite is abundant as minute crystals; priderite (identified by its optical properties) accompanies the amphiboles. Sanidine, forming the greatest part of the rock, was identified by X-ray diffractograms, which also indicate the absence of analcite.

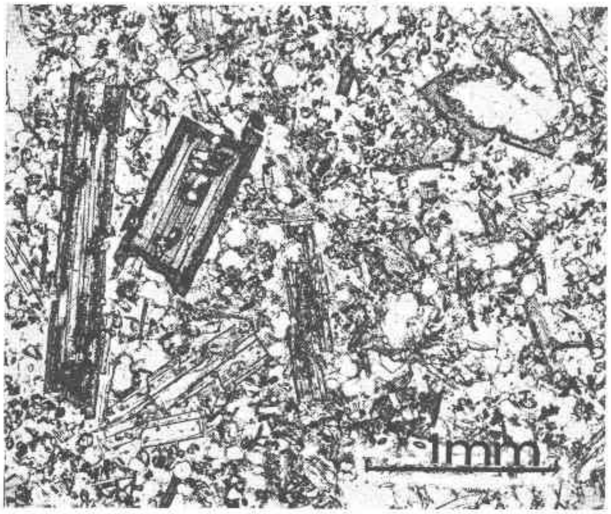

FIG. 5. Zoned phlogopite, with inclusions, in an analcite-rich groundmass of type III lamproite.

The *type II* rock is composed of phlogopite, diopside, altered olivine, armalcolite, and abundant elongated crystals of sanidine (Fig. 4) which, with some yellowish glass, form most of the groundmass. Pseudomorphs of similar shape and size as analcite crystals in type III are present. They resemble,

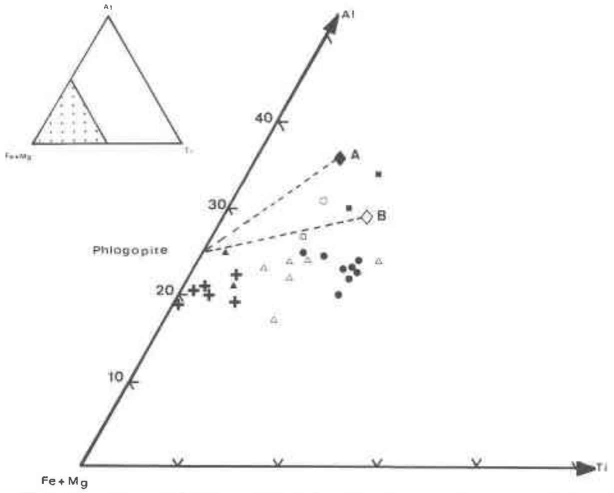

FIG. 6. Plot of observed natural phlogopite compositions comparing Al-Ti-(Total Fe + Mg) atomic proportions. Line A is the substitution studied experimentally by Robert (1973), line B that studied by Forbes and Flower (1974). Crosses are micas from kimberlite nodules (Aoki, 1974); triangles, micas from lamproites (Carmichael, 1967), solid triangles representing phlogopites from the Leucite Hills; solid circles, phlogopites from Smoky Butte; open square is a mica from a basalt (Flower, 1969); open circle represents 4 analyses of mica phenocrysts in a sardinian basalt (Velde and Coulon, unpublished); solid squares are phlogopites from a garnet lherzolite (Dawson *et al*, 1970).

however, the altered leucite crystals described in west Kimberley lamproites (Wade and Prider, 1940). They are mostly composed now of mixtures of sanidine, montmorillonite, and minor analcite, the turbid appearance noted in the Kimberley specimens being quite conspicuous.

The *type III* rock contains olivine, usually completely replaced by pale green montmorillonite, zoned phlogopite, clinopyroxene, analcite, and armalcolite in a pale yellow glass. A few sanidine crystals can be found with the usual quench-crystal morphology. The calculated mode, using a least squares program following the method described by Bryan, Finger, and Chayes (1969) gives the following percentages (weight percents): diopside, 13; analcite, 10; phlogopite, 13; armalcolite, 3; and glass, 61.

Other rock types were found in thin sections loaned by the Department of Geology of Montana State University. Exceptional among them is a rock made of potassium richterite, phlogopite, priderite, armalcolite, and sanidine with, possibly, some wadeite. Such a rock should be somewhat richer in alkalis (*i.e.,* potassium) and poorer in alumina than the ones analyzed here, but specimens were not available for study.

Table 4 gives chemical analyses, trace element data, and C.I.P.W. norms for the three rock types that have been described. It is obvious when one considers the mineralogical compositions listed above, and the chemical data, that such rocks can only be grouped under the general term of lamproites, that is volcanic or hypabyssal rocks with an alkali- (mostly potassium) and magnesium-rich composition (Niggli, 1923; Wade and Prider, 1940, p. 50). The Montana rocks are considerably richer in TiO_2 and poorer in K_2O than the Leucite Hills volcanics. Their chemical composition is close to that reported for the Australian lamproites, but the differences in mineralogical compositions preclude the use of the specific names defined in west Kimberly.

Conclusions

The Smoky Butte area represents another occurrence of peralkaline potassic eruptive rocks. Though similar in many respects to the volcanics described in the Leucite Hills, they differ on the following grounds: high or very high TiO_2 contents, expressed by the presence of Fe-Ti oxides; abundance, in certain types, of an Si-rich analcite, and absence of leucite; absence of any basic types that could be compared to members of the madupite-orendite series of Wyoming (Carmichael, 1967).

In spite of these differences, any satisfactory explanation of the origin of the Leucite Hills volcanics could probably be applied with little, if any, change to the Montana rocks. Available hypotheses have been reviewed at length by Carmichael (1967) and subsequent isotopic studies of all known examples of potassium-rich lamproites and Montana diatremes as well (Powell and Bell, 1970) have failed to provide any better lead in determining the origin of lamproites.

One should remember, however, the relatively close spatial association of the Smoky Butte volcanics with the ultramafic rocks of kimberlitic affinities (Hearn, 1968). The latter are characterized by phenocrysts of forsteritic olivine in a groundmass of monticellite and/or melilite, nepheline, phlogopite, perovskite and opaques. Unless contaminated, they do not contain any clinopyroxene and their potassium-to-sodium ratio is always higher than one.

It is difficult to propose any kind of precise mechanism by which such an assemblage, so extremely undersaturated with respect to silica, would produce a relatively low-temperature melting fraction, enriched in silica and peralkaline in character. Such a mechanism could nevertheless exist, but all we can say at the moment is that such liquids are not likely to be generated in the crust. At Smoky Butte their origin would then have to be very deep indeed, since the crust in this particular zone of the western United States is at least 45 kilometers deep (Asada and Aldrich, 1966).

Acknowledgments

I am very grateful to Dr. John Hower, Case Western Reserve University, who gave me the opportunity to study rocks sampled some 7000 miles away from my present residence, to R. E. Matson, Montana Bureau of Mines, who kindly sent me specimens from the Smoky Butte quarry, and to G. Thompson, Montana State University, who loaned me a number of thin sections from various parts of the outcrop.

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Manuscript received, June 26, 1972; accepted for publication, February 4, 1975.