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Petrogenesis of the ultrapotassic rocks from the Leucite Hills, Wyoming

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Abstract

The ultrapotassic rocks of the Leucite Hills, Wyoming consist of silica-undersaturated madupites; silica-saturated to oversaturated wyomingites, orendites and olivine orendites; and "mixed" rocks with characteristics of both wyomingites and madupites. Based on textural relations and on major and trace element analyses of the rocks and their major minerals, phlogopite-rimmed olivines, some of the spinel group minerals and the salites are considered to be xenocrysts. Although consistent trends in the major elements suggest a genetic relationship among the rock types, the high Cr, Ni and 100 Mg/(Mg+Fe) ratios preclude fractional crystallization as a major mechanism in their genesis. Comparison of the compositions of the rocks with high pressure phase relations in the systems Mg_2SiO_4–$KAlSiO_4$–SiO_2 and $KAlSiO_4$–MgO–SiO_2–H_2O–CO_2 indicates that madupite liquid can be generated at pressures between about 24–34 kbar (94–120 km) depending on conditions of melting from a K-enriched peridotitic mantle source. The wyomingite and orendite magmas may be produced from the same source at pressures between approximately 14 and 19.5 kbar (49–68 km) depending on conditions of melting. By fractionation of olivine, diopside and phlogopite, an olivine orendite cumulate may be produced from an orendite liquid.

Introduction

The petrogenesis of ultrabasic to intermediate rocks with $K_2O > Na_2O$ has attracted the attention of many petrologists. Such rocks are distinguished by K_2O ranging from 2–19 wt.%, K_2O/Na_2O often greater than 3, high concentrations of LIL elements (K, Rb, Ba, Zr, etc.), and high molecular Mg/(Mg+Fe) and weight percent Fe_2O_3/FeO ratios. Rocks with exceptionally high K_2O values are commonly referred to as ultrapotassic; the best known examples being those of the Leucite Hills, Wyoming; Jumilla, Spain; the West Kimberley region, Western Australia; the Toro-Ankole region, southwest Uganda; and Gaussberg, Antarctica. Although the majority of ultrapotassic rocks are silica-undersaturated, Q-normative rocks occur in the Leucite Hills and some other localities. Ultrapotassic rocks are generally peralkaline and some have $K_2O > Al_2O_3$. In contrast to most other localities, the ultrapotassic rocks of the Leucite Hills have no associated volcanic rocks of less extreme compositions.

A perplexing problem is that of reconciling the enrichments of K and other LIL elements in the ultrapotassic rocks, such enrichments being comparable to, or greater than, the values normally found in more siliceous and highly differentiated rocks with much lower Mg/(Mg+Fe) ratios. Bell and Powell (1969) and Sahama (1974) summarize the hypotheses for the genesis of these rocks. On both geological and chemical grounds, hypotheses of assimilation and fractionation at sub-crustal depths seem unlikely inasmuch as basalts and other rocks, such as eclogites, inherent in these hypotheses, are absent from most ultrapotassic rock localities. Recent geochemical and experimental studies (Edgar *et al.*, 1976; Turi and Taylor, 1976; Ryabchikov and Green, 1978; Hawkesworth and Vollmer, 1979; Wendlandt and Eggler, 1980a,b) favor a partial melting hypothesis for these rocks, with the source being enriched in K.

Cross (1897), Kemp (1897), and Kemp and Knight

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(1903) first described, named, and analyzed the rocks of the Leucite Hills. Yagi and Matsumoto (1966) suggested that these rocks were derived by partial melting of a mica peridotite with no assimilation of sialic material. Based on an extensive geochemical study of the rocks and their minerals, Carmichael (1967) concluded that sialic contamination played little part in the genesis of the saturated to oversaturated wyomingite–orendite rocks and that the undersaturated madupite could be derived by high pressure fractionation of an eclogitic assemblage from a liquid produced by partial fusion of a garnet peridotitic mantle. Carmichael was unable to distinguish any genetic relationship between the silica-rich wyomingites and orendites and the silica-poor madupites. Kay and Gast (1973), using REE data, suggested that the madupite could be derived by very small degrees of partial melting of a garnet lherzolite and the wyomingites and orendites by very slightly higher amounts of partial melting of the same source material.

Experimental studies up to 30 kbar and with about 3 wt.% H_2O (Barton and Hamilton, 1979) indicate that madupite may be a product of partial melting of a metasomatized upper mantle with phlogopite being fractionated prior to segregation of the magma. Based on experiments up to 5 kbar, Barton and Hamilton (1978) concluded that wyomingite and orendite could not be related to madupite by low pressure fractional crystallization.

The close spatial association of the madupites, wyomingites, orendites, olivine orendites, and a series of rocks that appear to be intermediate between the wyomingites and madupites, as well as the lack of other volcanic rocks, has prompted us to reexamine the Leucite Hills ultrapotassic rocks in an attempt to find a genetic relationship between these rock types. Our model is based on detailed petrographic and geochemical studies of these rocks and on published phase relations in appropriate systems. The nomenclature and classification of the rocks used in this study are those of previous workers and have no genetic connotations.

Geology and petrography

Regional geology of the Wyoming Basin, in which the Leucite Hills are located, has recently been discussed by Prodehl (1976), Matthews and Work (1978), Stearns (1978), Stewart (1978), and Smithson *et al.* (1979). The Leucite Hills rocks are 1.1 m.y. in age (MacDowell, 1966) and are probably the youngest potassium-rich rocks in the Western U.S.A.

The petrography of the Leucite Hills volcanics has been described elsewhere (Cross, 1897; Kemp, 1897; Schultze and Cross, 1912; Carmichael, 1967; Kuehner, 1980). A synopsis of our observations of the thin sections of the main rock types is given in Table 1, together with two rocks with petrographic characteristics similar to both the madupite and wyomingite. The latter rocks contain analcime and other zeolites indicative of extensive secondary alteration. These rocks, referred to as transitional rocks for the remainder of this paper, are similar to the rocks described by Kemp and Knight (1903).

Geochemistry

X-ray fluorescence methods were used for the major and trace element analyses of the rocks and for trace element analyses of minerals which could be separated in sufficient quantities and with a high degree of purity (>95 percent). Ferrous–ferric ratios were determined titrimetrically, F and CO_2 by ion-sensitive electrode and gravimetric methods respectively. An automated MAC-400 microprobe was used for the major element analyses of the minerals. Details of analytical procedures are given by Kuehner (1980).

Mineral analyses

Olivines (Table 2). As shown in Table 1, olivines occur as both phenocryst and groundmass phases in the olivine orendites and in the groundmass of the orendites. Using the distribution coefficient of calcium between olivine and liquid ($D_{Ca}^{Ol/L}$) determined by Watson (1979) at 1 atm, the groundmass olivines could have crystallized from a liquid whose composition is comparable to the rocks in which they occur. A similar calculation for the phlogopite-rimmed olivine phenocrysts (Table 1) indicate they could be in equilibrium only with a liquid containing <1% CaO. Therefore these olivines may not have crystallized from a liquid represented by the rock in which they are now found as this rock contains 4.21 wt.% CaO (Table 8). Similarily, using the Mg and Fe^{2+} partition coefficients (K_D^{Fe-Mg}) determined at atmospheric pressure (Roeder and Emslie, 1970), the groundmass olivines could have crystallized from a liquid corresponding to the rocks in which they occur, whereas the phlogopite-rimmed olivines (Table 1) could not. The effects of pressure on the $D_{Ca}^{Ol/L}$ and the K_D^{Fe-Mg} between olivine and liquid have not been determined experimentally. However, Stormer (1973) has calculated that the $D_{Ca}^{Ol/L}$ should decrease with increasing pressure, and Mysen (1975) has shown that the K_D^{Fe-Mg}

Table 1. Mineralogy and paragenesis of rock types

Olivine Orendite	Orendite	Wyomingite	Transitional	Madupite
(chromites)** -resorbed, rimmed by phlogopite	(apatite) -resorbed grains	(hercynite) -resorbed grains	(hercynite) -resorbed grains	DIOPSIDE MP-G -twinning very common
(apatite) -resorbed grains	(dark phlogopite cores) -resorbed grains	apatite MP -subhedral grains	APATITE MP -microphenocrysts, euhedral	apatite
(dark phlogopite cores) -resorbed grains	PHLOGOPITE MP -microphenocrysts strongly resorbed	(dark phlogopite cores) -resorbed grains	(dark phlogopite cores) -resorbed grains	(leucite)
(olivine) P -rimmed by phlogopite	olivine G -may form reaction rim around phlogopite	(pale green diopside cores) -resorbed grains	(pale green diopside cores) -resorbed grains	perovskite G
PHLOGOPITE†† MP -microphenocrysts strongly resorbed	DIOPSIDE MP -twinning rare	PHLOGOPITE MP -microphenocrysts	PHLOGOPITE MP -microphenocrysts	wadeite G
olivine† G -may form reaction rim around phlogopite	apatite	DIOPSIDE MP -twinning not rare	DIOPSIDE MP -twinning very common	phlogopite G -ragged poikilitic masses
DIOPSIDE MP -twinning rare	SANIDINE G	apatite	(leucite)	(euhedral opaques) -found only in phlogopite
apatite	priderite -amber color	LEUCITE G	perovskite G	(skeletal opaques)
SANIDINE G	(wadeite)	(priderite) -dark brown in color	phlogopite G -ragged poikilitic masses	GLASS -dirty green color
priderite -amber color	LEUCITE G	(phlogopite) -narrow rims on the phlogopite phenocrysts	euhedral opaques -only found in phlogopite	
(wadeite)	RICHTERITE G	richterite G	GLASS -dirty green color	
LEUCITE G		(sanidine)		
RICHTERITE G		glass		

* Minerals listed in order of crystallization sequences
** Rare minerals in brackets
† Common minerals unbracketed
†† Major minerals in capitals

P - phenocryst
MP - microphenocryst
G - groundmass

for olivine and liquid may be dependent on pressure temperature , fO_2 and composition. Calculations of $D_{Ca}^{Ol/L}$ and K_D^{Fe-Mg} values for the phlogopite-rimmed olivines with those of the liquids from which they supposedly crystallized vary by about 5 times. This suggests that these olivines may be xenocrysts rather than phenocrysts. If the phlogopite-rimmed olivines are xenocrysts they may have been derived from an ultramafic mantle. The chromite inclusions in such olivines (Table 1) would therefore also be xenocrysts.

Pyroxenes (Table 3). The pyroxenes are predominantly diopside, although those from some wyomingites have cores of salite sharply bounded by diopside rims. This is a fairly common feature of potash-rich rocks, *e.g.*, the pyroxenes in the minettes from the Navajo field, Arizona (Roden and Smith, 1979) and those from the mafurites of southwest Uganda (Holmes, 1942). Based on high pressure experimental studies on one of the latter rocks (Edgar *et al.*, 1976) it seems unlikely that the reverse zoning involving salite and diopside is indicative of a difference in the depths of crystallization. Calculations, based on the Fe–Mg distribution coefficients (Thompson, 1974) between Ca-rich pyroxenes and

Table 2. Representative olivine analyses

	1	2	3	4
SiO_2	41.84	40.41	40.71	38.55
FeO*	7.14	9.73	9.85	12.27
MgO	50.82	48.77	49.18	47.56
CaO	0.03	0.15	0.11	0.62
NiO	0.08	0.18	0.21	0.25
MnO	-	0.33	0.38	0.61
Σ	99.91	99.57	100.44	99.86
$\frac{Mg}{Mg + Fe}$	93	90	90	87
Number of Cations on the basis of 4 Oxygens				
Si	1.001	0.981	0.994	0.965
Fe	0.145	0.201	0.201	0.258
Mg	1.830	1.789	1.793	1.777
Ca	-	0.004	0.002	0.016
Ni	0.002	0.004	0.002	0.004
Mn	-	0.008	0.004	0.012

* Total iron determined as FeO.
1 From olivine orendite (SK 36), South Table Mt., olivine rimmed by phlogopite, average of 4 analyses.
2 From orendite (SK 35), North Table Mt., groundmass olivine, average of 2 analyses.
3 From olivine orendite (SK 36), groundmass olivine, average of 2 analyses.
4 From orendite (SK 37), Spring Butte, single analysis of groundmass olivine.

Table 3. Representative pyroxene analyses

	Major Elements									
	1a	1b	1c	2a	2b	3	4	5	6	7
SiO_2	54.11	53.52	54.17	54.55	53.63	53.81	53.88	54.20	54.19	51.84
TiO_2	0.63	0.84	0.92	0.79	1.24	0.70	0.52	0.51	0.82	0.19
Al_2O_3	0.25	0.37	0.48	0.38	0.83	0.17	0.12	0.05	0.19	1.59
Cr_2O_3	0.02	0.00	0.00	0.01	0.11	0.02	0.00	0.00	0.00	0.03
FeO	3.15	3.04	2.56	2.31	2.63	2.48	3.31	2.08	2.81	10.89
MnO	0.08	0.12	0.13	0.14	0.09	0.14	0.21	0.00	0.08	0.42
CaO	25.19	24.51	25.06	25.45	25.00	23.20	22.90	24.96	24.85	22.50
MgO	16.43	16.45	17.05	16.61	16.17	18.28	17.72	17.92	17.34	12.05
Na_2O	0.20	0.32	0.18	0.16	0.38	0.22	0.20	0.20	0.30	0.47
K_2O	0.01	0.01	0.02	0.09	0.21	0.05	0.07	0.01	0.05	0.02
Σ	100.07	99.18	100.57	100.49	100.29	99.07	98.93	99.93	100.63	100.00
	Number of Cations on the Basis of 6 Oxygens									
Si	1.973	1.965	1.960	1.973	1.947	1.966	1.978	1.967	1.959	1.950
Al^{IV}	0.011	0.016	0.020	0.016	0.036	0.007	0.005	0.002	0.008	0.050
Al^{VI}	-	-	-	-	-	-	-	-	-	0.020
Fe^{3+}	0.038	0.053	0.042	0.042	0.066	0.049	0.041	0.050	0.060	0.055
Ti	0.017	0.023	0.035	0.022	0.034	0.019	0.014	0.014	0.022	0.005
Cr	0.001	0.000	0.000	0.000	0.003	0.001	0.000	0.000	0.000	0.001
Mg	0.893	0.900	0.919	0.896	0.875	0.996	0.970	0.969	0.934	0.676
Fe^{2+}	0.058	0.040	0.035	0.028	0.014	0.027	0.061	0.013	0.025	0.288
Mn	0.002	0.004	0.004	0.004	0.003	0.004	0.007	0.000	0.002	0.013
Ca	0.984	0.964	0.971	0.987	0.972	0.908	0.901	0.970	0.962	0.906
Na	0.014	0.023	0.013	0.011	0.027	0.016	0.014	0.014	0.021	0.034
K	0.000	0.000	0.000	0.004	0.010	0.003	0.003	0.000	0.002	0.001

	Trace Elements (ppm)						
	8	9	10	11	12	13	14
Cr	481	336	88	241	908	716	1188
Ba	3884	1855	2256	1563	2847	1856	1454
Ni	220	157	114	162	106	135	123
Nb	6	9	8	-	25	14	40
Zr	447	536	581	530	624	410	689
Y	12	15	10	20	7	1	7
Sr	2429	2531	2068	2196	2875	2013	2612
Rb	39	37	30	15	31	24	30
Pb	2	21	24	-	11	12	23
Co	35	40	40	33	30	22	26

1 From transitional rock (SK 32), Twin Rocks, (a) core of pyroxene grain, (b) intermediate between core and rim of pyroxene grain, (c) rim of pyroxene grain.

2 From orendite (SK 35), North Table Mtn., (a) core of pyroxene grain, (b) rim of pyroxene grain.

3 From olivine orendite (SK 36), South Table Mtn.

4 From olivine orendite (SK 36), South Table Mtn.

5 From wyomingite (SK 9), Zirkel Mesa.

6 From wyomingite (SK 9), Zirkel Mesa.

7 From wyomingite (SK 41), Spring Butte.

8 From wyomingite (SK 9), Zirkel Mesa (different sample from 5 and 6).

9 From wyomingite (SK 10), Emmons Mesa.

10 From wyomingite (SK 41), Spring Butte (different sample from 7)

11 From olivine orendite (SK 36), South Table Mtn., (different sample from 3 and 4).

12 From transitional rock (SK 32), Twin Rocks (different sample from 1).

13 From madupite (SK 28), Pilot Butte.

14 From madupite (SK 22), Pilot Butte.

+- Fe^{3+} calculated by assuming a coupled substitution of the type ^{VI}Ti-$2^{IV}Al$, ^{VI}Ti-$2^{IV}Fe^{3+}$, and ^{VI}Na-$^{VI}Fe^{3+}$.

liquids, suggest that the salites crystallized from a much more Fe-rich liquid than did the diopsides. Texturally, the diopsides in the wyomingites appear to have crystallized with, or after, the phlogopite (Table 1), suggesting that prior to phlogopite crystallization the liquid may have had a low FeO/MgO ratio and therefore could not have crystallized salite. Whether the salites represent accidental xenocrysts, as suggested by Roden and Smith (1979) for the Navajo rocks, or are the result of magma mixing (cf. Brooks and Printzlau, 1978) could not be determined.

The trace element analyses of the pyroxenes (Table 3), although variable, indicate concentrations of Cr and Nb are high and those of Y and Co are low in the pyroxenes of the madupites and transitional rock relative to those in the wyomingites.

Phlogopites (Table 4). Figure 1 illustrates the variations in phlogopite compositions in the madupites and transitional rock relative to those in the wyomingites and olivine orendites. Phlogopites from wyomingites, olivine orendites, and phlogopite cores in the transitional rock are generally richer in Al_2O_3 and poorer in TiO_2 than those from the madupites and rims and groundmass phlogopites in the transitional rock (Fig. 1a). Phlogopites from the wyomingites have higher Cr_2O_3 contents than the phlogopites from madupites and olivine orendites. The dark colored cores, often found in phlogopites from the olivine orendites, are enriched in Ti, Ba, Al, and Fe and depleted in Mg and Si relative to the phlogopite microphenocrysts in the same rocks.

Concentrations of Co, Pb, Y, Zr, Rb, Nb, and Ni in the phlogopites from the wyomingites, orendites, and olivine orendites are similar (Table 4) indicating that the liquids from which these phlogopites crystallized may also have had similar concentrations of these elements. No trace element analyses of the phlogopites rimming olivines in the olivine orendites were obtained. Assuming that $D_{Ni}^{Ol/L}$ is greater than $D_{Ni}^{Ph/L}$ and that the same relationship applies to Co, crystallization of olivine directly from a liquid of olivine orendite composition should preferentially partition Ni and Co into the olivine, thus depleting the later crystallizing microphenocryst phlogopites in these elements. However the similarities in the Ni and Co contents of the microphenocrysts of phlogopite in the olivine orendites with those in the wyomingites support the concept of the phlogopite-rimmed olivines in the olivine orendites being xenocrysts.

The lower Cr contents of phlogopites from the olivine orendites and orendites than those from phlogopites in the wyomingites (Table 4) may be a result of early crystallization of chromite from the orendite liquids. Chromite is rimmed by phlogopite in the orendites and olivine orendites but not in the wyomingites. The lower Cr in the rims of the phlogopites in the wyomingites relative to their cores may mark the onset of crystallization of diopside in the wyomingites. Petrographically diopside appears to have crystallized along with, or after, phlogopite (Table 1). Since $D_{Cr}^{Di/L} > 1$ (Gast, 1968), the drop in Cr concentration between core and rim in the phlogopites may reflect the preferential incorporation of Cr into the diopsides.

Leucites (Table 5). Leucites in the Leucite Hills rocks have high Fe and low Al contents relative to most leucites (cf. Cross, 1897; Carmichael, 1967). Sufficient leucite for trace element analyses could only be separated from three of the wyomingites. Impoverishment in Rb and enrichment in Sr and Ba relative to the whole rocks (Table 8) support the petrographic conclusions (Table 1) that the leucites are late-crystallizing minerals (cf. Henderson, 1965).

Sanidines. Analyses of sanidines are very similar to those given by Carmichael (1967), being rich in Fe and poor in Na. No compositional differences were detected in the sanidine microphenocrysts occurring in the orendites and olivine orendites from the very minor amounts of sanidines in the wyomingites. The latter sanidines are believed to be products of reaction between leucite and liquid as sanidines are absent in glassy varieties of wyomingite which contain more abundant leucite.

Richterites. Richterites from the orendites, olivine orendites and wyomingites have similar compositions, comparable to those given by Carmichael (1967).

Apatites and Perovskites (Table 6). Inclusions of apatite in the phlogopite from the wyomingite have the same composition as the anhedral microphenocrysts of apatite in the same rock, and both are similar to the euhedral apatite in the transitional rock. Apatites in the madupite, however, have higher Nd_2O_3 and lower P_2O_5 contents.

Perovskite in the transitional rock (Table 6) has the same REE chemistry as perovskite in the madupite (Carmichael, 1967, Table 11).

Oxides (Table 7). Five types of spinels (two varieties of magnesiochromite, two types of titanomagnetite, and an Fe-rich spinel close to pleonaste) occur in

Table 4. Representative phlogopite analyses

	1	2	3a	3b	4	5a	5b	6	7
Major Elements									
SiO_2	42.14	43.06	41.28	41.81	42.46	41.66	41.29	41.18	39.23
TiO_2	2.06	1.86	2.15	2.20	3.70	2.14	4.66	4.88	5.14
Al_2O_3	11.38	11.30	11.89	11.52	7.72	12.29	9.31	9.21	13.91
Cr_2O_3	0.08	0.34	0.89	0.66	0.00	1.40	0.07	0.06	0.04
FeO	3.00	2.76	2.96	3.18	5.64	2.98	5.91	5.69	6.86
MnO	0.00	0.00	0.07	0.00	0.02	0.06	0.10	0.16	0.05
MgO	25.84	25.47	24.85	25.44	23.56	23.97	23.35	22.87	20.23
BaO	0.33	0.27	0.52	0.48	1.47	0.45	2.74	2.42	1.01
Na_2O	0.00	0.00	0.13	0.22	0.62	0.09	0.27	0.26	0.04
K_2O	10.41	10.69	10.61	10.53	10.16	10.71	9.76	9.75	10.50
Σ	95.24	95.75	95.35	96.04	95.35	95.75	97.46	96.48	97.01
Number of Cations on the Basis of 22 Oxygen†									
Si	5.944	6.042	5.856	5.875	6.075	5.821	5.839	5.867	5.602
Al	1.892	1.869	1.988	1.909	1.301	2.023	1.551	1.547	2.341
Fe^{3+}	0.164	0.089	0.157	0.215	0.623	0.156	0.610	0.586	0.057
Ti	0.219	0.196	0.229	0.232	0.398	0.450	0.495	0.523	0.552
Cr	0.009	0.038	0.100	0.073	0.000	0.154	0.008	0.007	0.004
Mn	0.000	0.000	0.009	0.000	0.003	0.007	0.012	0.020	0.006
Fe^{2+}	0.190	0.234	0.194	0.159	0.052	0.192	0.089	0.092	0.762
Mg	5.434	5.327	5.256	5.330	5.025	4.992	4.922	4.857	4.306
Ba	0.016	0.013	0.026	0.024	0.075	0.023	0.138	0.122	0.051
Na	0.000	0.000	0.036	0.060	0.172	0.025	0.074	0.074	0.011
K	1.873	1.914	1.920	1.888	1.854	1.909	1.761	1.772	1.912

Trace Elements (ppm)

	8	9	10	11	12
Cr	5255	5555	4965	3936	3354
Ba	4884	3784	3785	3012	3164
Ni	2255	2435	2083	2471	2227
Nb	18	15	24	17	26
Zr	164	122	131	171	147
Y	5	7	5	9	4
Sr	207	128	162	135	273
Rb	287	250	270	275	274
Pb	2	4	8	3	6
Co	54	54	57	54	54

1 From olivine orendite (SK 36), South Table Mtn.
2 From olivine orendite (SK 36), South Table Mtn.
3 From wyomingite (SK 9), Zirkel Mesa, (a) core of phlogopite grain, (b) rim of phlogopite grain
4 From madupite (SK 28), Pilot Butte
5 From transitional rock (SK 32), Twin Rocks, (a) core of phlogopite grain, (b) rim of phlogopite grain
6 From transitional rock (SK 32), Twin Rocks. Phlogopite in groundmass
7 From olivine orendite (SK 36), South Table Mtn., average analysis of dark phlogopite core
8 From wyomingite (SK 9), Zirkel Mesa (different sample from 3)
9 From wyomingite (SK 10), Emmons Mesa
10 From wyomingite (SK 41), Spring Butte
11 From orendite (SK 35), North Table Mtn.
12 From olivine orendite (SK 36), South Table Mtn. (different sample from 7)
† Fe^{3+} calculated to fill the tetrahedral sites to 8.000

the Leucite Hills rocks. Both opaque and deep-red colored magnesiochromites occur in the olivine orendites. Red magnesiochromites have been reported from nodules in kimberlites (MacGregor, 1979) and in basanite (Frey and Prinz, 1978) of upper mantle origin. These are type I xenocrysts using the nomenclature of Frey and Prinz (1978). Mitchell and Clark (1976) have interpreted red magnesiochromites found in the Peuyuk kimberlite as having crystallized under relatively reducing conditions in the mantle.

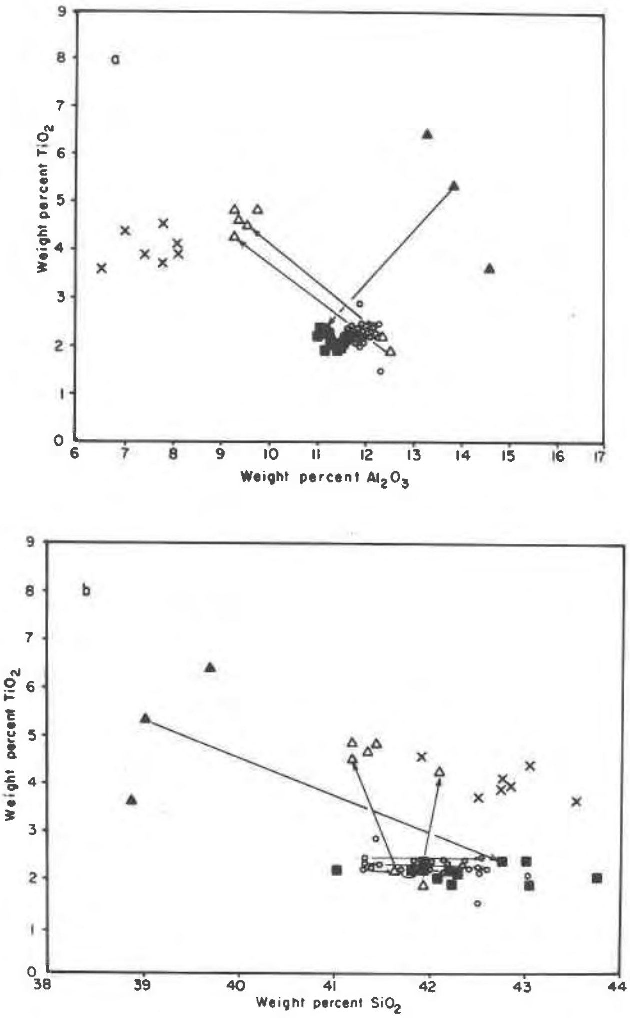

Fig. 1. Variations in (a) TiO_2 *vs.* Al_2O_3 (top) and (b) TiO_2 *vs.* SiO_2 for phlogopites from the Leucite Hills rocks (bottom). X—madupites; △—transitional rocks; ○—wyomingites; □—orendites; ■—olivine orendites; ▲—dark phlogopite cores in olivine orendites. Arrows indicate core → rim analyses.

As the red magnesiochromites in the olivine orendites have similar Fe_2O_3/FeO to those found in kimberlites (Mitchell and Clark, 1976; Mitchell, 1979), these chromites may also have formed under relatively reducing conditions. Reducing conditions seem to be inconsistent with the Fe_2O_3/FeO values for the Leucite Hills rocks (Table 8); thus the red magnesiochromites in the olivine orendites are considered to be xenocrysts.

The opaque magnesiochromite has high FeO, Fe_2O_3/FeO and low TiO_2 relative to other chromites (Deer *et al.*, 1962). This chemistry suggests it may have crystallized from a liquid with low Al_2O_3 and under highly oxidizing conditions similar to those inferred for the liquids from which the Leucite Hills lavas formed (Carmichael, 1967). The chemistry of this magnesiochromite suggests it was an early crystallizing mineral from an orendite liquid and is not a xenocryst.

The fine grain sizes of the skeletal titanomagnetite in the madupite and euhedral grains in the rims of the phlogopite microphenocrysts in the transitional rock were too small to analyze quantitatively but based on semiquantitative analyses appear to be Cr–Ti rich magnetites and pseudobrookites. The origin of these minerals is unknown.

The Fe-rich spinel (pleonaste) was found only in the wyomingites. Minerals of similar composition have been reported from granulite and clinopyroxenite nodules in basanites (Baldridge, 1979) and from mafic to ultramafic nodules in analcimitites (Wilkin-

Table 5. Representative leucite analyses

	Major Elements			
	1a	1b	2	3
SiO_2	56.88	57.96	59.54	57.71
Al_2O_3	20.32	18.38	18.27	19.62
Fe_2O_3*	1.81	2.90	2.67	2.03
BaO	0.03	0.04	0.00	0.09
Na_2O	0.08	0.04	0.18	0.05
K_2O	21.21	21.13	19.60	20.77
Σ	100.33	100.45	100.26	100.27
	Number of Cations on the Basis of 6 Oxygens			
Si	2.064	2.105	2.138	2.145
Al	0.869	0.786	0.773	0.837
Fe^{3+}	0.049	0.079	0.072	0.055
Na	0.006	0.003	0.013	0.004
K	0.982	0.979	0.848	0.959
Ba	0.000	0.001	0.000	0.001
	Trace Elements (ppm)			
	4	5	6	
Cr	17	31	nd	
Ba	1549	1994	2990	
Ni	7	4	nd	
Nb	23	25	25	
Zr	306	439	378	
Y	5	5	11	
Sr	323	309	418	
Rb	436	555	483	
Ba	13	19	11	
Co	10	12	10	

1 From wyomingite (SK 9), Zirkel Mesa, (a) core of leucite grain, (b) rim of leucite grain
2 From olivine orendite (SK 36), South Table Mtn.
3 From orendite (SK 34), North Table Mtn.
4 From wyomingite (SK 9), Zirkel Mesa (different sample from 1)
5 From wyomingite (SK 10), Emmons Mesa
6 From wyomingite (SK 41), Spring Butte
* Iron determined as FeO and recalculated to Fe_2O_3
nd not determined

Table 6. Partial analyses of apatites and perovskites

	1	2a	2b	3	4	5
MgO	0.23	0.52	0.40	0.32	-	-
K_2O	0.06	0.05	0.07	0.02	-	-
CaO	52.68	53.07	53.02	54.10	-	-
BaO	0.19	0.12	0.09	0.12	-	-
FeO *	0.13	0.04	0.11	0.05	-	-
P_2O_5	36.65	40.18	39.79	38.96	-	-
La_2O_3	0.27	0.17	0.12	0.25	2.03	2.05
Ce_2O_3	0.65	0.42	0.41	0.57	4.45	4.87
Pr_2O_3	0.09	0.07	0.00	0.07	1.47	1.64
Nd_2O_3	0.41	0.26	0.24	0.31	1.73	1.91
Σ	91.36	94.90	94.25	94.77	9.68	10.47

* Total iron determined as FeO.
1 Apatite from madupite (SK 23), Pilot Butte, average of 5 analyses
2 Apatite from wyomingite (SK 41), Spring Butte, (a) inclusion in phlogopite, (b) groundmass
3 Apatite from transitional rock (SK 32), Twin Rocks, average of 6 analyses
4 Perovskite from transitional rock (SK 32), Twin Rocks
5 Perovskite from madupite (SK 23), Pilot Butte

son, 1975). In both cases these nodules are believed to be xenoliths of type II (Frey and Prinz, 1978). Although similar nodules do not occur in the wyomingite, the chemistry of the pleonaste (Table 7), relative to that of the wyomingite, suggests it could be a xenocryst from a disaggregated (clinopyroxenite?) nodule. This hypothesis is supported by the common association in wyomingites of pleonaste with salite-cored diopsides, also believed to be xenocrysts.

Whole rock geochemistry

The high K_2O, low Na_2O and Al_2O_3 as well as the high values of Zr, Rb, Ba *etc.* (Table 8) indicate the highly potassic nature of Leucite Hills rocks. The

Table 7. Spinel analyses

	1a	1b	2
TiO_2	0.11	0.68	0.02
Al_2O_3	11.93	6.71	64.53
Cr_2O_3	57.52	44.07	0.10
FeO*	13.77	32.81	14.32
MgO	15.33	10.79	20.32
MnO	0.93	0.79	0.06
ZnO	nd	nd	0.32
Σ	99.59	95.85	99.67

* Total iron determined as FeO
1 Magnesiochromite from olivine orendite (SK 36), South Table Mtn., (a) red variety, (b) opaque variety
2 Pleonaste from wyomingite (SK 19), Deer Butte

orendites, olivine orendites, and wyomingites are chemically similar, except that the orendites tend to be richer in SiO_2 and poorer in CaO relative to the wyomingites. The olivine orendites are richer in MgO than either the orendites or wyomingites. The madupites are distinct from the other rocks in having lower SiO_2, Al_2O_3, and K_2O and higher Fe_2O_3, MgO, and CaO. The perpotassic–peralkaline nature of the olivine orendites, orendites, and wyomingites is indicated by their normative ks. In contrast only one madupite is perpotassic. All the rocks contain large normative pyroxene contents and the slight SiO_2-oversaturation in some rocks is indicated by normative Q. Based on their norms, the orendites and wyomingites are indistinguishable. The olivine orendites however can be distinguished from the other rock types by their higher modal and normative olivine. Modally, variations in the amounts of leucite and sanidine distinguish the orendites and olivine orendites from the wyomingites which have very minor sanidine and high leucite contents.

The analysis of the transitional rock (SK 32) is quite dissimilar from that of the other rocks of the Leucite Hills. The much higher H_2O and CO_2 contents of this rock are probably due to subsolidus alteration as indicated by vesicular carbonates, analcime, and other zeolites. Similarly, the madupite (SK 22) is also an altered rock as it has uncharacteristically high volatile and low alkalies. The other oxides,

Table 8. Whole rock analyses

	1	2	3	4	5	6	7	8
SiO_2	51.03	52.98	51.57	54.86	53.45	42.71	43.10	43.16
TiO_2	2.67	2.55	2.42	2.52	2.14	2.40	2.32	2.35
ZrO_2	0.17	0.17	0.17	0.18	0.17	0.17	0.17	0.16
Al_2O_3	9.81	10.49	10.10	10.80	10.27	7.67	8.58	8.37
Cr_2O_3	0.07	0.06	0.08	0.06	0.08	0.08	0.09	0.06
Fe_2O_3	3.67	2.64	2.85	3.00	3.60	4.72	5.34	4.85
FeO	0.65	1.94	1.63	0.96	1.00	1.46	0.80	0.84
MnO	0.07	0.07	0.08	0.05	0.08	0.14	0.13	0.14
MgO	7.24	7.23	7.78	6.56	9.61	11.75	11.60	8.21
CaO	5.37	4.31	5.03	3.57	4.21	13.13	10.71	12.84
SrO	0.32	0.22	0.26	0.20	0.20	0.58	0.38	0.57
BaO	0.78	0.60	0.73	0.52	0.34	0.91	0.48	1.38
Na_2O	1.03	1.29	1.31	1.24	1.26	0.50	0.93	2.23
K_2O	10.61	11.15	11.32	10.70	10.62	2.52	8.53	4.31
P_2O_5	1.71	1.37	1.61	1.32	1.28	1.60	2.13	2.55
SO_3	1.00	-	-	0.37	-	-	0.50	0.50
Cl	0.02	-	-	-	0.03	-	nd	0.08
F	0.69	-	-	-	0.06	-	0.71	0.59
CO_2	0.52	-	-	-	0.57	-	0.48	2.76
$H_2O\pm$	2.91	-	-	-	1.15	-	3.87	4.65
LOI	-	2.86	2.79	3.24	-	9.44	-	-
$O \equiv F$	0.29	-	-	-	0.29	-	0.30	0.25
$O \equiv Cl$	-	-	-	-	0.01	-	-	0.02
Σ	100.05	99.93	99.74	100.15	100.42	99.78	100.85	100.00
$\frac{100\ Mg}{Mg + Fe}$	74	75	77	76	80	78	79	74
$\frac{Na + K}{Al}$	1.21	1.19	1.25	1.11	1.16	0.39	1.10	0.78
				CIPW Norms				
Q	0.24	-	-	0.30	-	-	-	-
Z	0.18	0.37	0.18	0.18	0.18	0.18	0.37	0.18
or	53.38	57.27	52.82	58.94	56.16	15.01	1.11	25.58
ab	-	-	-	-	-	4.19	-	15.72
an	-	-	-	-	-	11.12	-	1.67
ne	-	-	-	-	-	-	-	-
lc	-	-	1.74	-	-	-	35.74	-
ns	-	0.61	0.37	-	-	-	-	-
ks	3.12	2.34	3.28	1.25	1.87	-	2.65	-
di	5.41	9.74	11.47	4.55	4.33	37.44	22.29	20.99
hy	15.56	4.22	-	14.26	8.23	2.31	-	4.32
ac	2.31	7.39	8.32	6.93	9.24	-	6.93	-
fo	-	6.44	9.80	-	9.52	7.14	12.95	4.48
il	1.52	4.10	3.50	1.98	2.13	3.19	1.98	2.13
hm	2.88	-	-	0.64	0.48	4.80	2.88	4.80
cr	-	0.22	0.22	0.22	0.22	0.22	0.22	-
pf	-	0.68	0.95	-	1.77	0.95	2.18	2.04
tn	4.51	-	-	3.72	-	-	-	-
th	1.70	-	-	0.71	-	-	-	0.85
fl	0.39	-	-	-	0.23	-	0.31	0.16
ap	4.03	3.36	3.70	3.02	3.02	3.70	5.04	6.05
cc	1.20	-	-	-	1.31	-	1.80	6.31
DI	60.51	67.61	66.53	68.13	67.27	17.20	46.44	42.15

Table 8. (continued)

				Trace Elements (ppm)				
Cr	506	417	565	402	565	577	607	427
Ba	7064	5408	6561	4621	3065	8180	4319	12355
Ni	270	274	267	309	428	163	162	149
Nb	42	58	53	48	45	120	137	90
Zr	1266	1250	1256	1298	1283	1233	1232	1152
Y	17	16	17	16	14	17	27	14
Sr	2677	1840	2179	1652	1674	4904	3196	4787
Rb	274	249	246	259	246	156	195	189
Pb	29	24	27	27	31	88	48	36
Co	22	58	22	61	26	33	34	57

1 Wyomingite (SK 9), Zirkel Mesa
2 Wyomingite (SK 10), Emmons Mesa
3 Wyomingite (SK 41), Spring Butte
4 Orendite (SK 35), North Table Mtn.
5 Olivine orendite (SK 36), South Table Mtn.
6 Madupite (SK 22), Pilot Butte
7 Madupite (SK 28), Pilot Butte
8 Transitional rock (SK 32), Twin Rocks

however, are comparable to those of other madupites analyzed in this study and reported in the literature.

In Figure 2 a D.I. plot of the major oxides given in Table 8, along with those from the literature, indicates two distinct groupings; the madupites, and the wyomingite–orendite–olivine orendite series. Excepting one analysis with a D.I. of 79 (Yagi and Matsumoto, 1966), the latter rocks plot within a narrow range of D.I. values (61–72). This indicates that crystal fractionation has probably not played a major role in their genesis, although the orendites and olivine orendites appear to be related by fractionation. Similarly, for the madupites, all analyses except SK 22 with D.I. = 17, which is altered, fall in the range of D.I. values between 33–46, indicating these rocks are probably only related to one another by minor fractionation. This variation could be produced by a maximum of 14% diopside fractionation, possibly as a result of flow differentiation. When SK 22 is recalculated on a volatile free basis and compared to the other madupites, using a molecular ratio diagram (Pearce, 1968), it has a D.I. of 39, well within the range of the other madupites. Although there is a compositional gap between the madupite and wyomingite–orendite–olivine orendite series (Fig. 2), the reasonably continuous trend on the D.I. plot implies a common genetic relationship between the rock types. Crystal fractionation based on the mineral analyses (Tables 2–7) cannot be the cause of this relationship as such fractionation requires removal or addition of excessive amounts of certain minerals which cannot be supported by the petrography and inferred paragenesis (Table 1).

Within wyomingites, orendites and olivine orendites the values of Nb, Zr, Y, Rb, and Pb are similar, Co and Cr are variable, and Ni is progressively depleted from the olivine orendites to the wyomingites to the madupites (Table 8). Madupites are depleted in Ni, Rb, and possibly Zr and enriched in Nb, Sr, and possibly Cr relative to the other rocks. The consistency in trace elements in the wyomingite–orendite–olivine orendite group supports the concept of these rocks being derived from a common source but unrelated by extensive crystal fractionation. The relationship of this group to the madupites is not clear on the basis of the trace element data.

Petrogenetic models

Small degrees of partial melting of a peridotitic mantle with 100 Mg/(Mg+Fe) of 92 produce a liquid with 100 Mg/(Mg+Fe) of 68–74 (Ringwood, 1975; Bence et al., 1980). With increasing pressure, such a liquid will be more Mg-rich. Partial melting of the same source with 2000 ppm Ni and 3000 ppm Cr yields a primary liquid with 300–400 ppm Ni and 400–500 ppm Cr (Ringwood, 1975; Green, 1980). Deviations from these values may be caused by larger degrees of partial melting, by partial melting of an anomalous mantle source, by crystal fractionation, or possibly by changes in pressure and fO_2 which may alter K_D values of material being fractionated (cf. Mysen, 1975).

For the Leucite Hills rocks, the molecular 100 Mg/(Mg+Fe) values (where iron mol. ratios have been calculated from FeO + 0.8998 (Fe_2O_3) and shown as Fe_T) are generally greater than 72. The Ni values for the orendites, olivine orendites, and wyomingites are within the range predicted for small degrees of partial melting of peridotite while the Ni values for the madupites are lower. The Cr contents of all rocks are

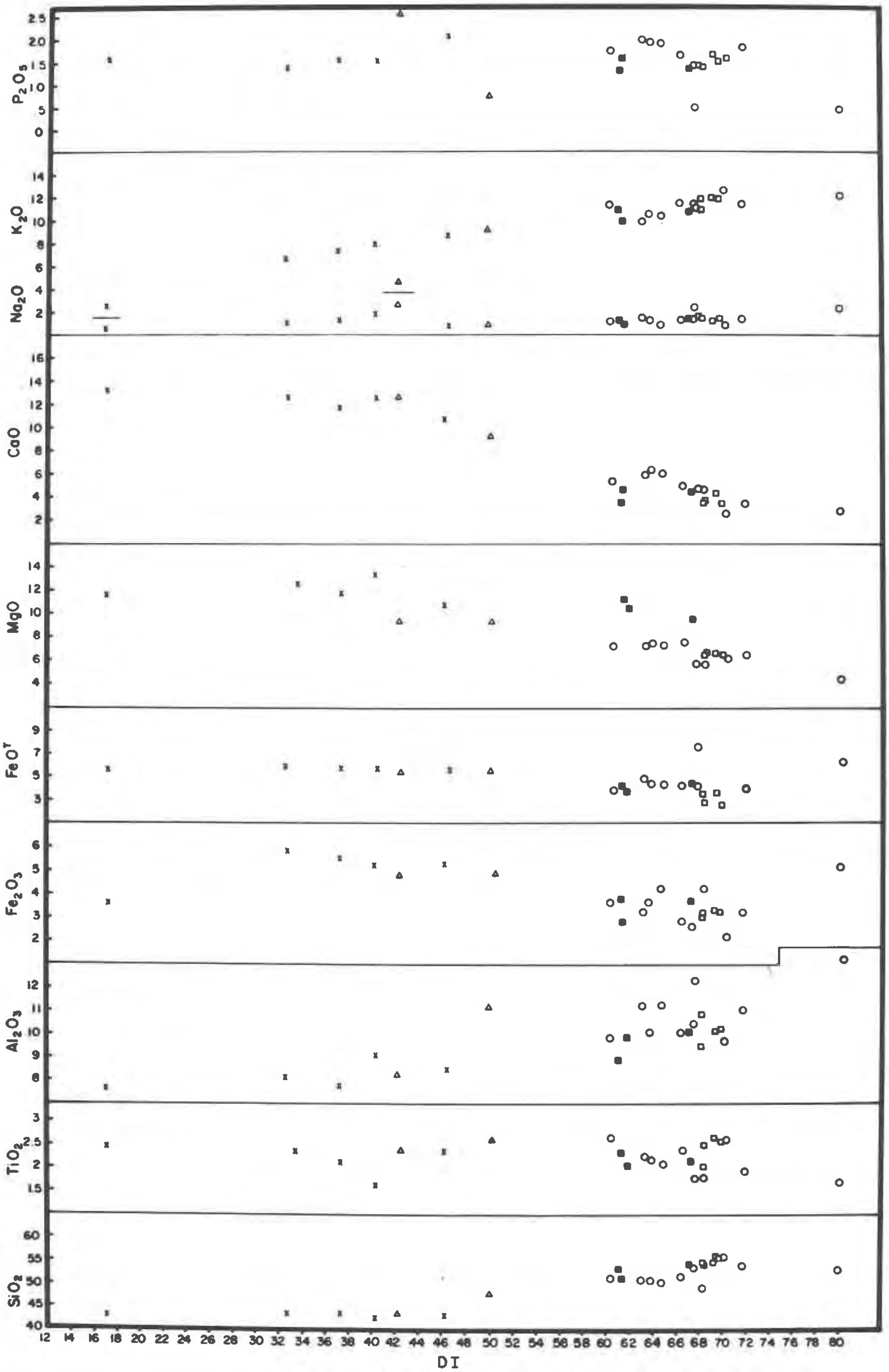

Fig. 2 Differentiation Index (D.I.) diagram for Leucite Hills rocks. Data from present study, and from Cross (1897), Schultze and Cross (1912), Johnston (1959), Smithson (1959), Yagi and Matsumoto (1966), Carmichael (1967), Barton and Hamilton (1978). Symbols as in Figure 1.

within, or slightly greater than, the predicted values for such partial melting.

The lower Ni values in the madupite cannot be explained by fractional crystallization of diopside and/or phlogopite as such fractionation would also lower the Cr contents of the madupite. Based on the $D_{Ni}^{Ol/L}$ (cf. Ringwood, 1975) and the equilibrium fractional crystallization equation (Greenland, 1970), the amount of olivine fractionation to derive the madupite can be calculated. Assuming an initial approximation of 350 ppm Ni in the parental madupite liquid, using the $D_{Ni}^{Ol/L}$ values of Ringwood (1975), fractionation of 8% olivine (100 Mg/(Mg+Fe) = 90) from such a liquid would be necessary to lower the Ni content to the 160 ppm observed in the madupite (Table 8). However, if the madupite (100 Mg/(Mg+Fe_T) = 78) is the product of 8% olivine fractionation, the parental madupite liquid must have had a 100 Mg/(Mg+Fe_T) value of 85. For a 'normal' mantle source, very high degrees of partial melting would be required to produce a derivative liquid with such a high 100 Mg/(Mg+Fe_T) ratio or the K_D^{Fe-mg} value for olivine must be greater than 0.33 as given by Roeder and Emslie (1970) for atmospheric pressure. The absence of olivine in the madupites also seems to preclude olivine fractionation as a viable model.

A mantle source for all of the rocks seems likely based on their consistently high 100 Mg/(Mg+Fe_T) ratios and their Ni and Cr contents. The possibility of large degrees of partial melting of a normal peridotite mantle are unlikely as such a mechanism fails to explain the high concentrations of LIL elements in the Leucite Hills rocks. High concentrations of LREE led Kay and Gast (1973) to propose that the madupites, wyomingites and orendites were derived by very small degrees of partial melting of an extremely enriched REE mantle source.

The genesis of the rocks of the Leucite Hills can be modeled on the basis of phase relations in the systems $KAlSiO_4$–Mg_2SiO_4–SiO_2–CO_2 and $KAlSiO_4$–MgO–SiO_2–CO_2–H_2O which have been determined under upper mantle conditions (Wendlandt and Eggler, 1980a, b). Although diopside, an important constituent of the Leucite Hills rocks, is not included in this system, Wendlandt and Eggler (1980a, p. 414) state that the absence of CaO in this system "... is not expected to have a major effect on phase relations defining the minimum pressure of generation of alkali basalts". Figures 3a, b show phase relations in the 28–34 and 18.5–19.5 kbar sections respectively of the anhydrous $KAlSiO_4$(Ks)–Mg_2SiO_4(Fo)–

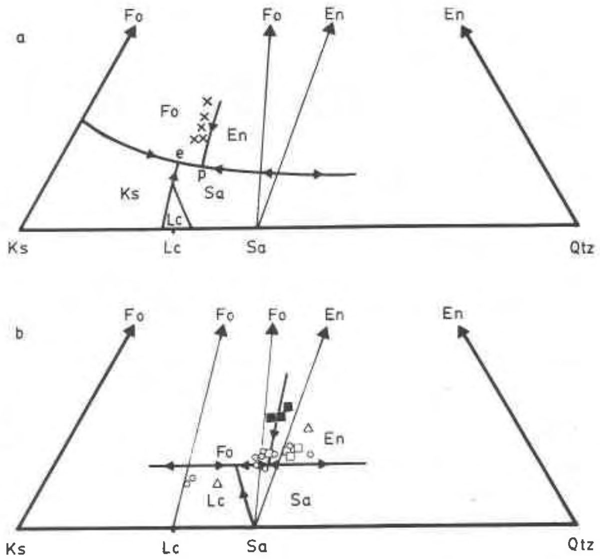

Fig. 3 Schematic liquidus projections for the system $KAlSiO_4$ (Ks)–Mg_2SiO_4 (Fo)–SiO_2 (Qtz) (after Wendlandt and Eggler, 1980a) with recalculated normative compositions of Leucite Hills rocks. (a) projection at 28–34 kbar under volatile-absent conditions; (b) projection at 18.5–19.5 kbar under volatile-absent conditions. Data sources and symbols as in Figures 1 and 2.

SiO_2(Qtz) system, together with recalculated normative compositions of the Leucite Hills rocks. As shown by Wendlandt and Eggler (1980a, Fig. 6), the primary phase fields of enstatite and sanidine expand with increasing pressure producing increasingly silica-undersaturated liquids on partial melting at increasing depths. Figure 3a, representing the 28–34 kbar section under volatile-absent conditions, shows that the madupite compositions plot near the invariant point (P) involving enstatite, forsterite, and sanidine. This suggests that partial melting of a source material, whose composition can be represented within the Fo–En–Sa subsystem, may produce a liquid comparable to that of the madupite.

If the same source material from which the liquids of composition comparable to the madupite were produced, rises diapirically to a depth corresponding to 18.5–19.5 kbar (55–59 km) under volatile-absent conditions, melting will recur. This situation is modeled by the phase relations in Figure 3b and in Figure 4. In Figure 3b members of the recalculated wyomingite–orendite series plot near the enstatite–forsterite–sanidine eutectic for pressures between 18.5–19.5 kbar, suggesting that liquids corresponding to the wyomingite–orendite compositions are generated at this point.

The path of the rising solid material and the gener-

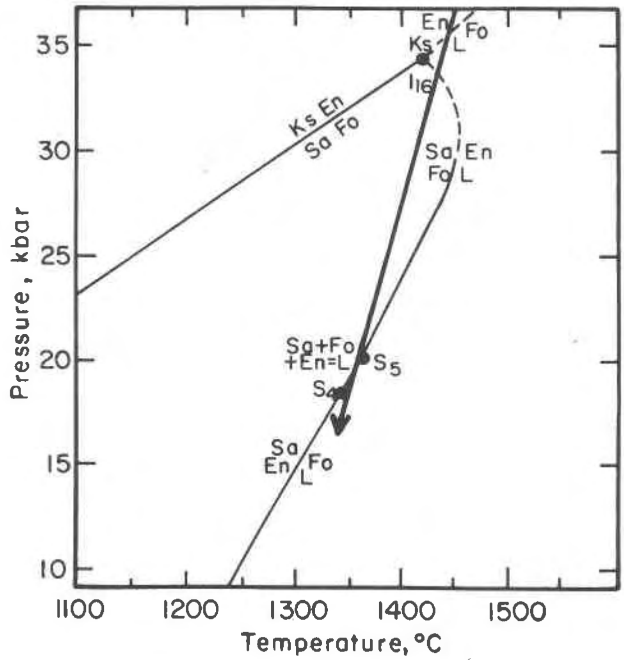

Fig. 4 *P–T* diagram (modified from Wendlandt and Eggler, 1980a, Fig. 1) showing path (thick line) of diapiric uprise of material producing Leucite Hills lavas. Madupite liquids are produced around 34 kbar (I_{16}) and wyomingite–orendite liquids between 18.5–19.5 kbar (S_4–S_5).

ation of the madupite and wyomingite–orendite liquids is shown schematically in Figure 4 (modified from Wendlandt and Eggler, 1980a, Fig. 1) which represents the *P–T* section of the phase relationships shown in Figure 3. In Figure 4 the first liquid, corresponding to madupite composition, appears at some pressure and temperature above I_{16} (~34 kbar, 1350°C). With decreasing pressure and a very slight decrease in temperature, liquid of madupite composition will continue to be produced until the diapir reaches a depth corresponding to a pressure just less than I_{16} (corresponding to Figure 3a). The path of the diapir then enters the Sa + Fo field by crossing the Sa + Fo $\rightleftarrows$ En + L reaction curve which has a negative slope. From just below $I_{16'}$ the solid will rise, without generating a melt, until it again passes from the Sa + Fo field into the En + L field defined by the Sa + Fo $\rightleftarrows$ En + L reaction curve between S_4 and S_5 at 18.5–19.5 kbar (corresponding to Figure 3b). At this stage a liquid corresponding in composition to wyomingite and orendite will be produced.

The relationships described above are modeled on the volatile-absent system. If the Mg_2SiO_4–$KAlSiO_4$–SiO_2–CO_2 system is used, under CO_2-saturated conditions, the corresponding pressures for production of madupite will be 24–27 kbar (84–94 km) and the production of the wyomingite–orendite liquids will occur at 14.5–15.5 kbar (51–54 km). These relationships can be illustrated by referring to Wendlandt and Eggler (1980a, Figs. 6 and 3). In Figure 3 of Wendlandt and Eggler (1980a) the production of madupite liquids will occur at pressures just above I_{14} to just below I_{18} and that of wyomingite–orendite liquids between S_{10} and S_{11}.

If the madupite and wyomingite–orendite liquids have been generated under conditions of CO_2-saturation and H_2O-present, but buffered at low a_{H_2O}, the phase relations in the $KAlSiO_4$–MgO–SiO_2–H_2O–CO_2 system (Wendlandt and Eggler, 1980b) can be used as a model. Under these conditions, the madupite liquid would form around 24 kbar (see Wendlandt and Eggler, 1980b, Fig. 10, point I_{15}), corresponding to 84 km depth, whereas the wyomingite–orendite liquids would form between 14 and 17 kbar (see Wendlandt and Eggler, 1980b, Fig. 10, points S_{14}–S_{17}) corresponding to 49–60 km depth.

The mechanism postulated to explain the generation of the two chemically distinct primitive liquids of the Leucite Hills suite is caused by the negative dP/dT slopes of the high pressure solidus curves and the positive slopes of the same curves at lower pressures in the systems used to model this suite. This model also implies that the Leucite Hills rocks have been generated by partial melting of a single K-enriched peridotitic mantle source. $^{87}Sr/^{86}Sr$ data for madupite and orendite (Ogden and Vollmer, 1980) support a single source model.

The Ni, Co, Cr, and Zr contents of the Leucite Hills rocks do not correlate with the concept of very small but increasing degrees of partial melting model proposed by Kay and Gast (1973) based on REE data. Whether these trace elements fit a two-stage melting process from a single parental K-enriched peridotite source, as we propose, is difficult to assess. Trace element behavior in such a process is not well known, nor is the behavior of such elements based on bulk distribution coefficients of an anomalous, rather than a normal, mantle source. The mineralogy of such an anomalous mantle is speculative but, based on experimental studies, may contain phlogopite and apatite (Barton and Hamilton, 1978) or even wadeite (Arima and Edgar, 1980) for which bulk distribution coefficients are unknown. The distribution of Ba and Rb in these rocks suggests that phlogopite fusion may have been involved in their genesis.

Chemically the wyomingites and orendites are similar (Table 8) and are only distinguishable on the basis of their mineralogy (Table 1). Differences in their mineralogy are likely due to crystallization in the orendites having taken place at greater depths than in the wyomingites. Experiments on both the wyomingite and orendite (Barton and Hamilton, 1978) show that leucite is not stable above 1 kbar P_{H_2O} ($P_{H_2O} = P_{tot}$) and that sanidine is stable at higher pressures than leucite. In the orendites, sanidine tends to be more abundant than leucite and appears to have formed prior to leucite (Table 1), suggesting crystallization of both minerals may have occurred around 1 kbar $P_{H_2O} = P_{total}$. In contrast, the wyomingites contain very little sanidine but have appreciable glass, with leucite increasing as glass increases, indicating more rapid crystallization at shallower depths.

The olivine orendites are the only rocks in the Leucite Hills suite which have been produced by major fractionation processes. As shown in Figure 3 the olivine orendites are more Fo-rich than the orendites. This suggests that fractionation of olivine from the orendites may have taken place. Using the method of Reid *et al.* (1973) the fractionation of a total of less than 20% (consisting of 4% olivine, 4% diopside and 11% phlogopite) from the orendite would produce the olivine orendite.

Paucity of data and the altered nature of the transitional rocks do not permit us to determine their genesis. Preliminary results suggest a magma mixing model may explain these rocks. This problem is under investigation.

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References

Arima, M. and Edgar, A. D. (1980) Stability of wadeite ($Zr_2K_4Si_6O_{18}$) under upper mantle conditions: petrological implications. Contributions to Mineralogy and Petrology, 72, 791–795.

Baldridge, W. S. (1979) Mafic and ultramafic inclusion suites from the Rio Grande Rift (New Mexico) and their bearing on the composition and thermal state of the lithosphere. Journal of Volcanology and Geothermal Research, 6, 319–351.

Barton, M. and Hamilton, D. L. (1978) Water-saturated melting relations to 5 kilobars of three Leucite Hills lavas. Contributions to Mineralogy and Petrology, 66, 41–49.

Barton, M. and Hamilton, D. L. (1979) The melting relationships of a madupite from the Leucite Hills, Wyoming, to 30 kb. Contributions to Mineralogy and Petrology, 69, 133–142.

Bell, K. and Powell, J. L. (1969) Strontium isotopic studies of alkalic rocks. The potassium rich lavas of the Birunga and Toro-Ankole regions, East and Central Equatorial Africa. Journal of Petrology, 10, 536–572.

Bence, A. E., Grove, T. L., and Papike, J. J. (1980) Basalts and probes of planetary interiors: constraints on the chemistry and mineralogy of their source regions. Precambrian Research, 10, 249–279.

Brooks, C. K. and Printzlau, I. (1978) Magma mixing in mafic alkaline volcanic rocks: the evidence from relict phenocryst phases and other inclusions. Journal of Volcanology and Geothermal Research, 4, 315-331.

Carmichael, I. S. E. (1967) The mineralogy and petrology of the volcanic rocks from the Leucite Hills, Wyoming. Contributions to Mineralogy and Petrology, 15, 24–66.

Cross, W. (1897) Igneous rocks of the Leucite Hills and Pilot Butte, Wyoming. American Journal of Science, 4, 115–141.

Deer, W. A., Howie, R. H., and Zussman, J. (1962) Rock-Forming Minerals, Vol. 5, Non-Silicates. Longmans, London.

Edgar, A. D., Green, D. H., and Hibberson, W. O. (1976) Experimental petrology of a highly potassic magma. Journal of Petrology, 17, 339–356.

Frey, F. A. and Prinz, M. (1978) Ultramafic inclusions from San Carlos, Arizona: petrological and geochemical data bearing on their petrogenesis. Earth Planetary Science Letters, 30, 127–176.

Gast, P. W. (1968) Trace element fractionation and the origin of tholeiite and alkaline magma types. Geochimica et Cosmochimica Acta, 32, 1057–1086.

Green, T. H. (1980) Island-arc and continental building magmatism—a review of petrogenetic models based on experimental petrology and geochemistry. Tectonophysics, 63, 367–385.

Greenland, L. P. (1970) An equation for trace element distribution during magmatic differentiation. American Mineralogist, 55, 455–466.

Hawkesworth, C. J. and Vollmer, R. (1979) Crustal contamination versus enriched mantle: $^{143}Nd/^{144}Nd$ and $^{87}Sr/^{86}Sr$ evidence from the Italian volcanics. Contributions to Mineralogy and Petrology, 69, 151–165.

Henderson, C. M. B. (1965) Minor element chemistry of leucite and pseudoleucite. Mineralogical Magazine, 35, 596–603.

Holmes, A. (1942) A suite of volcanic rocks from southwest Uganda containing kalsilite (a polymorph of $KAlSiO_4$). Mineralogical Magazine, 26, 197–217.

Johnston, R. H. (1959) Geology of the Northern Leucite Hills, Sweetwater County, Wyoming. M.A. thesis, University of Wyoming, Laramie.

Kay, R. W. and Gast, P. W. (1973) Rare earth content and origin of alkali-rich basalts. Journal of Geology, 81, 653–682.

Kemp, J. F. (1897) The Leucite Hills of Wyoming. Geological Society of America, Bulletin, 8, 169–182.

Kemp, J. F. and Knight, W. C. (1903) Leucite Hills of Wyoming. Geological Society of America Bulletin, 4, 305–336.

Kuehner, S. M. (1980) Petrogenesis of ultrapotassic rocks, Leucite Hills, Wyoming. M.Sc. thesis, University of Western Ontario, London, Ontario.

MacDowell, F. W. (1966) Potassium Argon Dating of Cordilleran Intrusives. Ph.D. thesis, Columbia University, New York.

MacGregor, I. D. (1979) Mafic and ultramafic xenoliths from the Kao kimberlite pipe. In F. R. Boyd and H. O. A. Meyer, Eds. The Mantle Sample: Inclusions in Kimberlites and Other Volcanics. Proceedings of the Second International Kimberlite Conference, vol. 2, p. 156–172, American Geophysical Union, Washington.

Matthews, V. and Work, D. F. (1978) Laramide folding associated with basement block faulting along the northeastern flank of the Front Range, Colorado. In V. Matthews III, Ed. Laramide Folding Associated with Block Faulting in the Western United States. Geological Society of America Memoir, 151.

Mitchell, R. H. (1979) Mineralogy of the Tunraq kimberlite, Somerset Island, N.W.T., Canada. In F. R. Boyd and H. O. A. Meyer, Eds., Kimberlites, Diatremes and Diamonds: Their Geology, Petrology and Geochemistry. Proceedings of the Second International Kimberlite Conference, Vol. 1, p. 161–171, American Geophysical Union, Washington.

Mitchell, R. H. and Clark, D. B. (1976) Oxide and sulphide mineralogy of the Peuyuk kimberlite. Contributions to Mineralogy and Petrology, 56, 157–172.

Mysen, B. O. (1975) Partitioning of iron and magnesium between crystals and partial melts in peridotite upper mantle. Contributions to Mineralogy and Petrology, 52, 69–76.

Ogden, P. R. and Vollmer, R. (1980) Leucite Hills revisited. $^{87}Sr/^{86}Sr$ evidence. EOS, 61, 412.

Pearce, T. H. (1968) A contribution to the theory of variation diagrams. Contributions to Mineralogy and Petrology, 19, 142–157.

Prodehl, C. (1976) Comparison of seismic refraction studies in central Europe and the western United States. In P. Giese, C. Prodehl and A. Stein, Eds., Explosion Seismology in Central Europe, p. 385–395. Springer–Verlag.

Reid, M. J., Gancarz, A. J., and Albee, A. L. (1973) Constrained least squares analysis of petrologic problems with an application to lunar sample 12040. Earth Planetary Science Letters, 17, 433–445.

Ringwood, A. E. (1975) Composition and Petrology of the Earth's Mantle. McGraw-Hill, New York.

Roden, M. F. and Smith, D. (1979) Field geology, chemistry and petrology of the Buell Park minette diatreme, Apache County, Arizona. In F. R. Boyd and H. O. A. Meyer, Eds. Kimberlites, Diatremes and Diamonds: Their Geology, Petrology and Geochemistry. Proceedings of Second International Kimberlite Conference, Vol. 1, p. 364–381, American Geophysical Union, Washington.

Roeder, P. L. and Emslie, R. F. (1970) Olivine-liquid equilibration. Contributions to Mineralogy and Petrology, 29, 275–289.

Ryabchikov, I. D. and Green, D. H. (1978) The role of carbon dioxide in the petrogenesis of highly potassic magmas. In Problems of Petrology of the Earth's Crust and Upper Mantle, Trudy Instituta Geologii Geofiziki, 403, Nauka, Novosibirsk.

Sahama, Th.G. (1974) Potassium-rich alkaline rocks. In H. Sørensen, Ed. The Alkaline Rocks, p. 94–109, Wiley and Sons, London.

Schultze, A. R. and Cross, W. (1912) Potash-bearing rocks of the Leucite Hills, Sweetwater County, Wyoming. U.S. Geological Survey Bulletin, 512.

Smithson, S. B. (1959) The Geology of the Southeastern Leucite Hills, Sweetwater County, Wyoming. M.A. thesis, University of Wyoming, Laramie.

Smithson, S. B., Brewer, J. A., Kaufman, S., Oliver, J. E. and Hurich, C. A. (1979) Structure of the Laramide Wind River Uplift, Wyoming from COCORR deep reflection data and gravity data. Journal Geophysical Research, 84, 5955–5972.

Stearns, D. W. (1978) Faulting and forced folding in the Rocky Mountains foreland. In V. Matthews, Ed. Laramide Folding Associated with Basement Block Faulting in the Western United States. Geological Society of America Memoir 151.

Stewart, J. H. (1978) Basin-range structure in western North America: a review. In R. B. Smith and G. P. Eaton, Eds. Cenozoic Tectonics and Regional Geophysics of the Western Cordillera. Geological Society of America Memoir 132.

Stormer, J. C. (1973) Calcium zoning in olivine and its relationship to silica activity and pressure. Geochimica et Cosmochimica Acta, 37, 1815–1821.

Thompson, R. N. (1974) Some high pressure pyroxenes. Mineralogical Magazine, 39, 768–787.

Turi, B. and Taylor, H. P. (1976) Oxygen isotope studies of potassic volcanic rocks of the Roman Province, central Italy. Contributions to Mineralogy and Petrology, 55, 1–31.

Watson, E. B. (1979) Calcium content of forsterite coexisting with silicate liquid in the system Na_2O–CaO–MgO–Al_2O_3–SiO_2. American Mineralogist, 64, 824–829.

Wendlandt, R. F. and Eggler, D. H. (1980a) The origins of potassic magmas: 1. Melting relations in the systems $KAlSiO_4$–Mg_2SiO_4–SiO_2 and $KAlSiO_4$–MgO–SiO_2–CO_2 to 30 kilobars. American Journal of Science, 280, 385–420.

Wendlandt, R. F. and Eggler, D. H. (1980b) The origins of potassic magmas: 2. Stability of phlogopite in natural spinel lherzolite and in the system $KAlSiO_4$–MgO–SiO_2–H_2O–CO_2 at high pressures and temperatures. American Journal of Science, 280, 421–458.

Wilkinson, J. F. G. (1975) An Al-spinel ultramafic-mafic inclusions suite and high pressure megacrysts in an analcimitite and their bearing on basalt magma fractionation at elevated pressures. Contributions to Mineralogy and Petrology, 53, 71–104.

Yagi, K. and Matsumoto, H. (1966) Note on the leucite-bearing rocks from the Leucite Hills, Wyoming, U.S.A. Journal of Faculty of Science, Hokkaido University, Series IV, 13, 307–312.

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