

Magma-mixing in the low velocity zone: kimberlitic megacrysts from Fayette County, Pennsylvania

ROBERT H. HUNTER¹ AND LAWRENCE A. TAYLOR

Department of Geological Sciences
University of Tennessee, Knoxville, Tennessee 37996

Abstract

Two chemically distinct populations of megacrysts/inclusions are present in hypabyssal-facies kimberlite from Fayette County, S.W. Pennsylvania: A Cr-rich suite and a Cr-poor suite. The Cr-rich suite consists of olivine (Fo 90–93), garnet (1.64–6.14% Cr₂O₃), diopside (1.89–2.34% Cr₂O₃), Cr-spinel, and immiscible sulfide-melt products. The suite possesses a limited range of *mg*, and major element compositions of minerals are similar to those in sheared garnet-lherzolite xenoliths. Minor element variations are consistent with crystal fractionation over a temperature interval from 1310° to 1055°C in the pressure range 48–39 kbar. The Cr-poor suite is more evolved and comprises olivine (Fo 81–85), garnet (<0.1% Cr₂O₃), picroilmenite (16–37 mol% MgTiO₃; 0.05–2.05% Cr₂O₃), and diopside (<0.1% Cr₂O₃). This suite represents a relatively advanced stage of fractionation of a Cr-poor melt within the low velocity zone (LVZ). Mixing of these populations and their host melts has resulted in rim compositions of Fo 88–89 on both groups of olivines and reverse-zoned rims on the ilmenite megacrysts (34–51 mol% MgTiO₃; 1.4–3.6% Cr₂O₃).

The Cr-poor suite crystallized from ponded melts segregated from rising diapirs in the LVZ. Crystal fractionation in these melts led to evolved mineral compositions and ilmenite saturation. The Cr-rich suite represents crystal and immiscible sulfide-melt fractionation in a melt within an uprising diapir; the melt/crystal ratio was low and major element compositions in the melt were buffered by ambient mantle compositions (undepleted garnet lherzolite). Uprise of this melt-bearing diapir disrupted evolved, Cr-poor crystal mushes incorporating megacrysts and resulting in mixing of the host melts. Back-reaction and crystallization of the cognate xenocrysts within the hybrid melt resulted in the zoning patterns observed.

Introduction

A characteristic feature of kimberlites worldwide is their, often abundant, megacrysts or discrete nodules (*e.g.*, Boyd and Dawson, 1972; Boyd and Clement, 1977; Boyd and Nixon, 1973; 1978; Egger *et al.*, 1979; Frantsesson, 1970; Garrison and Taylor, 1980; Gurney *et al.*, 1979; Mitchell, 1979; Nixon and Boyd, 1973a; 1979). Despite their widespread occurrence, there has been considerable debate concerning their origin, particularly the picroilmenite megacrysts and ilmenite-silicate intergrowths (*e.g.*, Boyd and Nixon, 1973; Dawson and Reid, 1970; Garrison and Taylor, 1981; Gurney *et al.*, 1973; Haggerty *et al.*, 1979; Mitchell, 1973; 1977; Pasteris *et al.*, 1979; Pasteris, 1980; Rawlinson and Dawson, 1979). The following features are characteristic of typical megacryst suites: (1) grain-sizes are in the range 1–10 cm, considerably larger than average grain-sizes in mantle-derived

peridotite xenoliths (0.2–0.4 cm); (2) they often possess compositional ranges distinct from associated mantle xenoliths; (3) calculated equilibration pressures place their origin within the depth range 100–200 km, and commonly, megacryst suites from a single intrusion display a wide range of equilibration temperatures, typically 200–300°C (*e.g.*, Pasteris *et al.*, 1979); and (4) temperature-related mineralogical and textural features frequently are present (*e.g.*, Boyd and Nixon, 1975). These textural and mineralogical characteristics suggest that an origin by disaggregation of mantle peridotite is unlikely and it is a widely held view among kimberlite petrologists, that these features are consistent with high-pressure crystallization from fractionating melts within the low velocity zone (LVZ). A principal point of contention is the relationship of these megacrysts to their host kimberlite. Are they xenocrysts (*e.g.*, Boyd and Nixon, 1973; Pasteris *et al.*, 1979; Pasteris, 1980) or phenocrysts in a kimberlitic liquid (*e.g.*, Dawson and Stevens, 1975; Egger *et al.*, 1979; Garrison and Taylor, 1980; Gurney *et al.*, 1979; Mitchell, 1973; Mitchell and Clarke, 1976)? In

¹ Present address: Grant Institute of Geology, University of Edinburgh, West Mains Road, Edinburgh, U.K.

addition, the origin and significance of rim-zonation in olivine (e.g., Boyd and Clement, 1977; Emeleus and Andrews, 1975) and ilmenite megacrysts (e.g., Agee *et al.*, 1982; Boctor and Boyd, 1980a; Haggerty *et al.*, 1979; Pasteris, 1980) have also provoked debate.

Fresh, hypabyssal-facies kimberlite from S. W. Pennsylvania contains a well-preserved megacryst and inclusion suite typical of kimberlites in general. Two chemically distinct populations of megacrysts and inclusions are present, a Cr-rich suite and a Cr-poor suite; these are analogous to those described from the Colorado-Wyoming kimberlites by Egger *et al.* (1979). The Cr-rich suite consists of olivine, garnet, orthopyroxene, clinopyroxene, Cr-spinel, and sulfides; this group shows minor element compositions that vary systematically with equilibration temperatures and pressures, although *mg* shows a restricted range. The Cr-poor suite consists of olivine, garnet, ilmenite, and clinopyroxene; these phases possess restricted compositional ranges but are significantly more evolved than the Cr-rich suite. Olivines in both groups possess identical rim compositions, and the ilmenite megacrysts of the Cr-poor suite display reverse zonation. These features are interpreted within the context of a magma-mixing model. This model is incorporated with aspects of models presented by other authors into a reasonably coherent scenario for evolution of kimberlitic melts within the LVZ.

Petrography and mineral chemistry

Megacrysts and xenoliths are set in a matrix largely composed of phlogopite, ilmenite, Cr-spinel-titanomagnetite_{ss}, perovskite, and carbonate with subordinate serpentine, apatite, rutile, and sulfides. The petrology of these groundmass phases, together with the location and general characteristics of the Fayette County kimberlite are described in a companion paper (Hunter *et al.*, 1983).

Groundmass grains are less than 2 mm in size. Megacrysts are rounded to subangular, and typically, dimensions are in the range 0.5–1.5 cm, but grains are present up to 3 cm. The term megacryst is used herein to describe optically continuous, single-phase grains or partial mosaics. Inclusions are single grains (except in the case of the sulfide blebs) within megacrysts, typically 0.5–1.0 mm in size, and are rounded to subhedral in form. One cm is usually considered to be the lower size limit of megacrysts *sensu stricto* (e.g., Harte, 1978). Commonly, in kimberlites where there is a distinct bimodal distribution of grain sizes, the lower limit of the large grain-size population is around one cm. However, in the discrete (i.e., non-xenolithic) olivine population, as a whole, there is a *seriate* size distribution (<2 mm to 3 cm), skewed towards the smaller size range. The cores of all the olivine grains less than 2 mm in size show chemical identity with the Cr-rich, olivine-megacryst population. Commonly, the smallest olivine grains and olivine inclusions fall within the size range exhibited by the other megacryst

inclusions, and clearly, there is an unambiguous cogenetic relationship between megacrysts and inclusions. Considerations of grain size distribution and chemistry of the olivines suggests that it would be misleading to consider the smaller olivine grains in isolation from the megacrysts, and usage of the term olivine herein includes the *entire* discrete olivine population.

Olivines with subordinate garnet comprise the phases in the Cr-rich suite; in order of decreasing abundance, inclusions consist of (Table 1): orthopyroxene, sulfide, olivine, garnet, clinopyroxene, and Cr-spinel. Within the Cr-poor suite, three phases exist as megacrysts. Relative modal abundances of these are olivine > ilmenite > garnet; ilmenite inclusions are more abundant than clinopyroxene. As a group, the Cr-rich suite is relatively more abundant than the Cr-poor suite, by a factor of 2–3. Phlogopite megacrysts are rare and may or may not belong to one of the two groups. Ultramafic xenoliths are not common; their characteristics are described briefly, but they are only discussed within the context of the megacryst evolution.

Mineral analyses were performed using an automated MAC 400S electron microprobe, using data reduction procedures of Bence and Albee (1968) and the data of Albee and Ray (1970). Whenever possible, 5–10 analyses were performed on each grain, and means and standard deviations computed. Core regions of grains were homogeneous, but zonation was present in the outer 100–200 μm of olivine and ilmenite megacrysts. Representative analyses of silicates and oxides are presented in Tables 2 through 6; ranges of chemical parameters of minerals in the Cr-rich and Cr-poor suites and in garnet-herzolite xenoliths are summarized in Table 7.

Olivine

As noted above, olivine grains possess a seriate size distribution with the largest megacrysts 3 cm in size; usually, they are rounded to oblate. Many olivines display a variety of strain features. The majority of the grains are fractured; often incipient serpentization is present along more prominent fractures. Higher degrees of stress have resulted in recrystallization into a mosaic of

Table 1. Megacryst inclusions

Group	Megacryst	Inclusion(s)	
Cr-rich	Olivine:	garnet	(2)*
		garnet + cpx	(1)
		cpx	(2)
		opx	(7)
		opx + sulphide	(1)
		sulphide	(5)
		Cr-spinel	(2)
		olivine	(4)
Cr-poor	Garnet:	ilmenite	(2)
		ilmenite + cpx	(1)

Note: The data-base was thirty 4.5 x 2.5 cm thin-sections. * The numbers in parentheses indicate the number of megacryst samples bearing that particular assemblage.

Table 2. Representative olivine analyses

	Cr-rich megacryst		Cr-poor megacryst		Lherzolite
	core	rim	core	rim	
SiO ₂	40.8 (4)*	40.2 (2)	39.3 (2)	40.1 (4)	40.7 (2)
Cr ₂ O ₃	0.07(2)	n.d.	n.d.	n.d.	0.10(2)
MgO	51.2 (5)	48.1 (5)	45.8 (1)	48.5 (9)	50.7 (4)
FeO**	8.05(15)	10.9 (1)	15.2 (1)	11.2 (5)	8.08(5)
MnO	0.06(1)	0.11(4)	0.18(2)	0.16(2)	0.12(2)
NiO	0.37(2)	0.16(1)	0.20(2)	0.23(2)	0.42(3)
CaO	0.04(1)	0.14(1)	0.04(1)	0.11(2)	0.11(1)
Total	100.59	99.61	100.72	100.30	100.23
Cations on the basis of 4 oxygens					
Si	0.988	0.995	0.981	0.989	0.991
Cr	0.001	-	-	-	0.002
Mg	1.849	1.775	1.708	1.781	1.838
Fe	0.162	0.226	0.317	0.229	0.164
Mn	0.001	0.003	0.003	0.002	0.003
Ni	0.007	0.003	0.003	0.004	0.009
Ca	0.001	0.003	0.000	0.002	0.003
Σ	3.009	3.005	3.014	3.007	3.010
Fo	92.0	88.7	84.3	88.6	91.8

* The numbers in parentheses represent 1σ standard deviations in terms of the last digit cited for the value to their immediate left, thus 40.8 (4) indicates a standard deviation of 0.4
 ** All Fe is calculated as Fe²⁺
 n.d. = not detected (<0.02%).

unstrained, randomly oriented, inequigranular neoblasts, 20–100 μm in size. In mild cases, only a few neoblasts, sometimes euhedral in outline, are present in an optically continuous but strained host. In more advanced stages, large areas of grains are recrystallized, often around the margins. The most intense strain is manifest in complete recrystallization resulting in a 'mosaic dunite'. All these strain features may be present within a single olivine

Table 3. Representative garnet analyses

	Cr-rich megacryst	Cr-rich megacryst	Cr-poor megacryst	Garnet lherzolite	
				A	B
SiO ₂	41.4 (6)*	41.4 (6)	41.5 (1)	41.3 (2)	42.3 (4)
TiO ₂	0.36(5)	0.68(4)	0.43(1)	0.40(12)	0.48(2)
Al ₂ O ₃	22.6 (2)	17.5 (6)	22.3 (2)	17.8 (2)	21.4 (2)
Cr ₂ O ₃	1.69(3)	6.13(44)	0.07(2)	6.48(7)	2.36(3)
MgO	20.4 (3)	20.0 (3)	18.0 (2)	21.2 (3)	22.0 (2)
FeO**	8.53(8)	7.85(23)	12.4 (2)	5.91(8)	6.96(6)
MnO	0.52(2)	0.43(13)	0.42(5)	0.28(2)	0.22(1)
CaO	4.10(7)	6.64(4)	4.58(3)	5.85(4)	4.48(6)
Total	99.60	100.63	99.60	99.22	100.20
Cations on the basis of 12 oxygens					
Si	2.992	2.996	3.007	2.993	2.992
Ti	0.012	0.036	0.023	0.021	0.024
Al	1.908	1.496	1.896	1.524	1.785
Cr	0.096	0.350	0.002	0.370	0.130
Mg	2.172	2.157	1.944	2.292	2.318
Fe	0.504	0.471	0.752	0.357	0.410
Mn	0.024	0.020	0.025	0.017	0.012
Ca	0.312	0.510	0.352	0.450	0.335
Σ	7.990	8.030	8.001	8.021	8.007

* The numbers in parentheses represent 1σ standard deviations in terms of the last digit cited for the value to their immediate left, thus 41.4 (6) indicates a standard deviation of 0.6
 ** All Fe is calculated as Fe²⁺

Table 4. Representative orthopyroxene analyses

	Garnet lherzolite	Inclusion Cr-rich	Inclusion Cr-poor
	SiO ₂	57.1 (6)*	57.1 (6)
TiO ₂	0.14(2)	0.18(2)	0.09(2)
Al ₂ O ₃	1.11(2)	1.12(2)	0.70(1)
Cr ₂ O ₃	0.29(1)	0.41(1)	0.23(3)
MgO	35.2 (7)	34.0 (3)	35.8 (3)
FeO**	4.78(8)	5.98(22)	5.13(14)
MnO	0.06(2)	0.16(4)	0.14(2)
CaO	1.09(1)	1.20(4)	0.52(1)
Na ₂ O	0.19(3)	0.24(1)	0.22(5)
Total	99.96	100.39	100.73
Cations on the basis of 6 oxygens			
Si	1.962	1.968	1.973
Ti	0.003	0.004	0.002
Al	0.044	0.044	0.028
Cr	0.007	0.010	0.006
Mg	1.801	1.745	1.821
Fe	0.137	0.172	0.146
Mn	0.001	0.004	0.003
Ca	0.039	0.043	0.018
Na	0.012	0.016	0.014
Σ	4.005	4.006	4.009

* The numbers in parentheses represent 1σ standard deviations in terms of the last digit cited for the value to their immediate left, thus 57.1 (6) indicates a standard deviation of 0.6
 ** All Fe is calculated as Fe²⁺

megacryst. Many of the olivines possess a distinct rim, generally <200 μm wide, optically distinct from the core.

The Cr-rich suite olivines have core compositions in the range Fo 90–93, and the Cr-poor, cores of Fo 81–85. This latter group constitutes approximately one-third of the analyzed megacryst grains; all the olivines in the <1 mm size-range fall within the compositional range of the Cr-rich group. This group has Cr-contents in the range 0.05–0.10% Cr₂O₃. Neither the rims of Cr-rich olivines nor the Cr-poor group olivines possesses Cr-contents above nominal detection limits for routine analysis (0.02%). The discrete chemical nature of the groups is also reflected in their NiO and CaO contents (Fig. 1). The rim compositions of both groups fall in the range Fo 88–89. Within the Cr-rich olivines, there appears to be a correlation of NiO and CaO with decreasing Fo content, positive for NiO and negative for CaO. These are trends to be expected from olivines undergoing normal fractionation from a melt. However, many of the rims of these olivines are slightly lower in NiO and CaO than would be expected if these correlations were extrapolated to Fo 88.5. If linear correlations of NiO and CaO with Fo could be assumed over ranges of ~5 mol% Fo, then it would appear that the rims are diluted with respect to these minor components. The Cr-poor olivines contain low NiO and CaO.

These olivine megacrysts are typical in their petrographic and chemical features of olivines in kimberlites in general. Emeleus and Andrews (1975) and Boyd and Clement (1977) have described occurrences exhibiting olivines with core compositions in the range Fo 93 to Fo

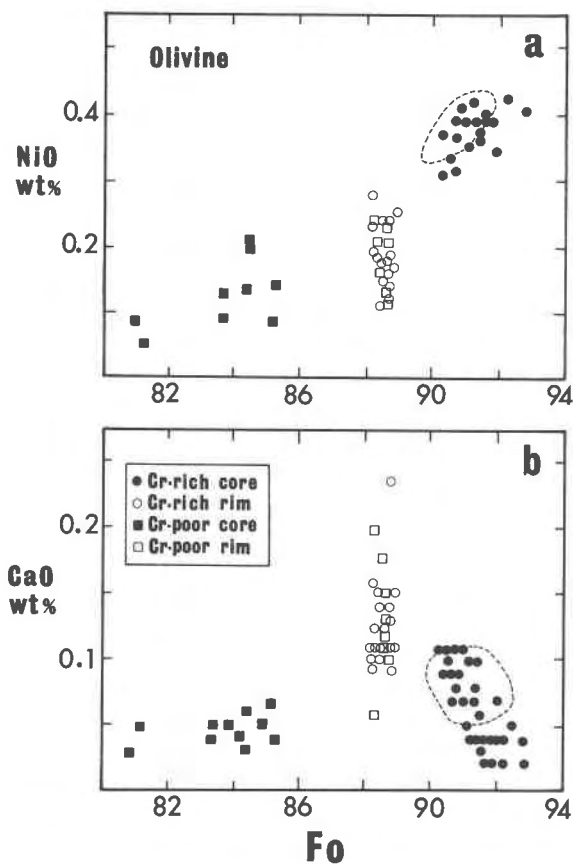


Fig. 1. Plots of NiO vs. Fo (a) and CaO vs. Fo (b) for olivine cores and rims. The outlined fields indicate the range of compositions in the garnet lherzolites.

80, where the rim compositions converge towards a constant value of Fo 88–89. In the Fayette County occurrence, the two populations are chemically distinct.

Garnet

Garnet is the second most abundant megacryst phase, forming 3 modal % of the >1 mm grains. Typically, grains are 0.5 to 2.0 cm in size and form flattened or oblate spheroids. Optically, the two types of garnet megacryst are different; those in the Cr-rich suite are wine-red, whereas the garnets in the less abundant Cr-poor suite are pale-orange. Both varieties have well-developed kelyphitic rims. These rims consist of spinel, aluminous Ca-rich and Ca-poor pyroxene, olivine, and glass, the whole reaction assemblage being mantled by interlocking phlogopite flakes. Assemblages of spinel, cpx, opx and glass also form veins and pockets within the garnets. In some cases, a residual kelyphitic assemblage is the only manifestation of the former presence of a garnet. These assemblages have resulted from incongruent melting of the garnets in the presence of alkalis and volatiles and

have been described in detail by Hunter and Taylor (1982).

The dichotomy of compositions recorded in the olivines is also reflected in the garnet-megacryst compositions (Fig. 2). The Cr-rich garnets have $mg = 0.85$ – 0.80 ; Cr_2O_3 ranges from 1.64 to 6.14%. The Cr-poor variety are more Fe-rich ($mg = 0.72$) and possess <0.1% Cr_2O_3 . Individual garnets are homogeneous; if compositionally different rims ever were present, they are obscured now by the kelyphitic rims described above.

Chemical identity of inclusions of Cr-rich olivines in the Cr-rich garnets and *vice versa* indicates co-precipitation of these phases. No inclusions of garnet in Cr-poor olivines or *vice versa* were observed. Within the Cr-rich garnets, there is a positive correlation of CaO and Cr_2O_3 . However, uvarovite substitution does not correlate significantly with mg , nor does TiO_2 which varies from 0.18 to 1.00%.

Pyroxene

Pyroxene does not occur as a megacryst/phenocryst phase in this kimberlite but rather as inclusions in olivine and garnet megacrysts (pyroxene also occurs within garnet-lherzolite nodules, described below). The inclusions are rounded and generally in the size range 0.5–1.0 mm.

Orthopyroxene inclusions are confined to the Cr-rich olivines, up to three inclusions occurring within a single megacryst. Within the eleven inclusions analyzed, mg is relatively constant at 0.93–0.91. However, with decreasing CaO content (a function of temperature), there is a systematic variation in minor components; CaO varies from 1.37 to 0.51%, Al_2O_3 from 1.14 to 0.69%, and Cr_2O_3 from 0.61 to 0.20%. Variation in TiO_2 and Na_2O is less systematic, values ranging from 0.08 to 0.24% and 0.31 to 0.18% respectively.

As might be expected, the clinopyroxene inclusions in the Cr-rich olivines and the Cr-poor garnets are chemically distinct (Table 5). The apple-green clinopyroxenes in the Cr-rich olivines are richer in Cr_2O_3 (1.89–2.34%), Al_2O_3 (2.11–2.63%), and Na_2O (1.55–2.81%) than the less variable, colorless variety in the Cr-poor garnets, which contain mean values of 0.11% Cr_2O_3 , 2.06% Al_2O_3 and 1.60% Na_2O . However, the Cr-poor clinopyroxenes are richer in TiO_2 (0.26%) than the Cr-rich (0.16%). The Cr-rich clinopyroxenes contain 84–88 mol% Di and the Cr-poor, 78–82 mol% Di.

Picroilmenite

Rounded to ovoid picroilmenite megacrysts, up to 1.5 cm in diameter, form the third most abundant group of minerals in the >1 mm size range (1–2 modal %). The majority of the megacrysts are monocrystalline; some show undulose extinction, bending of Ti-magnetite exsolution lamellae, and varying degrees of recrystallization to poly-crystalline mosaics. Reaction rims of Ti-magne-

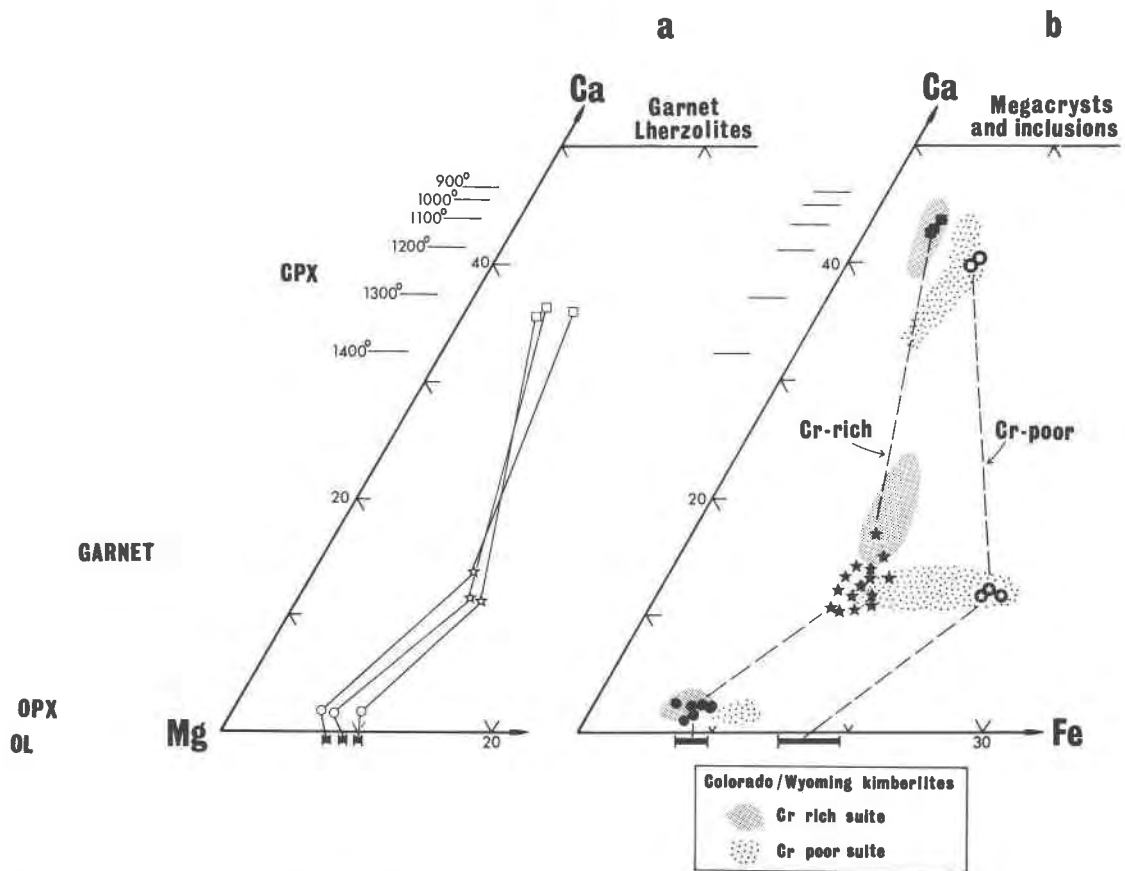


Fig. 2. Pyroxene, garnet, and olivine compositions in the Ca-Mg-Fe ternary. The temperature scale is adapted from the 20 kbar solvus of Lindsley and Dixon (1976). (a) Garnet lherzolite mineral compositions. Tie-lines join coexisting phases in the three xenoliths recovered. (b) Megacryst and inclusion compositions. The range in olivine compositions is indicated by the bar. The dashed lines indicate the general tie-line orientations of the Cr-rich and Cr-poor groups. The shaded areas are the ranges of compositions in the Sloan-Nix kimberlite pipes in Colorado and Wyoming (Eggler *et al.*, 1979) for comparison.

tite and perovskite are a ubiquitous feature; these fretted rims are usually 100–200 μm in thickness. The reaction mantles are described in detail in the accompanying paper (Hunter *et al.*, 1983). Euhedral ilmenite grains, 30–50 μm in size, form abundant inclusions in the Cr-poor garnet megacrysts.

The compositional range of the ilmenites, with respect to their principal end-member components, ilmenite (FeTiO_3), geikielite (MgTiO_3), and hematite (Fe_2O_3), is shown in Fig. 3. The analyses plot within the kimberlite field as defined by Mitchell (1977). The core compositions range from 16–37 mol% MgTiO_3 ; the rims are enriched in MgO (34–51% MgTiO_3) relative to the cores. Microprobe traverses across the megacrysts (Fig. 4) indicate that this Mg-enrichment is restricted to the outer 200–300 μm , with a relatively smooth compositional profile. This Mg-enrichment is a common feature of kimberlitic microilmenite megacrysts (*e.g.*, Haggerty *et al.*, 1979; Elthon and Ridley, 1979; Boctor and Boyd, 1980a; Pasteris, 1980; Agee *et al.*, 1982), and its origin has been the subject of considerable debate. $\text{Fe}^{3+}/$

($\text{Fe}^{3+} + \text{Fe}^{2+}$) shows a gradual but slight increase from core to rim (Fig. 4). There is variability in the Cr_2O_3 and Fe_2O_3 contents of the megacryst cores (Fig. 5). The rims possess a restricted range of Fe_2O_3 , but Cr_2O_3 varies by a factor of two. The rims are enriched in Cr_2O_3 (1.4–3.6%) relative to the cores (0.05–2.05%). The ilmenite inclusions in the Cr-poor garnet megacrysts are similar in composition to the megacryst cores (Table 6), containing 24–33 mol% MgTiO_3 , although their Fe^{3+} contents are lower. The enclosing garnets apparently shielded these ilmenites from reactions which formed the MgO increase in the megacryst rims and from subsequent changes in $f\text{O}_2$.

Phlogopite

Subhedral to ovoid phlogopite megacrysts, up to 1 cm in size, are rare in comparison with the other megacryst phases. No optical zonation is evident within the reverse pleochroic cores; the rims are strongly abraded, but thin (50–100 μm), partial rims of normally pleochroic phlogopite, similar to the groundmass micas, are present on two

Table 5. Representative clinopyroxene analyses

	Garnet Iherzolite	Inclusion Cr-rich	Inclusion Cr-poor
SiO ₂	54.8 (2)*	53.9 (4)	54.4 (11)
TiO ₂	0.29(1)	0.17(1)	0.25(5)
Al ₂ O ₃	2.06(8)	2.63(6)	2.07(9)
Cr ₂ O ₃	0.82(7)	2.18(6)	0.11(1)
MgO	19.5 (5)	16.4 (2)	16.8 (2)
FeO**	3.78(5)	22.87(5)	5.22(22)
MnO	0.12(2)	0.14(1)	0.09(1)
CaO	17.6 (6)	18.6 (3)	18.7 (1)
Na ₂ O	1.37(5)	2.48(9)	1.64(19)
Total	100.34	99.37	99.28
Cations on the basis of 6 oxygens			
Si	1.966	1.963	1.990
Ti	0.007	0.004	0.007
Al	0.086	0.112	0.090
Cr	0.023	0.062	0.003
Mg	1.042	0.891	0.916
Fe	0.112	0.086	0.160
Mn	0.003	0.003	0.002
Ca	0.676	0.762	0.731
Na	0.095	0.174	0.114
Σ	4.011	4.022	4.013

* The numbers in parentheses represent 1σ standard deviations in terms of the last digit cited for the value to their immediate left, thus 54.8 (2) indicates a standard deviation of 0.2
 ** All Fe is calculated as Fe²⁺.

of the megacrysts. On the basis of their chemistry, there is no rationale for allocating any of these megacrysts to either Cr-rich or Cr-poor suites. The cores are low in Cr₂O₃ (<0.5%) and TiO₂ (<0.7%), Al₂O₃ is in the range 11–12%, and total iron as FeO varies from 3.45 to 8.46%. They are similar to megacryst phlogopites reported from other occurrences in their low Cr₂O₃ and TiO₂ (e.g., Smith *et al.*, 1978; Delaney *et al.*, 1980), but their FeO contents overlap the range of MARID suite micas (Dawson

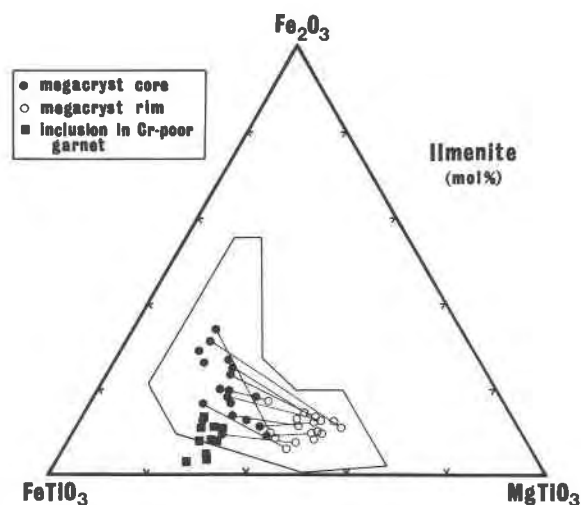


Fig. 3. Picroilmenite compositions plotted in the ilmenite (FeTiO₃)–hematite (Fe₂O₃)–geikielite (MgTiO₃) ternary. Tie-lines join coexisting cores and rims. The outlined area is the kimberlitic ilmenite field defined by Mitchell (1977).

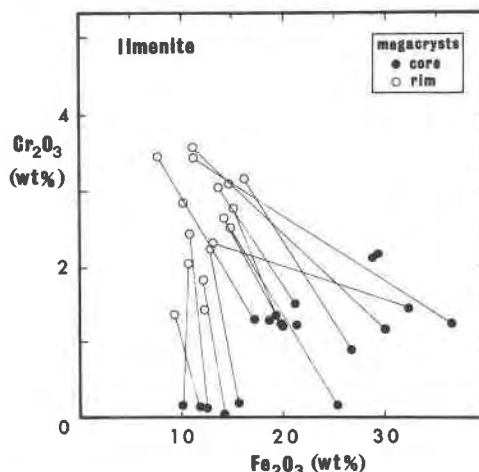


Fig. 4. Ilmenite core and rim compositions on a plot of Cr₂O₃ vs. Fe₂O₃. Tie-lines join coexisting core and rim analyses.

and Smith, 1977). They are chemically distinct from the suite of phlogopites present as phenocrysts, as mantles on garnet kelyphite rims, and in the groundmass of this kimberlite, which themselves show systematic chemical variability (Hunter *et al.*, 1984).

Sulfides

Spherical inclusions, up to 1.5 mm in diameter, composed of sulfide phases occur in the Cr-rich olivine megacrysts. They represent the second most abundant inclusion type in the Cr-rich suite (after orthopyroxene—see Table 1), and several inclusions may be present within a single megacryst; one olivine megacryst contains inclusions of both sulfide and opx. The blebs are composite and consist of intimate intergrowths of pentlandite and pyrrhotite. Chalcopyrite constitutes a minor phase and is concentrated at the margins of the blebs. No analyses of

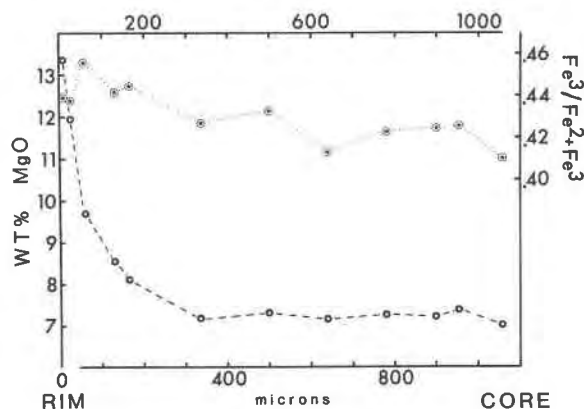


Fig. 5. Core to rim microprobe traverse across an ilmenite megacryst indicating changes in MgO (lower curve) and Fe³⁺/(Fe³⁺ + Fe²⁺) (upper curve). Note that the Mg-enrichment is confined to the outer 300 μm.

Table 6. Representative oxide analyses

	Ilmenite core	megacryst rim	Ilmenite inclusion	Cr-spinel inclusion
TiO ₂	44.1 (5)*	48.7 (5)	47.9 (3)	6.00(6)
Al ₂ O ₃	0.18(2)	0.65(9)	1.30(2)	10.1 (1)
Cr ₂ O ₃	1.25(4)	2.65(16)	0.14(2)	37.4 (2)
Fe ₂ O ₃ **	19.8	14.3	13.8	14.3
MgO	7.29(7)	14.3 (2)	8.27(8)	13.7 (4)
FeO	26.6 (3)	17.9 (3)	28.3 (3)	17.1 (3)
MnO	0.29(2)	0.49(2)	0.32(2)	0.36(2)
Total Cation basis	99.51 (3)	98.99 (3)	100.03 (3)	98.96 (4)
Ti	0.802	0.841	0.856	0.147
Al	0.005	0.018	0.036	0.388
Cr	0.024	0.048	0.003	0.968
Fe ³⁺	0.361	0.248	0.245	0.351
Mg	0.263	0.490	0.292	0.668
Fe ²⁺	0.539	0.345	0.562	0.468
Mn	0.006	0.010	0.006	0.010
Σ	2.000	2.000	2.000	3.000

* The numbers in parentheses represent 1σ standard deviations of replicate analyses in terms of the last digit cited for the value to their immediate left, thus 44.1 (5) indicates a standard deviation of 0.5

** Fe³⁺ is calculated assuming stoichiometry

these sulfides were performed for this study; texturally, the inclusions resemble features described by Boctor and Boyd (1980b) and Haggerty and Tomkins (1982). The assemblages present now have certainly re-equilibrated on cooling, but the original blebs undoubtedly represent an immiscible sulfide liquid entrapped during growth of the Cr-rich olivine megacrysts.

Cr-spinel

Two euhedral inclusions of titanian Mg-Al-chromite, 30 μm in size, have been recorded from the outer portion of the cores of Cr-rich olivine megacrysts. Compositionally (Table 6), they plot at the Cr-rich end of the spectra of spinels present as inclusions in phlogopite phenocrysts and as discrete euhedral-subhedral grains within the groundmass of this kimberlite. These mid- to late-stage phases are described in detail in the accompanying paper (Hunter *et al.*, 1984).

Ultramafic xenoliths

Mantle-derived xenolithic material is uncommon in this kimberlite; despite extensive slabbing of collected material, only three garnet lherzolite xenoliths were recovered. Like the megacrysts, these xenoliths are ovoid in shape and have similar dimensions (1.5 × 2.5 cm). All are porphyroclastic to mosaic-textured; rounded porphyroclasts of wine-red garnet, pale-green diopside, and colorless enstatite, 1–4 mm in size, are set in a mosaic of inequigranular olivine neoblasts, generally <100 μm in size. The pyroxenes are not free of strain features; locally, the enstatite is partially neoblastic and diopsides

commonly possess strong, stress-induced cleavage. Garnets are fractured to varying degrees and are partially or wholly replaced by symplectic intergrowths of spinel + pyroxenes + glass (described in detail by Hunter and Taylor, 1982). Locally, thin mantles of secondary phlogopite are present partially surrounding these reaction assemblages.

The compositions of garnet-lherzolite phases are presented in Tables 2–5 and 7, and are plotted in the Mg–Ca–Fe ternary (Fig. 2a). The clinopyroxenes contain 70–74 mol% Di, and Ca/(Ca+Mg+Fe) in the garnets varies from 0.10 to 0.15; correlating positively with Cr₂O₃ contents. On the basis of Cr₂O₃ content in the garnets and pyroxenes, one of the xenoliths (xenolith A in Table 3) is different from the other two. This sample contains mean values of 6.48%, 1.45% and 0.47% Cr₂O₃ in garnet, diopside and enstatite, respectively; in comparison, the other two contain 2.28%, 0.80–0.84% and 0.28–0.32% Cr₂O₃, in these three phases, respectively. Olivine compositions vary from Fo 90–93; their ranges in NiO and CaO are indicated in Figure 1. Texturally and mineralogically, the garnet lherzolites are similar to sheared garnet lherzolites from other kimberlite localities worldwide (e.g., Boyd and Nixon, 1978).

Mosaic dunite is relatively abundant in comparison with the lherzolite xenoliths. However, as noted above, complete, stress-induced recrystallization of olivine megacrysts may result in formation of a "mosaic dunite"; it is not apparent, in these cases, whether these represent primary, fragmented, lherzolite assemblages, or recrystallized megacrysts. In some cases, mineral inclusions are present in partially recrystallized olivines. That these are not xenoliths is indicated by regions of optically continu-

Table 7. Salient features of the megacryst, inclusion, and garnet lherzolite mineral chemistry

	Mg/Mg+Fe ²⁺	Ca/Ca+Mg+Fe ²⁺	Cr ₂ O ₃ wt%
Cr-rich Suite:			
Olivine (core)	0.90–0.93		0.05–0.10
(rim)	0.88–0.89		<0.02
Garnet	0.81–0.85	0.10–0.17	1.65–6.13
Diopside	0.91–0.92	0.42–0.44	1.89–2.34
Enstatite	0.91–0.93	0.090–0.025	0.20–0.61
Cr-poor Suite:			
Olivine (core)	0.81–0.85		<0.02
(rim)	0.88–0.89		<0.02
Garnet	0.71–0.72	0.11–0.12	0.07–0.09
Diopside	0.84–0.86	0.39–0.41	0.10–0.12
Ilmenite (core)	0.16–0.37*		0.05–1.50
(rim)	0.34–0.37*		1.42–3.74
Garnet Lherzolite:			
Olivine	0.90–0.92		0.07–0.10
Garnet	0.85–0.87	0.10–0.15	2.28–6.48
Diopside	0.90–0.93	0.35–0.37	0.80–1.54
Enstatite	0.91–0.93	0.020–0.022	0.28–0.48

*Mg/Mg + Fe²⁺ + Fe³⁺ in ilmenite.

ous but strained olivine within the recrystallized mosaic. If this recrystallization had been complete, there would be no way of telling petrographically if the mosaic had been a former megacryst or a xenolith. However, true xenoliths are rare in this kimberlite, and their mineral chemistry is sufficiently different from the observed inclusions to prevent ambiguity in this regard.

Geothermobarometry

Before undertaking a synthesis of the petrographic and mineral chemical data, it is pertinent to assess the pressure and temperature conditions under which the various assemblages appear to have equilibrated. Numerous mineral geothermometers applicable to garnet–two pyroxene–olivine assemblages are present in the literature. Typically, application of different thermometers, or even different calibrations of the same thermometer, return values that may be discordant by as much as 200–300°C. The problems associated with the various methods have been reviewed by Carswell and Gibb (1980). Several methods were applied to the megacryst and lherzolite assemblages of the Fayette County kimberlite, and as expected, discordant temperatures were returned. It is not intended herein to make a detailed evaluation of the relative merits of the different methods. Instead, we have applied methods which have been widely used for these assemblages in order to be able to compare our results with other reported occurrences in the literature. For temperatures, we have applied the 20 kbar diopside solvus limb of Lindsley and Dixon (1976) and the empirical enstatite solvus limb of Boyd and Nixon (1973). Using these values, we have calculated pressures of equilibration using the Al_2O_3 content of orthopyroxene coexisting with garnet (Wood, 1974). The reader's attention is drawn to two points. First, the calculated 40 kbar solvus of Lindsley (1980) indicates that temperatures returned from the 20 kbar solvus of Lindsley and Dixon (1976) may be underestimated by as much as 100°C. Second, the Wood (1974) isopleths for Al_2O_3 in orthopyroxene coexisting with garnet were generated by modelling the data of MacGregor (1974). Recent investigations in the system $\text{MgO}-\text{Al}_2\text{O}_3-\text{SiO}_2$ from 15–40 kbar and 900–1600°C using only reversed determinations (Perkins *et al.*, 1981) show lower Al_2O_3 contents in orthopyroxene at a given P and T , in comparison with MacGregor's results over most of the garnet field. These data indicate that for a given T and Al_2O_3 content in orthopyroxene, the pressures calculated from the Wood (1974) expression will be over-estimated by 3–5 kbar.

Use of the pyroxene solvus thermometer for the megacryst/inclusion suite requires that the two pyroxenes were in equilibrium. Although no contact relationships were observed, both clino- and orthopyroxene are present as inclusions in the Cr-rich olivines. Pressure calculations for the orthopyroxene inclusions requires that this phase be in equilibrium with garnet; the presence of Cr-rich

garnet inclusions in the Cr-rich olivine megacrysts justifies this assumption. For the purposes of these calculations, a mean Cr-rich garnet megacryst composition was chosen with $\text{Ca}/(\text{Ca}+\text{Mg}+\text{Fe}) = 0.129$ and $\text{Cr}/(\text{Cr}+\text{Al}) = 0.114$. Data obtained from the Cr-rich pyroxenes and the garnet lherzolite xenoliths have been plotted in Figure 6.

The garnet lherzolites yield temperatures in the range 1330–1250°C and pressures of 49–51 kbar. These pressures correspond to a depth of approximately 150 km and, together with the temperatures, place the equilibration conditions well above the static shield geotherm of Clarke and Ringwood (1964), but outside the stability field of diamond.

The Cr-rich orthopyroxene inclusions equilibrated over a temperature range from 1310–1055°C and at pressures between 39–48 kbar, corresponding to a depth range of 120–150 km. The data points appear to form clusters in P – T space (Fig. 6); however, this possibly reflects the limited sampling, since the suite shows systematic chemical changes with decreasing CaO (*i.e.*, temperature). The clinopyroxene inclusions return temperatures of 1115–1165°C; this limited range is also probably caused by a sampling problem.

The equilibration conditions of the Cr-poor suite cannot be constrained. Since only Ca-rich pyroxene is present, there is no basis for assuming that the chemistry of the pyroxene could be solvus controlled. However, Mg/Fe distribution between garnet and clinopyroxene (Akella and Boyd, 1974) has yielded temperatures of 1184–1227°C. In the absence of coexisting orthopyroxene, no estimates of pressure can be made for this group.

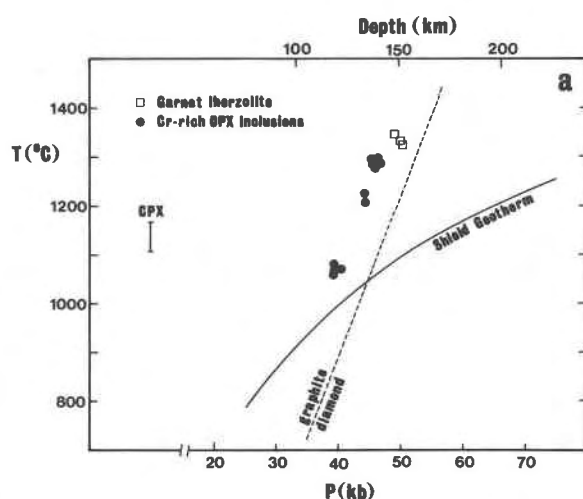


Fig. 6. A plot of temperature (°C) vs. pressure (kbar) and depth (km) for Cr-rich orthopyroxene inclusions and garnet lherzolite xenoliths. The temperature range for the Cr-rich clinopyroxene inclusions is shown on the left. The static-earth geotherm of Clarke and Ringwood (1964) and the diamond-graphite stability curve of Bundy *et al.* (1961) are shown for reference.

Provenance of the megacrysts

A notable feature of the megacryst and inclusion suites in the Fayette County kimberlite is the dichotomy of mineral compositions. Populations of megacrysts with contrasting chemical characteristics have been noted from other kimberlite occurrences, perhaps the best characterized being the Cr-rich and Cr-poor suites in the Colorado–Wyoming Sloan–Nix kimberlites (McCallum *et al.*, 1975; Egger *et al.*, 1979). The disparity of compositions between the Cr-rich and Cr-poor megacryst populations in the Fayette County kimberlite is manifest principally in *mg* and Cr₂O₃ contents and these two groups correlate, in these regards, with the analogous suites reported in the Colorado–Wyoming kimberlites.

The range of compositions of pyroxenes and garnets in the Colorado–Wyoming pipes is indicated in Figure 2b. The Fayette County Cr-rich pyroxenes plot within the field defined by the Colorado–Wyoming Cr-rich suite. Furthermore, their Cr-contents (1.89–2.34% Cr₂O₃ in cpx; 0.20–0.61% in opx) are similar to those of the Colorado–Wyoming Cr-rich suite (0.83–2.40% Cr₂O₃ in cpx; 0.39–0.69 in opx). The Cr-rich garnets in the Fayette County occurrence (1.65–6.13% Cr₂O₃) span the Cr₂O₃ range of the Colorado–Wyoming Cr-poor garnets (0.03–4.80% Cr₂O₃) and the gap to the Cr-rich garnet (6.3–13.0% Cr₂O₃). The positive correlation of CaO and Cr₂O₃, the trend in Figure 2b, and association of Cr-rich garnets with Cr-rich pyroxenes in the Fayette County kimberlite, suggests that they are lower-Cr representatives of the Cr-rich group of Egger *et al.* (1979). An additional similarity between the Cr-rich groups in both occurrences is their restricted range in *mg*. The Cr-poor suite in the Colorado–Wyoming kimberlites is more Fe-rich than the Cr-rich suite, and the garnets show a wide range in *mg* at constant Ca/(Ca+Mg+Fe) (Fig. 2b). The Cr-poor garnets from the Fayette County kimberlite plot within this Cr-poor range, near its Fe-rich limit. The Cr-poor clinopyroxene inclusions lie close to the field of Cr-poor clinopyroxene megacrysts in the Colorado–Wyoming kimberlites and correlate with the steeper part of the trend defined by the Sloan–Nix pipes.

The Cr-poor suite

Mineral compositions in the Cr-poor suite are similar to the more Fe-rich compositions of megacrysts and inclusions in the Cr-poor suites of kimberlites from Northern Lesotho (Nixon and Boyd, 1973a), the Monastery kimberlite (Gurney *et al.*, 1979), and the Colorado–Wyoming kimberlites, among others. Egger *et al.* (1979) suggest that the Cr-poor suite represents the “normal” trend of fractional crystallization within kimberlites at high *P* and *T*. Harte and Gurney (1981) proposed a model in which complex magma sheets and apophyses exist in a temperature gradient, allowing the simultaneous existence of different stages of magma differentiation under essentially isobaric conditions. The attraction of such a model lies in

circumventing the necessity for extensive magma columns existing throughout a *P–T* gradient.

Any consideration of the provenance of Cr-poor megacrysts in kimberlitic rocks must involve the origin of ilmenite megacrysts and ilmenite-silicate intergrowths. This has been the subject of considerable debate in the last decade (*e.g.*, Boyd and Nixon, 1973; Dawson and Reid, 1970; Haggerty *et al.*, 1979; Garrison and Taylor, 1981; Mitchell, 1973; 1977; Pasteris, 1980; Wyatt, 1975). Recently, Pasteris (1980) proposed that the megacrysts are xenocrystic in origin; they crystallized in the low velocity zone (LVZ) from a mantle partial melt that is unrelated to kimberlite. Kimberlite-bearing diapirs rise through the LVZ and disrupt the ilmenite megacryst mush, incorporating the ilmenites as xenoliths. Mitchell (1973; 1977) interpreted the ilmenite megacrysts to be high-pressure phenocrysts precipitated during the early stages of evolution of a proto-kimberlite melt. The early formation of ilmenite is not compatible with much of the more recent data on kimberlitic megacryst suites. However, Mitchell also noted that a xenocrystic origin could not be discounted.

A general feature of kimberlites with ilmenite megacrysts and pyroxene-ilmenite intergrowth/inclusion relations is that the coexisting diopside is calcic, typically yielding solvus temperatures at the lower end of the temperature interval exhibited by the diopside megacryst suite as a whole. On the basis of these relationships, ilmenite is usually considered to be a constituent of the Cr-poor megacryst suites (*e.g.*, Boyd and Nixon, 1978; Gurney *et al.*, 1979; Egger *et al.*, 1979). In the Monastery kimberlite (Gurney *et al.*, 1979), the earlier silicate phases (*i.e.*, higher *mg*; higher *T*) show an increase in TiO₂ with decreasing *mg* until ilmenite appears as an inclusion phase (and by inference a megacryst phase), whereafter TiO₂ decreases. The incoming of ilmenite as a liquidus phase occurs at *mg* = 0.74 in the garnets, at temperatures (inferred from the diopsides) close to 1300°C.

The bulk of the evidence from kimberlites containing silicate megacrysts with a wide-crystallization interval is consistent with ilmenite megacrysts being late-stage phenocrysts from a Cr-poor kimberlitic melt. Ilmenite is not commonly found as a phase within sheared, garnet-hercynite xenoliths in kimberlite. These undepleted mantle compositions (plus H₂O and CO₂) are believed to represent the parent from which kimberlite is derived by small degrees of partial melting (*e.g.*, Wyllie, 1979; Ellis and Wyllie, 1979). The evidence presented above indicates that ilmenite probably is not a liquidus phase in these early partial-melts, but these melts intersect an ilmenite saturation surface after some degree of fractional crystallization.

An interesting point arises from consideration of crystal/melt FeO/MgO *K_D*'s (*e.g.*, Schulze and Hoover, 1982). The *K_D*'s for olivine/melt and pyroxene/melt are less than unity and the *K_D* for garnet/melt is probably near to one; the *K_D* for ilmenite is considerably greater

than this (probably 2–5). Before ilmenite saturation, crystallization of silicate phases will deplete residual liquids in Mg. However, soon after ilmenite saturation, the *bulk* crystal/melt K_D may approach unity; consequently, *mg* in residual liquids could be buffered at or near a value of unity, and continued fractionation would not change *mg* in the melt significantly. Two lines of evidence are in support of this contention. First, in kimberlites where a considerable temperature interval is exhibited by the Cr-poor pyroxenes (*e.g.*, Sloan–Nix), the early part of the trend shows a decrease in *mg* with decreasing temperature; after ilmenite saturation (as evidenced by the appearance of intergrowths and inclusions), the trend shows an inflection towards constant *mg* with decreasing temperature (see Fig. 2). Second, very few kimberlites possess olivines with *mg* < 0.80 or garnets with *mg* < 0.70, suggesting that a near-constant *mg* is reached in the melt during crystal fractionation.

The Cr-rich suite

Whatever the origin of the Cr-poor megacryst suite, we must consider it to be generally applicable, since Cr-poor suite megacrysts are the rule rather than the exception in kimberlitic rocks. On the other hand, the Cr-rich suite apparently is of restricted occurrence, and their provenance is more problematical.

The Fayette County kimberlite Cr-rich megacryst and inclusion data show some correspondence with data for garnet lherzolites reported extensively in the literature (*e.g.*, Boyd and Nixon, 1978; Cox *et al.*, 1973; MacGregor, 1979; Nixon and Boyd, 1973). Indeed, there is overlap in the major element compositions with the garnet lherzolite data for this occurrence, although minor element ranges differ. In consideration of these compositional similarities, the possibility arises that the Cr-rich megacrysts may have been derived by disaggregation of garnet peridotite wallrocks and xenoliths.

Typically, grain sizes within mantle-derived peridotite xenoliths fall within the range 2–4 mm; in the “coarse-grained” peridotites, grain sizes may be as great as 1 cm. Only in rare cases have lherzolites been recovered in which grains are coarser than 1 cm (*e.g.*, Boyd and Nixon, 1978). In the Fayette County kimberlite, the Cr-rich megacrysts are up to 3 cm in size. It would appear unlikely that these largest grains could be derived by disaggregation of mantle peridotite. The presence of inclusions within these megacrysts with grain-sizes in the 0.5–1.5 mm range clearly indicates that grain sizes in the Cr-rich suite, as a whole, ranged from 0.5 mm to at least 3 cm. There seems no *a priori* reason to assume that just because crystals in a rock have the same size as those in the medium through which they passed, they represent disaggregated products of this medium (garnet lherzolite, in this case). Sizes of crystals grown from melts are a function of various kinetic parameters; a seriate size distribution is not unexpected in kimberlitic rocks, which may have a complex cooling history. Some of the inclu-

sions are euhedral to subhedral in form. This, together with evidence for coexistence of an immiscible sulfide melt phase, is more consistent with the host megacryst phase crystallizing from a melt rather than merely representing mechanical disaggregation of wall rock material. A further point of note is the absence of megacrystic pyroxene in this occurrence, a feature atypical of kimberlites in general. If the Cr-rich population were derived by disaggregation of peridotite, it would be expected that pyroxene megacrysts would be at least as abundant as garnets in the present rock.

While the disaggregation mechanism remains a possibility for the Cr-rich suite, we feel that the weight of evidence supports an origin through fractional crystallization of a Cr-rich melt at elevated *P* and *T*. The Di and Al₂O₃ contents of the Cr-rich orthopyroxene inclusions are consistent with equilibration over a temperature interval of approximately 300°C within the pressure range 40–50 kbar. If the parental melt were fractionating by removal of olivine + garnet + two-pyroxenes, it would be expected that *mg* in the liquid would decrease with increasing degrees of fractionation. However, the *mg* range of the Cr-rich assemblage is relatively restricted (see Table 7). In addition to the increase in Ca and Al recorded in the orthopyroxenes, the olivines show a correlation of increasing Ca and decreasing Ni with decreasing Fo contents, even within the restricted range from Fo 93–90. A mechanism must be sought to allow *mg* to remain relatively constant while minor elements vary during fractionation. An increase in *f*O₂ relative to mantle buffer curves over the observed temperature interval would have the effect of reducing Fe³⁺/(Fe²⁺ + Fe³⁺) and hence contribute to maintaining constant *mg*. However, the data of Eggler (1982) suggest that megacryst suites in kimberlites crystallized under buffered *f*O₂. Buffering of the mineral compositions by the host mantle (garnet lherzolite) should result in little change in *mg* during fractionation but would require low melt/solid ratios. Such low ratios are indicated at some stage during the evolution of the megacrysts, since many of them exhibit shearing and recrystallization. Buffering could result in major element concentrations remaining fairly constant (similar to the host lherzolite), whereas minor elements could fractionate by a zone refining process. Assimilation of refractory lithosphere could also contribute to maintaining constant major element patterns; the net result would be a buffering similar to that caused by zone refining. An assimilation mechanism was favored by Eggler *et al.* (1979). An attractive mechanism for maintaining constant *mg* values in a silicate melt, is by immiscible sulphide melt fractionation. This would be a contributing factor in the Fayette County Cr-rich suite, as sulphide blebs are not uncommon as inclusions. The bulk *mg* of the fractionating Cr-rich silicate assemblage based on modal proportions and compositions of the phases (*i.e.*, approx 2% cpx; 3% opx; 5% gar; 90% ol) would be ~0.90. The effect of removal of 5% of immiscible sulfide

melt (containing negligible Mg) from the silicate melt would be to change the bulk mg in the fractionating assemblage to ~ 0.86 . Clearly, the effect of only small amounts of sulphide melt (in comparison to the bulk silicate phases) is quantitatively significant, and may be sufficient to reduce the change in mg in the host melt during fractionation.

It is not possible to evaluate the relative roles of these processes in contributing to the restricted range in mg in the Cr-rich assemblage. Immiscible sulfide melt fractionation alone may account for the observed effect, but it does not account for the high Cr-contents. Zone refining and/or assimilation may both be contributing factors and are to be expected in partially melted diapirs where relatively low melt/solid ratios may be present. In addition, the high Cr-contents may reflect partial melting of a source region that is heterogeneous with respect to chromium. Clearly, the range of Cr-contents observed in minerals of the three lherzolites recovered from the Fayette County kimberlite indicate that such heterogeneities do exist.

Magma-mixing

The bulk of the evidence presented herein supports the existence of two chemically distinct populations of megacrysts/inclusions within the Fayette County kimberlite: (1) A Cr-rich suite recording high-pressure crystal and immiscible-melt fractionation over a temperature interval of approximately 300°C, in which mg did not vary significantly; and (2) A Cr-poor suite representing a later-stage in the evolution of a Cr-poor kimberlitic melt, also at elevated P and T .

Clearly, since both suites are intimately intermingled in the present rock, a mixing event must have taken place during the evolution of the kimberlite. The convergence of the Fo contents of the two olivine populations to a constant value of Fo 88.5 provides evidence to support the contention that the host melts of the two populations mixed thoroughly, resulting in an *homogeneous hybrid melt*. Mixing of the two disparate chemistries will result in a composition linearly intermediate between the two end-members. The NiO and CaO contents of the Cr-rich suite olivines indicate that the host melt was relatively enriched in the former and relatively depleted in the latter. The Cr-rich melt was undergoing enrichment in CaO and depletion in NiO as evidenced by their correlations with Fo content in the Cr-rich suite olivines (Fig. 1). Mixing of the two melts would result in NiO and CaO contents intermediate between the two. This is reflected in the dilution of both components in the Fo 88.5 rims of the Cr-rich suite olivines below the levels predicted. The Cr-poor suite rims similarly are enriched to levels higher than would be expected from back-extrapolation to Fo 88.5.

A principal reason for the debate concerning the origin of ilmenite megacrysts in kimberlites is the common,

reverse-zonation of MgO observed within the rims of these megacrysts (*e.g.*, Agee *et al.*, 1982; Boctor and Boyd, 1980a; Haggerty *et al.*, 1979; Pasteris, 1980). This enrichment in MgO is often associated with an increase in Cr₂O₃. These reverse-zoned rims led Pasteris (1980) to the conclusion that the ilmenite megacrysts are xenocrystic in origin; they are derived from a mantle partial melt that is unrelated to kimberlite. Haggerty *et al.* (1979) attributed the increase to a magmatic trend; a function of decreased P_{tot} . Boctor and Boyd (1980a) and Agee *et al.* (1982) suggest back-reaction of the ilmenite megacrysts within an environment of increased Mg-activity; the former authors calling upon a metasomatic Mg-rich fluid and the latter, incorporation of Mg-carbonate peridotite. In the Fayette County kimberlite, the megacrysts of the Cr-poor assemblage also possess rims enriched in MgO and Cr₂O₃ relative to the cores, indicating an increase in Mg-, and Cr-activity. *This reverse zonation would be a natural consequence of the mixing of the Cr-poor assemblage with the host melt of the Cr-rich (and Mg-rich) suite*, and is interpreted as such. Possible evidence of this mixing in the garnet megacrysts of the present study has been obliterated by their later-stage, metasomatic melting.

Many of the models for the origin of megacrysts in kimberlite, particularly ilmenite megacrysts, requires them to be xenocrysts within their host kimberlitic melt. The megacrysts are disrupted from cumulates crystallized from partial melts within the LVZ and were scavenged by kimberlite-bearing diapirs (*e.g.*, Pasteris, 1980), or the kimberlites represent the interstitial melt from these mushes (*e.g.*, Gurney *et al.*, 1979). Other authors have suggested the megacrysts crystallized from a "proto-kimberlitic melt" (*e.g.*, Mitchell, 1973; Egger *et al.*, 1979; Garrison and Taylor, 1980). Indeed, derivation of the megacryst phases from a kimberlitic melt at high P and T is consistent with the experiments of Egger and Wendlandt (1979). Furthermore, melting relations in the system peridotite-CO₂-H₂O (Ellis and Wyllie, 1979; Wyllie, 1979; Wyllie, 1980) indicate the kimberlitic melts can exist throughout the depth interval of the LVZ (260–80 km). Therefore, there is no *a priori* reason to assume that the megacrysts *cannot* be derived from these kimberlitic melts; the bulk of the evidence from the megacrysts can be interpreted within the context of such an origin (*i.e.*, the megacrysts are cognate to the kimberlite—Egger *et al.*, 1979).

Model

Any model for the origin of the Fayette County kimberlite containing megacrysts must include explanation of the following observations: (1) two discrete suites of megacrysts occurring within a single intrusion—a Cr-rich suite possessing a restricted range of mg , and a Cr-poor suite exhibiting a wide variation in mg ; (2) the presence of ilmenite; (3) reverse zonation within the rims of ilmenite

megacrysts, and reverse *and* normal zonation in olivine rims within the same intrusion; (4) a wide range of equilibration temperatures (200–300°C) exhibited by both the Cr-rich and Cr-poor suites of megacrysts; and (5) shearing within both megacrysts and xenoliths.

The possible modes of formation of the two megacryst suites has been discussed above. Any model for the origin of the Cr-poor megacryst suites, such as outlined by Harte and Gurney (1981), implies the existence of discrete bodies of magma within the LVZ. In contrast, features recorded by the Cr-rich suites seem to require low melt/solid ratios in which zone-refining and/or assimilation processes may be operative. These conditions are more likely to be attained within rising diapirs than in discrete bodies of magma. In addition, sulphur and/or oxygen fugacities may play an important role.

Magma mixing is proposed to explain the various zonation patterns within the megacryst populations in the Fayette County kimberlite. Wyllie (1980) has presented a model for the origin of kimberlite into which the provenance of megacrysts and the magma mixing model can be incorporated, thus providing a coherent general scenario for the development of kimberlite within the LVZ.

Before optimum conditions for crack propagation occur (*i.e.*, extension in the lithosphere and sufficient volatile build-up), it might be expected that a series of minor partially melting diapirs would be rising adiabatically through the LVZ, with segregation of melts into ponded magma bodies at 120–80 km depth where these diapirs reach a temperature maximum (thermal barrier on the solidus). These melts would be fractionating in thermal gradients under essentially isobaric conditions (*e.g.*, Harte and Gurney, 1981). Thus, within the upper levels of the LVZ, there would be a concentration of discrete magma bodies representing the various stages of high *P* and *T* evolution of kimberlitic melts. These low-Cr melts would achieve ilmenite saturation with subsequent buffering of *mg* by bulk crystal/melt FeO MgO K_D 's.

Once a conduit has been established, and the solidus barrier breached, uprise of partially melted diapirs will occur from greater depths. The absence of a thermal barrier presents little opportunity for segregation and ponding of melts to occur. These diapirs produce optimum conditions for buffering of partial melts with respect to *mg* (*i.e.*, low melt/crystal ratios). These buffered partial melts are essentially identical to the parental melts of the Cr-poor suite megacrysts; occasionally, they may be Cr-rich (as in the case of the Colorado/Wyoming and Fayette County kimberlites). Whether a Cr-rich or an "average Cr" melt is present may depend on heterogeneities within the diapir source region. As a consequence of the buffered *mg* conditions, phases crystallizing from these partial melts will have *mg* values typical of garnet lherzolite assemblages (*i.e.*, Fo 93–90, *etc.*).

These melt-bearing diapirs will rise through the LVZ disrupting all the megacryst-bearing, Cr-poor crystal mushes in their paths, incorporating megacrysts and host

melts. The kimberlitic melt that rises through the cracks to the surface will be a hybrid melt, slightly more Fe-rich than the initial partial melt, but more Mg- and Cr-rich than the melt in equilibrium with ilmenite. Furthermore, it will contain megacrysts flushed out from the entire column within the LVZ (Cr-poor suite) mixed with those crystallizing from the rising diapir (Cr-rich or primitive Cr-poor suite). The more evolved megacryst populations will back-react with the more primitive hybrid melt, resulting in reverse zonation. The more primitive megacrysts will be normally zoned to the composition of the hybrid melt. The extent of back-reaction will depend upon the diffusion coefficients within the phases concerned and the residence time of the hybrid melt within the mantle. Zoned rims may be destroyed by abrasion during transport to the surface, or by reactions with the host melt resulting in formation of new phases. The megacrysts will show various degrees of shearing which are the effect of stresses within the diapir or have resulted from disruption of the cumulate mushes.

Individual kimberlite occurrences will be variations on a general theme of kimberlite evolution as discussed above. Differences between kimberlites will reflect different degrees of mixing of both melt and megacrysts. The different mineralogical trends characteristic of individual kimberlites (*e.g.*, variations in $Fe^3/(Fe^{3+}+Fe^{2+})$ trends within ilmenites) will be, in part, a function of differences in fO_2 between various end-members involved in the mixing. Stability of later-stage phases (*i.e.*, groundmass ilmenites and spinels) will depend upon parameters influencing the hybrid melt during its final uprise through the upper mantle and crust. In this scenario, *the kimberlite that reaches the upper crust will be hybrid in nature*; therefore, the megacryst inclusions are not phenocryst *sensu stricto*. However, since the end-member melts in the mixing process may be very similar, or even derivative, xenocryst (*sensu stricto*) may also be a misleading term. Within the context of a magma mixing model, megacrysts would be autocrysts or cognate xenocrysts.

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